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Basin hydrologic response relations to distributed physiographic descriptors and climate

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Abstract

The long-term (climatic) hydrologic response of a basin may be quantified using the evaporation efficiency (E/E_p or actual evaporation ratio with potential evaporation) and runoff ratio (R/P or ratio of runoff loss to precipitation). A key question is the degree to which the basin's physiographic features and regional climate can explain or predict these hydrologic response measures.

In this paper we present the results from 10 basins in diverse climates and terrains. The long-term hydrologic response is estimated using an equilibrium surface water–groundwater interaction model. We investigate variability between basins with an examination of the relationships between various physical characteristics and the hydrologic properties of basins. Neither climate nor physiography alone can explain observed interbasin variability. Six variables are selected to represent the basins' climate, geomorphology, and lithology, each of which has a conceptual relationship to basin-scale equilibrium hydrology. The parameters include median slope, relief ratio, drainage density, wetness ratio, infiltration capacity, and a saturated zone efficiency index. Two hydrologic variables (runoff ratio and evaporation efficiency) are selected from the output of a distributed hydrologic equilibrium model. We perform a stepwise regression to identify which combinations of variables are valuable in predicting the basin-average hydrologic fluxes. A combination of two variables estimate the runoff ratio with an R^2 or explained-variance fraction of 0.76; use of all six variables increases the prediction to an R^2 of 0.90. The stepwise regression technique fails to achieve a statistically significant model for evaporation efficiency, but a regression model using all six variables nonetheless achieves an R^2 of 0.79.

This paper demonstrates that physiographic and climate descriptors can explain a large fraction of basin-to-basin differences in modeled hydrologic response. The case has been built on modeled surface water–groundwater interaction and should next be extended to hydrologic response descriptors derived from observations alone.

Keywords: Physiographic descriptors; Surface water–groundwater interaction; Climate descriptor

1. Introduction

Physical features of a landscape influence the hydrologic behavior of a basin in many complex ways. Steep slopes, for example, may cause a sharp differentiation in the hydrology of upslope and downslope regions. This phenomenon may, however, be offset by fine-texture soils or a very humid environment. This study is designed to examine how multiple basin characteristics combine to affect the long-term average water balance.

The key science question is how the long-term

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hydrologic response of the basin is related to metrics that can be derived purely from auxiliary topographic and climate data. Digital elevation data at high resolution (e.g. 25–30 m) and climate variables such as long-term precipitation (P) and potential evaporation (E_p) are readily available for many territories. However, the actual evaporation (E) and distributed runoff (R) that depend on the soil moisture distribution (itself related to the surface water–groundwater interactions across the basin) are not always at hand. The question is what fraction of the basin-to-basin differences in hydrologic partitioning (E/E_p and R/P, both in the strict range of [0,1]) can be deduced from physiographic and climatic indicators.

In this paper we apply the equilibrium distributed hydrologic model of Berger (2000) to 10 basins in diverse topographic, soil, and climate conditions. The model yields spatial fields of actual evaporation, runoff, and recharge as a result of surface water–groundwater interaction. The basin-aggregate hydrologic response (E/E_p and R/P) for the 10 basins is related to several metrics that are strictly based on terrain, soil texture, and climate characteristics. It is shown that a large fraction of the total basin-to-basin variability in long-term E/E_p and R/P (approximately 80 and 90%, respectively) can be predicted based on these auxiliary data.

This is a preliminary study in that: (1) the actual evaporation (E) and runoff (R) are modeled rather than observed, (2) a limited number of physiographic and climatic indicators are used, and (3) only linear relationships are captured between the predictant and predictors. The hydrologic fluxes used for comparison are modeled rather than observed because of the scarcity of long-term, reliable observations of evaporation and runoff in the modeled locations. The use of modeled predictants leads to an unduly large dependence of the predictants on model characteristics, since only the input parameters have the ability to influence the output. Use of observed predictants may reduce the correlation between physical conditions and hydrologic fluxes. Following this feasibility study, application to observed predictants is necessary. High-quality actual evaporation and total runoff data for a number of basins need to be assembled. This is a major task in itself and introduces measurement and water imbalance errors that further cloud any existing partial predictability. This study may be used as a basis to screen the set of predictors and explore whether any fruitful relationships may be found.

2. Background

Two recent studies used statistical techniques to identify relationships between geomorphology and hydrology. Zecharias and Brutsaert (1988) identified eight morphologic variables with a known or theorized effect on groundwater outflow. The parameters included purely geomorphologic descriptors, such as basin area and relief, and hydromorphic characteristics, such as the length of perennial streams. Principal component analysis was performed on parameter values for 19 basins along the Appalachian plateau. The first three components, representing size, slope, and dissection, explained over 98% of the variance. Sefton and Howarth (1998) used the same technique to examine the relationships between modeled dynamic hydrologic response and physical basin descriptors. Hydrologic characteristics included loss and routing parameters; physical descriptors included morphology, soil type, land cover, and climatic indices. Sefton and Howarth (1998) estimated the principal components from the physical descriptors and then regressed the most significant components against the hydrologic variables. The first four components explained 63% of the variance, but no significant relationships were established between the components and the six hydrologic variables.

Our statistical analysis differs from the investigations cited above. Zecharias and Brutsaert (1988) only considered morphologic indices related to groundwater outflow and did not use any explicit hydrologic variables in the analysis. Like Sefton and Howarth (1998), we use modeled hydrologic variables. However, their hydrology included only dynamic hydrologic response characteristics. We look at long-term hydrologic variables that include both subsurface and surface processes.

This study is also directed towards developing regionalized water balance partitioning parameters with the belief that surface water–groundwater interactions are dominant factors in a basin's hydrologic response. The goal is to develop a set of predictors based on simple topographic and climatic parameters.

