JP1.8 Hydroclimatological Predictions Based on Basin's Humidity Index

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1. Introduction

The use of long-term records of observational data to determine the variability of land surface moisture fluxes and storages is essential for answering key science questions concerning the degree to which the global hydrologic cycle is intensifying in response to potential anthropogenic climate forcings and to assess the potential for improving predictability of floods and droughts. In addition to quantifying climate anomalies, an accurate characterization of land surface states will improve runoff predictability. Energy fluxes and precipitation control runoff and evapotranspiration rates. In arid regions where available energy supply is greater than the latent heat required to evaporate total precipitation, evapotranspiration equals or approaches total precipitation. In more humid regions the available energy supply can be less than the latent heat required to evaporate all precipitation, resulting in the actual evapotranspiration approaching the potential evapotranspiration.

Here, we present a new empirical model that predicts the sensitivity of available energy and precipitation in determining values of. This model is a simple one-parameter formulation that account for the expected dominant vegetation type, while maintaining Budyko's (Budyko 1950) assumptions related to the concept of "geographical zonality": that (a) the biome type is determined by the aridity (or humidity) index and (b) the evaporation efficiency is a function only of the aridity (or humidity) index. Computed evapotranspiration values for 314 watersheds within the Arkansas/Red River system agreed reasonably well with the empirical model predictions. The model was extended to develop two other formulae that were used successfully to predict the sensitivity of runoff and evapotranspiration to precipitation variations (Shairf et al. 2005). The empirical model can serve as a useful tool to estimate the changes in various components of the hydrologic cycle as a result of climate change.

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2. ET Model

Remotely Energy fluxes and precipitation control the runoff and ET rates. In arid and semi-arid regions where available energy supply is greater than the latent heat required to evaporate total precipitation, equals or approaches total precipitation. In more humid regions the available energy supply can be less than the latent heat required to evaporate all the precipitation and hence actual ET approaches potential ET. Validation of TOPLATS simulated energy fluxes were limited by the relative scarcity of energy flux observations within the domain. However, Crow and Wood (2002) spatially averaged TOPLATS flux predictions during the 1994 growing season over the ARM-CART SGP site and validated them against spatially interpolated flux measurements made with nine Bowen ratio towers located in Oklahoma and Kansas. Errors in TOPLATS energy flux predictions were comparable to typical levels of systematic closure errors in the surface energy budget measured by eddy covariance (Twine et al., 2000); see Crow and Wood (2002) for further discussion.

Another way to evaluate model simulations is to examine how they reflect the control of precipitation and available energy, manifested by potential ET (PET), on the mean annual ET at the sub-basin scale. To obtain unbiased results, observed PET is used rather than estimates by the land surface model. Several empirical formulae that compute mean annual ET for a basin as a function of the ratio of mean annual precipitation and PET have been derived based on the assumption that annual change in soil moisture storage is negligible. Among the most frequently cited are those suggested by Schreiber (1904), Ol'dekop (1911), and Budyko (1948).

$\frac{E}{E_p} = \phi(1 - e^{-1/\phi})$	Schreiber (1904)
$\frac{E}{E_p} = \tanh(\phi)$	Ol'dekop (1911)
$\frac{E}{E_p} = \left[\phi \tanh(\phi)\left(1 - e^{-1/\phi}\right)\right]^{1/2}$	Budyko (1948)

Table 1

Definition of three empirical formulae for the relationships between the ratio of actual to potential evapotranspiration and ratio of precipitation to potential evapotranspiration

These three formulae are listed in Table 1. Budyko's formula, the most widely used, is simply the geometric mean of the formulae suggested by Schreiber (1904) and Ol'dekop (1911). All these formulae express ET as a function of the humidity index, which is the ratio between precipitation and PET (Equation 2 below).