Table 1
Values of basin characteristics for 10 study basins (see text for variable definition)

Basin	S_{50} (–)	R_r (–)	D_d (km^{-1})	P/E_p (–)	i_r/K_s (–)	α_s (–)	E/E_p (–)	R/P (–)
Bear	3.3×10^{-1}	1.7×10^{-1}	1.8	1.2	1.3×10^{-2}	5.1×10^{-2}	0.36	0.21
Big Creek	4.3×10^{-1}	6.4×10^{-2}	1.2	0.3	4.5×10^{-3}	2.0×10^{-2}	0.11	0.72
Brushy	1.0×10^{-1}	5.8×10^{-3}	2.3	1.1	8.9×10^{-2}	2.5×10^{-5}	0.91	0.11
Midland	3.3×10^{-2}	9.0×10^{-3}	2.0	1.0	1.5×10^{-1}	1.4×10^{-4}	0.90	0.07
Moshannon	1.2×10^{-1}	1.6×10^{-2}	1.0	1.7	1.5×10^{-1}	8.7×10^{-5}	0.98	0.36
Ogden	7.1×10^{-2}	1.5×10^{-2}	1.6	0.4	2.4×10^{-1}	4.1×10^{-4}	0.41	0.00
Sacramento	1.0×10^{-2}	3.7×10^{-3}	3.4	0.2	1.4×10^{-1}	9.4×10^{-5}	0.23	0.00
Schoharie	1.9×10^{-1}	5.7×10^{-2}	1.7	1.4	1.1×10^{-2}	8.4×10^{-3}	0.30	0.53
Tombstone	7.0×10^{-2}	3.6×10^{-2}	2.4	0.2	1.3×10^{-2}	3.3×10^{-2}	0.25	0.14
Yreka	3.1×10^{-1}	8.6×10^{-2}	1.0	0.6	9.3×10^{-3}	2.1×10^{-2}	0.22	0.26

The idea of regionalization in hydrology has a rich heritage and ultimately important because the observing networks does not adequately sample the hetereogeneity of real watersheds. Regionalization has been applied to extremes (flood quantiles and distributional parameters as well as low flow conditions). Regionalization has also been used to develop large-scale water balance estimates. Much of the history of regionalization in hydrology and its current status is summarized in the recent publication by Diekkruger et al. (1999).

2.1. Data and model

The parameter values used in this analysis come from multiple sources. The topographic parameters (median surface slope (S_{50}), relief ratio (R_r), and drainage density (D_d); see Table 1 and Section 3.1) come from 30 m digital elevation maps available from the US Geological Survey. Precipitation characteristics (both annual precipitation P and mean storm intensity i_r) come from data provided by the National Climate Data Center and processed by Hawk and Eagleson (1992). The soil hydraulic conductivity and soil depth are taken from the STATSGO on-line soil database (National Soil Survey Center, 1991) and converted to Brooks–Corey parameters according to the relationships for soil type given by Bras (1990). The STATSGO soil characteristics were made by generalizing detailed survey data and were designed for use in broad regional planning. They provide a broad sense of the lithology but sacrifice representation of small-scale variability. Fine-scale hydrologic fluxes may be inaccurate as a result, but the impact is less significant on basin-averaged hydrology.

The hydrologic fluxes, evaporation and runoff, are generated by a distributed equilibrium model. A conceptual figure of the relationship between groundwater and surface water fluxes is provided in Fig. 1. The equilibrium hydrology model is designed around capturing surface water–groundwater interactions. Thirty-meter resolution digital topography is used to define the basin and its physiography. A finite-difference groundwater code is used to model the spatial response of the water table to patterns of long-term recharge and discharge. The unsaturated zone is modeled using the statistical–dynamical approach of Eagleson (1978), pp. 705–748 modified to capture surface water–groundwater interaction. The vertical one-dimensional unsaturated zone processes are in statistical equilibrium and include the intermittent effects of storms and interstorms. The recharge and discharge at the bottom of the unsaturated zone couples groundwater processes and lateral redistribution. The climatic input parameters required are storm statistics such as mean storm depth, mean interarrival time, and mean duration as well as mean potential evaporation. Soil hydraulic properties as well as a representative soil mantle thickness also needs to be specified. No calibration is performed on the model since all the terrain and rainstorm parameters are observed. Appendix A provides the main equations that capture these described conditions.

The equilibrium model is built on a groundwater model for the basin that is forced at its upper boundary (net recharge) with a version of the Eagleson (1978) analytical unsaturated zone dynamics model that has

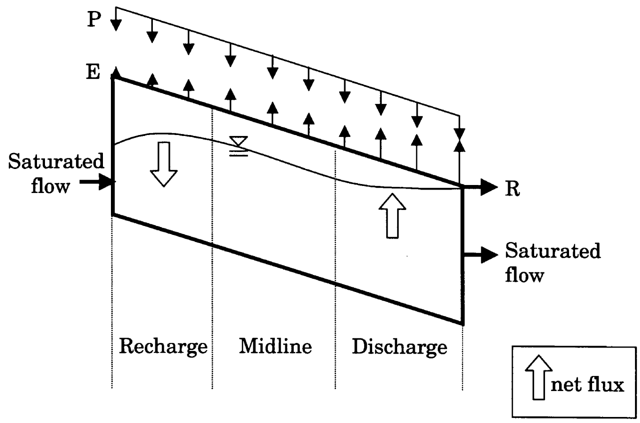

Fig. 1. Conceptual schematic of surface water-groundwater interaction along a hill-slope.

been adapted for bounded domain or shallow groundwater table conditions by Salvucci and Entekhabi (1995). The key to the surface water–groundwater coupling is the land-surface hydrologic partitioning as a function of water table depth. Fig. 2 shows an example for one of the 10 basins. The Eagleson (1978) model solves the unsaturated zone water balance that results from intermittent Poisson arrival of rectangular pulse storms. It includes switching boundary conditions and transitions from climate-limited to soil-controlled storm and interstorm fluxes in the soil column. The example in Fig. 2 is for a particular combination lead to high runoff and evaporation conditions. When coupled to a groundwater model, riparian zones develop in basins. Upslope regions with deep groundwater are net recharge zones. The riparian zones are net discharge zones. For each of the 10 basins, the spatial patterns of recharge and discharge are estimated using the coupled surface water–groundwater model. Between the net recharge and discharge zones there are often extended midline regions where the groundwater energy slope parallels

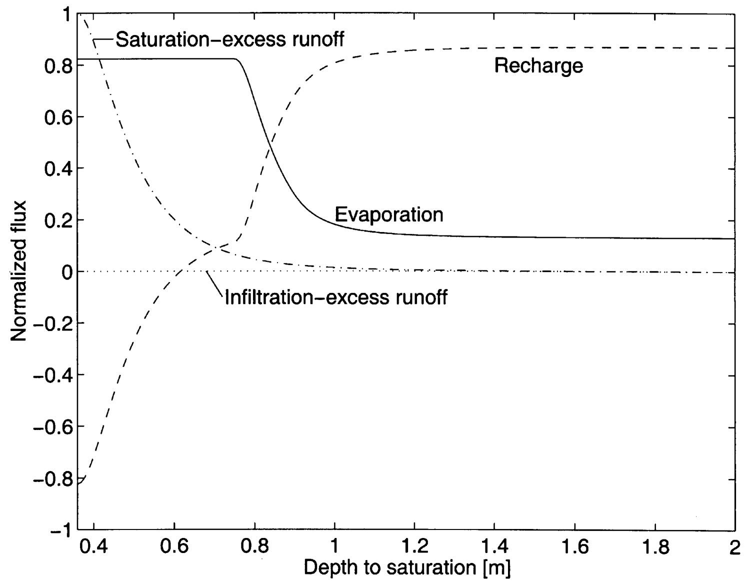

Fig. 2. Equilibrium hydrologic fluxes as a function of depth to saturation, Bear Valley, CA.

the surface topography. These regions transmit groundwater downslope but there is no divergence (which would equal the unsaturated zone recharge). Each basin, depending on its climate and soil texture has a critical groundwater table depth at which there is net zero recharge. The zero-recharge depth (called Z^*) is a useful combined soil and climate parameter for characterizing the spatial patterns of net recharge, discharge, or midline zones that may develop in a basin.

3. Basin descriptors

A major challenge in quantitative geomorphology is the development of indices to characterize physically important characteristics of a natural landscape. A number of criteria are employed in the selection of morphologic variables for the interbasin analysis: a known or theorized physical relationship to distributed equilibrium hydrology, a focus on equilibrium hydrology rather than on dynamic routing effects (i.e. we do not use the network-based Strahler or Horton ratios), nondimensionality (where possible), and independence from other variables. Three strictly topographic indices are selected: median slope, relief ratio, and drainage density. These three descriptors are added to variables that describe the climate, soil, and hydrologic responses of the basins. Below we describe the relevance of each of the nine parameters and its expected role. Table 1 contains the values of each parameter for all basins.

3.1. Definition of variables

Median surface slope (S_{50}). The median surface slope contains limited information on the distribution of slopes within a basin. We consider the median as opposed to the mean slope to avoid bias from a few very steep or shallow areas. Assuming that the bedrock is parallel to the ground surface, the surface slope represents the gradient driving lateral Darcy flow. In basins with a high slope, lateral flow may transport moisture out of a cell faster than it is replenished by recharge or incoming groundwater, causing the cell to dry out.

Relief ratio (R_r). The relief ratio is a measure of basinwide average slope. As opposed to the median surface slope, which is determined from the steepest slope at every pixel, R_r is the ratio of the total basin relief (the elevation difference between the highest and lowest points) and a representative basin length. That length is defined as the straight distance between the outlet and the farthest point in the basin from the outlet. This index provides an approximate estimate of the topographic gradient affecting the lateral movement of groundwater in near-surface saturated areas on the scale of the entire basin. High values of R_r should be correlated with efficient lateral redistribution of moisture.

Drainage density (D_d). Drainage density is defined as the ratio of total stream length to basin area. It is an approximate measure of the inverse mean horizontal hillslope length. It is often seen as a key indicator of the hydrologic response of a landscape, given the difference in velocity and residence time of water between the hillslope and stream channel. The drainage density also has implications for the extent of saturated areas and runoff generation. The soil is more likely to be saturated within the channel network than on the upper reaches of a hillslope; there is therefore a positive relationship between drainage density and the spatial extent of the riparian zone. A low value of D_d corresponds to a landscape with long hillslopes; a high D_d indicates a dissected landscape.

The estimation of a basin's drainage density requires delineation of the channel network. The identification of where a channel begins is an area of ongoing research. Channel heads may be estimated by field observation, visual estimation from topographic blue lines or aerial photographs, or from digital elevation models using automated techniques. We use the constant-drop approach described by Tarboton et al. (1991) for automated estimation of a channel network from digital elevation data. The approach holds that the elevation drop in each stream link should be independent of the Strahler order of that link. The minimum threshold contributing area is selected such that the resulting network has constant stream drops for different order links at the 95% significance level. In order to save space we only considered 95% level although other levels may as well as be studied. Given the limited sample size, an in-depth comparison of different significance levels is not very productive and meaningful.

Wetness ratio (P/E_p). The ratio of annual precipitation to annual potential evaporation has dual significance for hydrologic processes. Considered as the ratio of mean storm depth to evaporative depth ($i_r t_r / e_p t_b$, to which it is equal), it is a measure of the ratio of moisture input to output in the unsaturated zone. Here i_r is the mean storm depth, t_r the mean storm duration, e_p the potential evaporation rate, and t_b is the mean time between storms. These are parameters of a stochastic rectangular-pulse rainstorm model with Poisson storm arrival rates. Considered as the annual wetness index P/E_p, it is an indicator of the atmospheric supply and demand of moisture. Potential evaporation is used instead of actual evaporation because the index is designed to represent climatic forcing only; reduction in evaporation rates below their potential is caused by limited soil moisture or vegetation effects. High values signify a moist environment and a small midline area due to the abundance of available water in the unsaturated zone.