Zhang et al. (2001) suggest a formula with an additional parameter to take into account the dominant vegetation type:

$$\frac{E}{E_P} = \frac{\phi + w}{1 + \phi + w/\phi} \tag{1}$$

$$\phi = \frac{P}{E_p} \tag{2}$$

Where P is the mean annual precipitation, E the mean annual ET, Ep annual PET, w is the vegetation type available water coefficient, which ranges between 0.5 and 2.0 for grassy and forest regions, respectively, and ϕ is denoted as the basin's humidity index. Whereas Zhang et al. (2001) cited several studies indicating the effect of the vegetation type on ET, Budyko (1950) and others classified regions into desert, steppe, forest, or tundra according to the value of ϕ. None of the formulae in Table 1 explicitly takes into account the effect of vegetation type. However, the formula in Equation 1 does not work properly when ϕ approaches 2.0 (e.g. for very wet forested regions the value of ET/Ep can exceed 1.0). Since vegetation type is a function of ϕ and w is a function of the vegetation type, Equation 1 can be simplified by assuming that w varies linearly with vegetation type i.e. between the end of the desert region where ϕ is equal to 0.5 and into the forest region for ϕ approaching 2.0. Based on this simplification, the value of w will be equal to the value of ϕ in this region and Equation 1 is reduced to:

$$\frac{ET}{E_P} = \frac{2\phi}{2 + \phi} \tag{3}$$

For values of ϕ less than 0.5 the curve of Equation 1 is closest to Schreiber's and as ϕ approaches 2.0, it is closest to Ol'dekp's. These formulae can be used when Equation 3 is not applicable as seen in Figure 1.

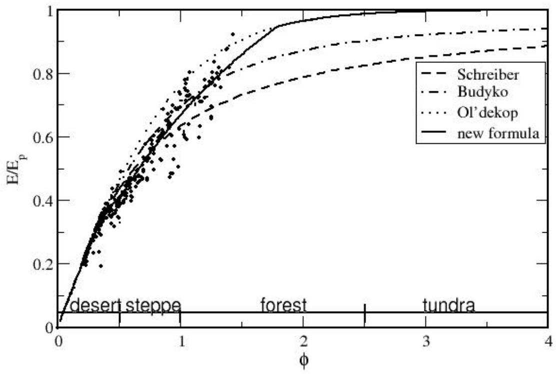

Figure 1
Empirical relationships curves predicted by the formulae of Table 1 and Equation 5 (solid line) together with the data computed for all 314 sub-basins

Figure 1 displays the three empirical curves predicted by the formulae of Table 1 and the curve predicted by Equation 3 (solid line) together with the data computed for all 314 sub-basins (points). The values of the index ϕ were computed after constructing a PET map based on the NOAA Evapotranspiration Atlas (Farnsworth et al., 1982) and observed precipitation data. TOPLATS-simulated ET was used to compute the evapotranspiration efficiency (ET/Ep) values. As seen in Figure 1, the formula of Equation 3 falls between Schreiber's and Ol'dekop's curves, as Budyko observed for real data, but it fits the simulated data better than Budyko's curve. Interestingly, Equation 3 still agrees with Budyko's (1950) assumptions when he first discussed the concept of "geographical zonality" as defined above.

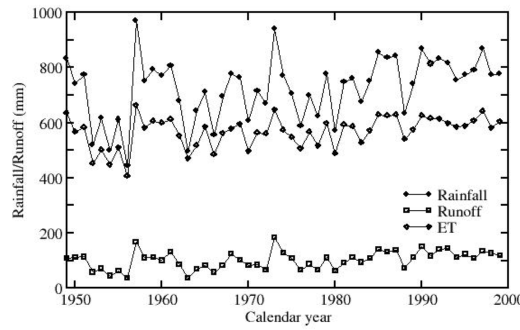

Figure 2
Time series of average annual precipitation, surface runoff, and evapotranspiration for the Arkansas/Red basin for the 1949-2000 period