Relative infiltration capacity (i_r/K_s). The ratio of mean precipitation intensity to saturated hydraulic conductivity (K_s) combines soil and climate characteristics to provide a rough indicator of the soil's ability to absorb the average rainfall. The hydraulic conductivity is the maximum rate at which water may be transmitted through the soil given a unit pressure gradient. If the precipitation intensity is much less than K_s, the soil can transport water through the soil column without ponding or surface runoff. Conversely, a relatively high ratio indicates a greater likelihood of infiltration-excess runoff.

Saturated zone efficiency (α_s). Salvucci and Entekhabi (1995) introduced an index to capture the ability of groundwater to laterally redistribute moisture downslope. This variable is modified here and called α_s. The variable is a nondimensional combination of saturated hydraulic conductivity, water table depth in the midline zone (Z^*), relief ratio, mean annual precipitation, and basin area:

$$\alpha_s \equiv \frac{K_s(Z_T - Z^*)R_r}{P\sqrt{A}}$$

where Z_T is total soil depth. The product of the saturated hydraulic conductivity K_s and the water table gradient ($Z_T - Z^*$) represents the Darcy potential for lateral flow in the saturated layer of the soil. This is scaled by a hillslope length and considered in proportion to the incident precipitation. The moisture input estimated in the denominator should technically be reduced by the fraction of contributing area above the midline. However, not only is the fraction unknown, it is also inherently related to other key variables such as hydrology and climate. We therefore use the square root of basin area ($\sqrt{A}$) as the representative length scale and neglect any adjustment for the extent of the recharge area.

It was found that α_s is at least partially responsible for an inverse correlation with the extent of midline. (The second important influence on the midline zone is P/E_p, a measure of the unsaturated zone tendency to provide recharge.) It is expected that α_s will play an important role in basin behavior beyond just the formation of the midline, since it combines many physical features in a single variable that represents the overall efficiency of the saturated zone.

The 10 basins selected for this study come from a range of environments; see Fig. 3 for their locations. The climatic range extends from dry (Sacramento, CA and Tombstone, AZ, where the wetness ratio is 0.2) to moist (Moshannon, PA, where the same ratio is 1.7). The topography also varies from flat to hilly to steep. The extremities of relief are found in the Big Creek and Sacramento basins where the median slope is 43 and 1%, respectively. Basins were selected to maximize the diversity of their characteristics, so that not all steep catchments were also humid and fine-soiled. The details of the individual basins' characteristics are found in Table 1.

4. Results

4.1. Pairwise correlations

The first step in identifying patterns in interbasin behavior is to examine the correlations between pairs of variables. Table 2 contains the correlation coefficients for the 10-basin, nine-variable dataset. Five pairs of basins are significant at the 95% level (correlation coefficient greater than 0.75). No coefficients are significant at 99% confidence.

Calculation of the correlation matrix of a multivariate dataset provides information on the relationships between pairs of variables, but it cannot identify significant multi-dimensional relationships. To understand the simultaneous interactions between climate,

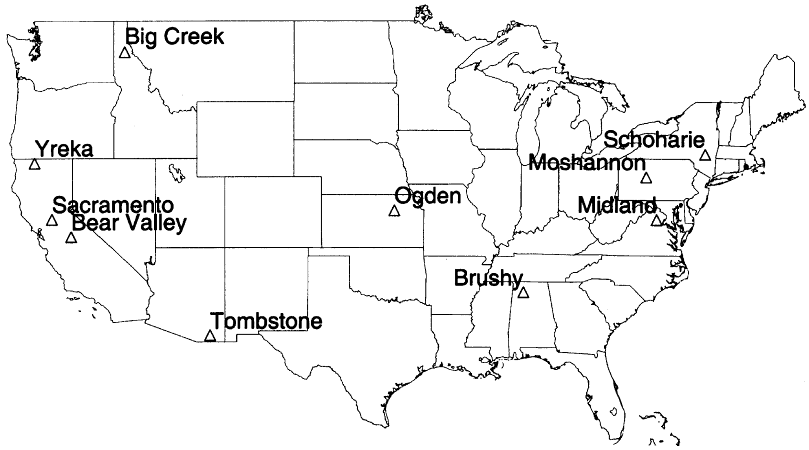

Fig. 3. Map of basin locations.

physiography, and hydrology, it is necessary to identify significant relationships between more than two variables at a time. We are interested in combinations of descriptors that vary in unison and the roles of the different clusters in determining the behavior of the study basins. These multi-variable combinations are investigated through the use of stepwise regression. The limitation is that only linear predictant–predictor relationships are explored. As a first-step feasibility study, this is adequate. The advantage of restricting the relations to linear form is that the statistical significance and confidence levels may be easily demonstrated. Methods and results are presented below.

5. Stepwise regression

In this analysis, we are interested in assessing whether combinations of topographic, soil, and climate variables can predict hydrologic response with reasonable accuracy. Stepwise regression is used to identify which variables contribute a significant amount toward the prediction of the runoff ratio (R/P) and evaporation efficiency (E/E_p). One limitation of this approach is the assumption of a linear relationship between predictant and predictors. Much of the dynamic response in the hydrologic cycle is nonlinear in nature. The consideration on linear relationships only is able to capture a significant fraction of the hydrologic variability between basins given that the focus is on equilibrium hydrologic response. It is appropriate as a preliminary analysis of the 10 basins considered here.

The stepwise regression is performed for each of the hydrologic fluxes — evaporation efficiency and runoff ratio — and the six non-hydrologic variables: median slope, relief ratio, drainage density, basin wetness, infiltration capacity, and saturated efficiency

Table 2
Correlation coefficients between variables used in principal component analysis, $N = 10$. Values significant at the 95% level are in boldface

Basin	S_{50}	R_r	D_d	P/E_p	i_r/K_s	α_s	E/E_p	R/P
S_{50}	1.00	–	–	–	–	–	–	–
R_r	**0.76**	1.00	–	–	–	–	–	–
D_d	−0.62	−0.33	1.00	–	–	–	–	–
P/E_p	0.07	0.17	−0.41	1.00	–	–	–	–
i_r/K_s	−0.68	−0.64	0.16	−0.01	1.00	–	–	–
α_s	0.60	**0.88**	−0.12	−0.11	−0.71	1.00	–	–
E/E_p	−0.43	−0.42	−0.11	0.65	0.51	−0.52	1.00	–
R/P	0.75	0.32	−0.59	0.26	−0.60	0.20	−0.26	1.00

Table 3
R^2 values between runoff ratio and individual basin characteristics

Variable	R^2
S_{50}	0.01
α_s	0.01
R_r	0.05
i_r/K_s	0.06
D_d	0.23
P/E_p	0.70

index. The F-tests are performed for a significance level of 95%.

In order to determine the relevant variables in estimation of the runoff ratio, we begin by looking, for each individual variable, at the R^2 values, or the percent explained variance between basins in diverse terrain and climate conditions. These values are provided in Table 3. The R^2 for models with increasing number of variables, together with the coefficients on each variable, are summarized in Table 4. Only the first two variables, P/E_p and i_r/K_s, pass the F-test as contributing a significant new amount of information to the regression model. However, inclusion of all of the variables improves the model prediction to explain 90% of the total variance. The results for both the two-variable and six-variable models are presented in Fig. 4. The model predictions are compared against the R/P values generated by the model.

The climatic wetness is the most significant variable; it represents the moisture available at the surface for runoff or evaporation. The infiltration capacity is related to runoff primarily because of the hydraulic conductivity in the denominator of the index: the soil conductivity governs the rate at which the saturated zone transmits moisture downslope to the saturated areas where it emerges as runoff. The remaining variables primarily capture physical features of the basin such as slope and channel density. The topographic characteristics improve the fit of the regression model, but remain of secondary importance relative to the large role of the climatic wetness.

A similar procedure is followed to determine the regression model for E/E_p. The R^2 values for single variables are summarized in Table 5. Of the individual variables, the basin wetness is again of greatest importance. P/E_p is correlated to E/E_p with an R^2 of 0.42. The F value for the significance test of the regression between E/E_p and P/E_p is 5.81, however, which is less than the 5.99 required for statistical significance. Using the variables included in this study, there is no statistically significant regression model for predicting the evaporation efficiency at 95% significance.

If we relax the strict requirements for passing the F-test and examine the performance of the regression models for increasing number of variables, we find that linear combinations of the six variables can explain up to 79% of the variance in the evaporation ratio. The results for the one- to six-variable regression models for evaporation efficiency are summarized in Table 6. The one-variable and six-variable models are compared against the model-generated fluxes in Fig. 5. Although the evaporation efficiency cannot be predicted as reliably as the runoff ratio, the six-variable model brings the predicted values to within 20% of the model values. This is a significant improvement over a single-variable model using the climatic wetness, which has errors on the scale of 50%.

The evaporation efficiency is more difficult to predict than the runoff ratio due to the pattern in

Table 4
R/Pstepwise regression results for increasing numbers of model variables for runoff ratio prediction. Model coefficients are provided in the order listed in the first column. The regression model constant (y-intercept) is provided in the final column. The horizontal line represents the cut-off below which additional variables do not add statistically significant information to the model

Variable(s)	R^2	Coefficients						Constant
P/E_p	**0.70**	0.16						0.05
P/E_p, i_r/K_s	**0.76**	0.16	−0.32					0.02
P/E_p, i_r/K_s, S_{50}	**0.77**	0.16	−0.46	−0.12				−0.02
P/E_p, i_r/K_s, S_{50}, D_d	**0.87**	0.13	−0.75	−0.46	−0.07			−0.27
P/E_p, i_r/K_s, S_{50}, D_d, α_s	**0.89**	0.13	−0.62	−0.52	−0.08	1.36		−0.26
P/E_p, i_r/K_s, S_{50}, D_d, α_s, R_r	**0.90**	0.15	−0.51	−0.39	−0.07	2.73	−0.59	−0.21