3. Runoff Sensitivity

Equation 3 can be extended to develop empirical expressions of runoff sensitivity to inter-annual variability in precipitation. For example, the derivative of ET can now be written as a function of the derivative of precipitation. Expressions for the relationship between the means and variances of ET and P can also be derived. Assuming that (a) inter-annual variability of soil moisture is negligible and (b) the PET is constant over time, the following expression for the relationship between the coefficients of variation of annual precipitation and runoff, CVP, and CVR, respectively, can be derived:

$$\frac{CV_R}{CV_P} = \frac{4+\phi}{2+\phi} \tag{4}$$

Similar expressions can derived from the formulae in Table 1. Figure 3 shows the ratio of coefficients of variation computed for the data for the 314 sub-basins. The solid line represents the curve predicted by Equation 4. For small values of the humidity index, the curve of Equation 4 can be joined with the one predicted by Schreiber's formula. Figure 3 shows that runoff sensitivity to precipitation variability increases for drier basins. The data points follow the predicted curve reasonably well. However, the scatter of data points is higher for dry basins, which can be due to the fact that these basins are more affected by the precipitation frequency and the size of individual events. In addition, the soil type and soil storage may play a more significant role in precipitation partitioning, even at the annual time-scale for drier watersheds.

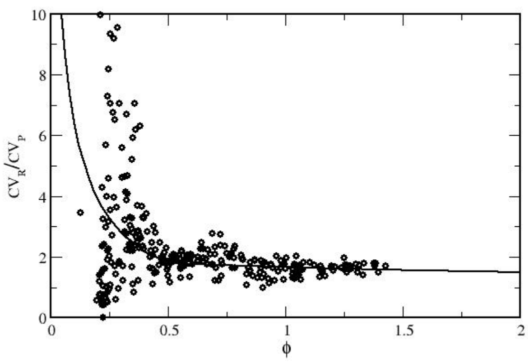

Figure 3

Ratios of coefficients of variation of annual runoff and precipitation for the 314 sub-basins (circles) and the curve predicted by the empirical relationship (solid line)

4. ET Variability

Comparing the average increase in annual precipitation and surface runoff indicates that ET in the basin has also increased during the simulation period, as seen in Figure 2. The increase in ET is significant (about 1.4 mm/year), but much less than the increase in precipitation. It has to be noted that observed discharge (USGS, 2005) for the 1949-1982 period, although not a direct measure of the actual runoff, as mentioned earlier due to the effects of water management, shows a decreasing trend. Increased ET over the AR River basin agrees with results reported in previous observation-based studies for several regions of the United States (e.g. Szilagyi et al., 2002 and Walter et al., 2004). It can be expected that an increase in precipitation will result in an increase in runoff and soil moisture, which in turn would lead to increase in ET, except when ET is near-equal to PET. In fact, Berger and Entekhabi (2001) found that, for a given value of potential evaporation, increased precipitation is the most important variable that leads to increased evaporation when they compared the effect of several variables related to a basin's climate, geomorphology, and lithology. It is worth mentioning here that several studies have indicated decreasing trends in PET, using pan evaporation, over wide regions in Asia, Europe, and South and North America (Lawrimore and Peterson, 2000; Chattopadhyay and Hulme, 1997; Quintana-Gomez, 1997). However, Brutsaert and Parlange (1998) showed that decreasing pan evaporation could actually be an indicator of increased terrestrial evaporation.

Equation 3, and other equations in Table 1, can be extended to show that the interannual variance in ET is always smaller than the variance in precipitation. Again by differentiating Equation 3, squaring, and averaging over time, the following expression can be obtained, assuming that PET is constant:

$$\frac{\sigma_E}{\sigma_P} = \frac{4}{(2+\phi)^2} \qquad (5)$$

Equation 5 suggests that the ET variance decreases with increase of the basin wetness for a given precipitation variance. For very dry basins, where ET is controlled by available moisture from precipitation, the ET variance approaches the precipitation variance. For very wet basins, where ET approaches its potential value, the variance becomes negligible. In Figure 4, the ratio of standard deviations of computed ET values and observed precipitation for the 314 sub-basins are plotted together with the curve predicted by Equation 5, joined with Schreiber's formula for very small values of the humidity index. It is clear from Figure 4 that the empirical formula reasonably predicts the relationship between the precipitation and ET variabilities.