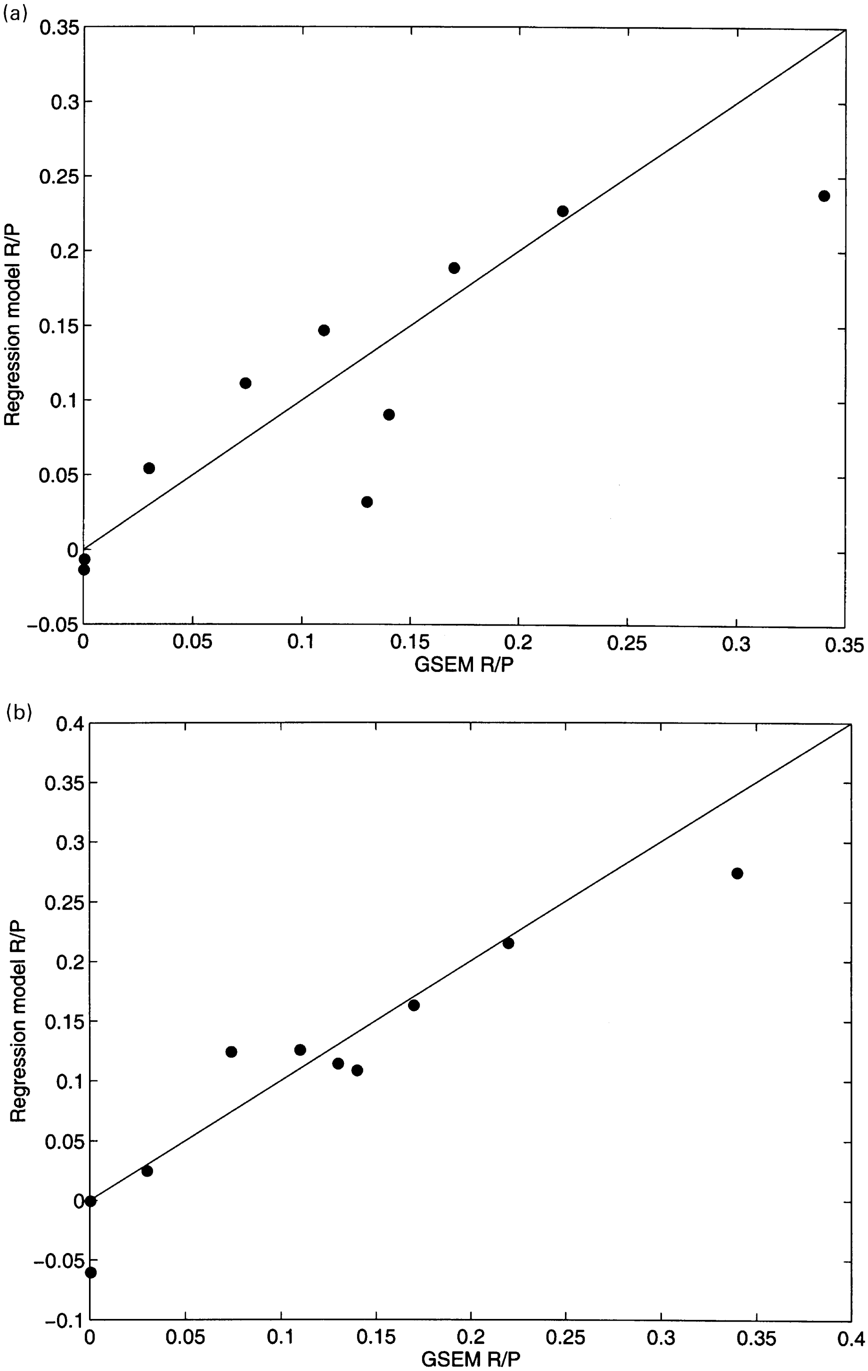

Fig. 4. Performance of (a) two-variable regression model and (b) six-variable regression model in *R*/*P* prediction. The 1:1 line is plotted for reference.

Table 5
R^2 values between evaporation efficiency and individual basin characteristics

Variable	R^2
D_d	0.01
R_r	0.19
S_{50}	0.21
α_s	0.25
i_r/K_s	0.25
P/E_p	0.42

which evaporation varies with water table position. A sample relationship between evaporation and Z_w is found in Fig. 2. In coarse-soiled basins, evaporation essentially behaves as a step function. When the water table is shallow, evaporation occurs at the climatically determined potential rate. As the water table deepens, evaporation abruptly drops to a low value. As a result, for a smooth distribution of water table depths, the distribution of evaporative flux tends to be bimodal with very few intermediate values. This pattern makes it difficult to predict the basin-average evaporation rate used in calculating the evaporation efficiency. Despite these limitations, incorporation of multiple basin characteristics allows development of a predictive linear model which can explain nearly 80% of the variability in model evaporation values.

6. Summary

In this paper we have characterized the dominant modes of interbasin variability influencing the equilibrium fluxes. Stepwise regression is performed in an effort to predict the two hydrologic partitioning parameters, evaporation efficiency and runoff ratio. The basin wetness and infiltration capacity together predict the runoff ratio with an R^2 or explained variance of 0.76. A model that uses all six variables improves the fit to an R^2 of 0.90. The same technique fails to achieve a statistically significant model for evaporation efficiency, but a regression model using all six variables is able to achieve an R^2 of 0.79.

In this feasibility study to predict basin-to-basin differences in hydrologic partitioning based on terrain and climate variables alone, it is shown that there is some promise to extract long-term hydrologic information from strictly auxiliary topographic, soil texture, and rainstorm statistics data. The limitation is that the hydrologic predictants are drawn from the application of a physically based model of surface water–groundwater interaction. The model is designed to produce a long-term equilibrium scenario; consideration of extreme events would require perturbing the equilibrium conditions and observing the transient response. The next step is to assemble observed actual evaporation and runoff loss data for a number of basins. Due to difficulties in measuring actual evaporation and obtaining river outflow measurements with adequate (unbiased) sampling and accuracy to close the basin water balance, this task requires a major undertaking. The conclusion here is that terrain, soil, and rainfall statistics measures, when combined to descriptive non-dimensional measures, may capture the basin-to-basin variability in hydrologic partitioning. When a larger sample of basins (greater than the 10 used here) is used, the further assumption of linear correspondence between predictants and predictors may also be relaxed. Nonetheless, even with linearity assumptions, the terrain, soil texture, and rainform statistics descriptors based on auxiliary data do succeed in explaining up to 80% of the evaporation efficiency

Table 6
R/P stepwise regression results for increasing numbers of model variables for evaporation efficiency prediction. Model coefficients are provided in the order listed in the first column. The regression model constant (y-intercept) is provided in the final column

Variable(s)	R^2	Coefficients						Constant
P/E_p	**0.42**	0.16						0.17
P/E_p, R_r	**0.70**	0.44	−3.37					−0.05
P/E_p, R_r, i_r/K_s	**0.74**	0.43	−2.25	1.06				0.06
P/E_p, R_r, i_r/K_s, α_s	**0.79**	0.50	−4.98	1.52	10.57			0.16
P/E_p, R_r, i_r/K_s, α_s, S_{50}	**0.79**	0.52	−5.95	1.78	12.57	0.30		0.21
P/E_p, R_r, i_r/K_s, α_s, S_{50}	**0.79**	0.53	−6.44	1.98	13.51	0.57	0.04	0.36

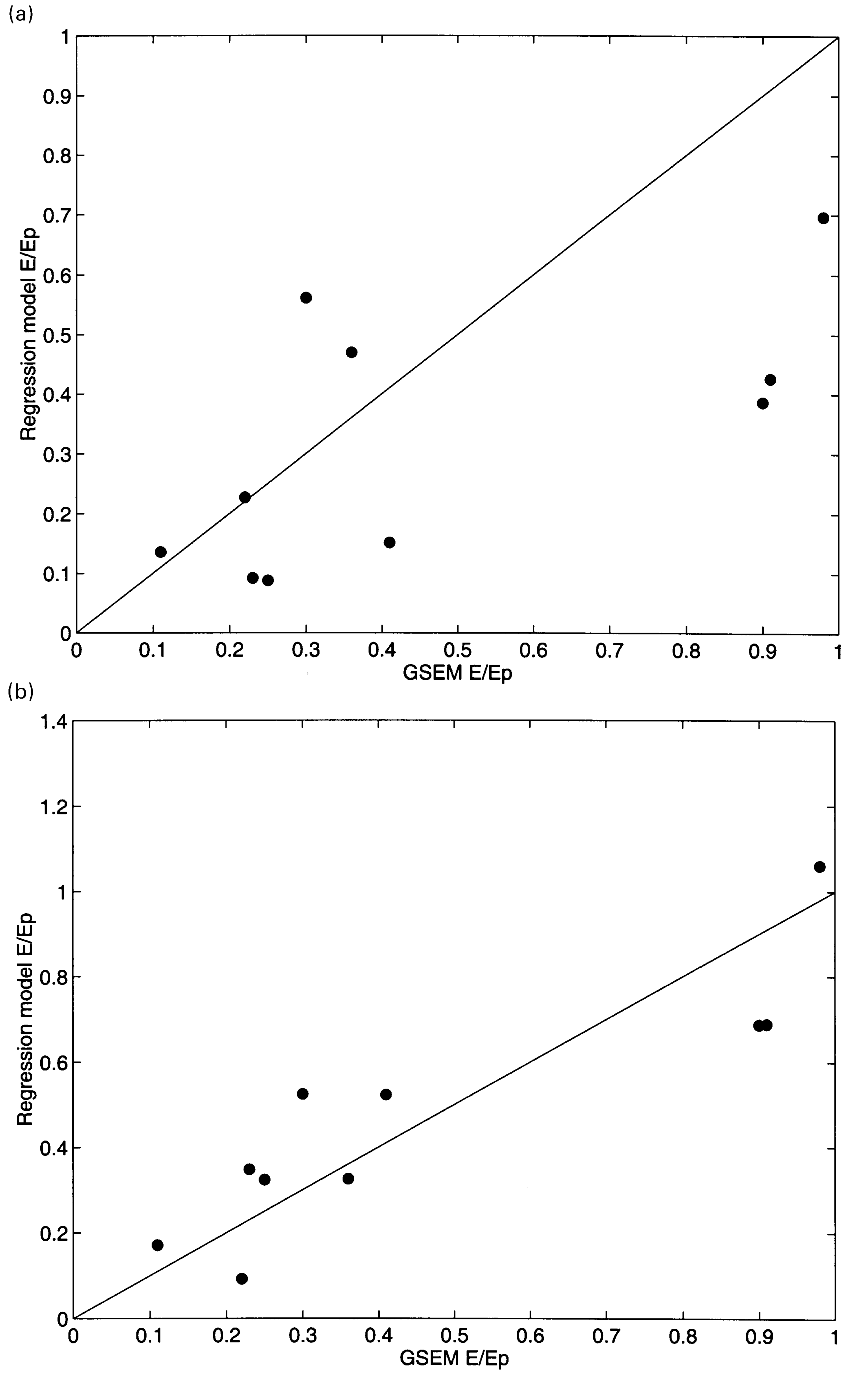

Fig. 5. Performance of (a) two-variable regression model and (b) six-variable regression model in E/E_p prediction. The 1:1 line is plotted for reference.

and 90% of the runoff ratio variance between basins in diverse physiographic and climatic conditions.

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Appendix A. Model equations

The following paragraphs describe the relationships for unsaturated zone fluxes used in the model. For a full derivation of fluxes for an infinite water table, see Eagleson (1978); modifications for a shallow water table are described in detail in Salvucci and Entekhabi (1995).

A.1. Infiltration-excess runoff

Infiltration-excess runoff occurs when precipitation intensity exceeds the capacity of the soil to transport water vertically downward. Eagleson (1978), pp. 741–748 uses the Philip solution (1957) to characterize infiltration. The mean annual flux is the integral of all possible storm intensities from the time when ponding first occurs (t_p) to the time when the soil saturates (t_s)

$$R_{ie} = \int_{A_0}^{\infty} \alpha\,\mathrm{e}^{-\alpha}\,\mathrm{d}i \int_{t_p}^{t_s} \delta\,\mathrm{e}^{-\delta\tau}(i - A_0)(\tau - t_p)\,\mathrm{d}\tau \quad \text{(A1)}$$

The integral can be solved to give the following expression for mean annual value of infiltration-excess runoff:

$$R_{ie} = P\left\{\exp(-\alpha A_0 - \mu)\Gamma\left(1 + \frac{1}{2}\mu\right)\left(\frac{1}{2}\mu\right)^{-1/2\mu}\right\} \quad \text{(A2)}$$

where P is mean annual precipitation, $\Gamma(\cdot)$ the Gamma function, and μ is defined as

$$\mu = [\alpha^2 \delta S_i^2]^{1/3} \quad \text{(A3)}$$

where S_i is sorptivity. Salvucci and Entekhabi (1995) modified this expression in two ways

$$R_{ie} = R_{ie}^{\text{Eagleson}}(1 - \mathrm{e}^{-\delta t_s}(1 + \delta t_s)) \quad \text{(A4)}$$

A.2. Saturation-excess runoff

Saturation-excess runoff due to wetting from above can be calculated by integrating over all possible storm intensities and durations, given time to soil column saturation t_s, using the following equation:

$$R_{se} = \int_0^{\infty} \alpha\,\mathrm{e}^{-\alpha i_r}\,\mathrm{d}i_r \int_{t_s}^{\infty} i_r(t - t_s)\delta\,\mathrm{e}^{-\delta t}\,\mathrm{d}t \quad \text{(A5)}$$

We assume that infiltration is not soil-controlled, since storms large enough to cause soil saturation will tend to have large total storm depths. As a result, time to saturation is a simple function of available storage and storm intensity ($t_s = \forall_e / i_r$), giving the solution for mean annual runoff

$$R_{se} = 2m_v \forall_e K_2[2\sqrt{\forall_e \alpha\delta}] \quad \text{(A6)}$$

where $\forall_e$ is the antecedent available storage in the unsaturated zone and $K_2[\cdot]$ is the modified Bessel function of the second order.

A.3. Bare-soil evaporation

When the soil is not covered by vegetation, evaporation is limited either by the ability of the soil to transport water upward to the ground surface or by the capacity of the atmosphere to evaporate available moisture. The average flux is therefore determined by integration of the smaller of potential evaporation (atmospheric control) and exfiltration capacity (soil control) over all possible interstorm durations

$$E_{bs} = m_v \int_0^{\infty} \left[\int_0^{\tau} \min(f_e^*, e_p)\,\mathrm{d}t\right] f_{t_b}(\tau)\,\mathrm{d}\tau \quad \text{(A7)}$$

where f_e^* is the exfiltration capacity and $f_{t_b}(\tau)$ is the probability distribution of time between storms. The solution of the integral results in Eagleson's expression (1978), pp. 741–748 for mean annual evaporation

$$E_{bs} = \bar{e}_p\left\{1 - (1 + \sqrt{2E}\,\mathrm{e}^{-E} + \sqrt{2E}\left(\Gamma\left[\frac{3}{2}\right] - \gamma\left[\frac{3}{2}, E\right]\right)\right\} \quad \text{(A8)}$$

where $\bar{e}_p$ is the mean annual potential evaporation rate and $\gamma(\cdot,\cdot)$ is the incomplete gamma function. E is evaporation effectiveness, which considers the interstorm climate and desorptivity characteristics of the soil.

With a near-surface water table, the unsaturated hydraulic conductivity at soil saturation must be considered; it was neglected by Eagleson (1978), pp. 731–739 because an infinite-depth soil column never saturates. This results in the following modified equation for mean annual bare-soil evaporation:

$$E_{\rm bs} = \frac{\bar{e}_{\rm p} m_{\rm v}}{\eta}\Big\{1 - (1 + \sqrt{2\Lambda}E + 2(\Omega)^{-1/2})\,{\rm e}^{-\Lambda E} + ((2\Omega)^{-1/2} + \sqrt{2\Omega}E)\,{\rm e}^{-\Omega E} + \sqrt{2E}\cdot\left(\gamma\left[\frac{3}{2},\Omega E\right] - \gamma\left[\frac{3}{2},\Lambda E\right]\right)\Big\} \tag{A9}$$

Λ and Ω are dimensionless parameter groups defined in the Appendix.

A.4. Transpiration by vegetation

The presence of vegetation affects the rate of moisture flux from the surface because roots provide access to deeper regions of the soil profile that are moister than the near-surface soil. The equation for bare-soil evaporation, Eq. (A9), is modified for the presence of vegetation represented by a full vegetation cover and a uniform rooting depth to give the following equation for transpiration from a fully-vegetated surface (Levine and Salvucci, 1999):

$$E_{\rm vs} = \begin{cases} \frac{\bar{e}_{\rm p} m_{\rm v}}{\eta}\{1 - (1 + \sqrt{2\Lambda}E + (2\Omega)^{-1/2})\,{\rm e}^{-\Lambda E} + ((2\Omega)^{-1/2} + \sqrt{2\Omega}E)\,{\rm e}^{-\Omega E} + \sqrt{2E}\left(\gamma\left[\frac{3}{2},\Omega E\right] - \gamma\left[\frac{3}{2},\Lambda E\right]\right) & w < \kappa \\ \frac{\bar{e}_{\rm p} m_{\rm v}}{\eta}\{1 - \left(1 + \sqrt{2\Lambda}E + \frac{\kappa}{e_{\rm p}} - \frac{w}{e_{\rm p}}\right){\rm e}^{-\Lambda E} + \sqrt{2E}\left(\Gamma\left[\frac{3}{2}\right] - \gamma\left[\frac{3}{2},\Lambda E\right]\right) & \kappa < w < \kappa + e_{\rm p} \\ \frac{\bar{e}_{\rm p} m_{\rm v}}{\eta} & w > \kappa + e_{\rm p} \end{cases} \tag{A10}$$

where κ is defined as

$$\kappa \equiv \frac{K_{\rm s} s_*^{\rm c}}{2} \tag{A11}$$

A.5. Saturated flow

The saturated zone is modeled using MODFLOW, the three-dimensional finite difference flow model developed by the US Geological Survey (McDonald and Harbaugh, 1988). The equation for flow in the saturated zone is

$$\frac{\partial}{\partial x}\left(K_x\frac{\partial h}{\partial x}\right) + \frac{\partial}{\partial y}\left(K_y\frac{\partial h}{\partial y}\right) + \frac{\partial}{\partial z}\left(K_z\frac{\partial h}{\partial z}\right) = S_y\frac{\partial h}{\partial t} \tag{A12}$$

where K_i is the hydraulic conductivity in the i direction, h the hydraulic head, and S_y is the specific storage of the soil. When conductivity is isotropic ($K_x = K_y = K_z = K_s$), the equation simplifies to

$$K_s\left[\frac{\partial^2 h}{\partial x^2} + \frac{\partial^2 h}{\partial y^2} + \frac{\partial^2 h}{\partial z^2}\right] = S_y\frac{\partial h}{\partial t} \tag{A13}$$

The coupling of the Eagleson–Salvucci unsaturated fluxes to the groundwater model was originally done by Ateljevich (1995).

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