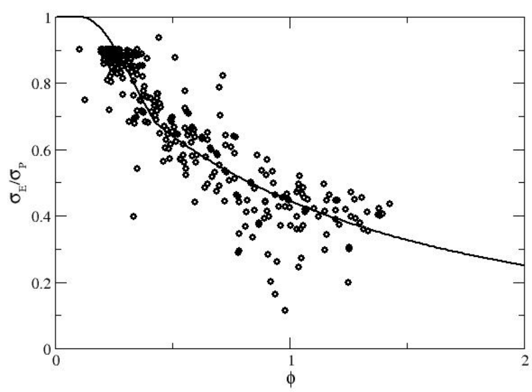

Figure 4

Computed ratios of standard deviation of annual evapotranspiration and precipitation for the 314 sub-basins (circles) and the curve predicted by the empirical relationship (solid line)

5. Summary and Conclusions

A new empirical model that predicts the control of available energy and precipitation in determining values of ET was developed. The model is a very simple one-parameter formula that takes into account the expected dominant vegetation type while maintaining Budyko's assumptions related to the concept of "geographical zonality". Computed ET values for 314 sub-basins agree reasonably well with the empirical model predictions. The model was extended to develop two other formulae that were used successfully to predict the sensitivity of runoff and ET to precipitation variations. The empirical model can serve as a useful tool to estimate the changes in various components of the hydrologic cycle as a result of climate change.

This is an ongoing study and additional sub-basin validations and analyses are planned. Forcing data at higher resolutions, e.g. reliable radar-estimated precipitation, will be used as they become available. The preliminary results presented here make us comfortable with the quality of soil moisture and surface and latent heat flux simulations. These results are being further analyzed, as part of a manuscript in preparation, as the accuracy achieved in surface runoff simulations implies that simulations of the water and energy balance relationships are adequate. The correlation between mean-monthly, seasonal, and annual variations in surface energy fluxes, soil moisture, stream flow, and large-scale atmospheric patterns is also being examined. It is hoped that the results of this analysis will help clarify the sources of long-term hydrologic variability within the AR River basin.

References

Berger K. P., D. Entekhabi, 2001: Basin hydrologic response relations distributed physiographic descriptors and climate. *J Hydrol.*, (247), 169-182

Brutsaert, W. 1982: Evaporation into the atmosphere: theory, history, and applications. D.Reidel, Hingham MA, 299 pp.

Budyko, M. I., 1950: Klimaticheskie faktori vneshnego fiziko-geografichesogo

protsessa (Climatic factors of the external physical-geographical processes). *Gl. Geofiz. Observ*., 19:25–40

Budyko, M.I., 1948: Evaporation under natural conditions, *Gidrometeoizdat*, Leningrad. English translation by IPST, Jerusalem

Chattopadhyay, N. and M. Hulme, 1997: Evaporation and potential evapotranspiration in India under conditions of recent and future climate change, *Agric. For. Meteor*., 87, 55-72

Crow, W.T. and E.F. Wood, 2002: The value of coarse-scale soil moisture observations for regional surface energy balance modeling, *J. Hydrometeor*., 3, 467-482

Lawrimore, J. H., and Peterson T. C., 2000: Pan evaporation trends in dry and humid regions of the United States. *J. Hydrometeor.*, 1, 543–546

Quintana-Gomez, R., 1997: Changes in evaporation patterns detected in northernmost South America. Homogeneity testing. *Proc. Seventh Int. Meeting on Statistical Climatology*, Whisler, BC, Canada, NRCSE, 25–29

Schreiber, P., 1904: Ueber die Beziehungen zwischen dem Niederschlag und der Wasseruhrung der Flusse in Mitteleuropa. *Meteorol. Z.*, **21**, 441-452

Szilagyi, J., G. G. Katul, and M. B. Parlange, 2002: Evapotranspiration intensifies over the conterminous United States. *J. Water Resour. Plann. Manage.*, 127, 354–362

Twine, T.E., W. P. Kustas, J. M. Norman, D. R. Cook, P. R. Houser, T. P. Meyers, J. H. Prueger, P. J. Starks, and M. L. Wesely, 2000: \Correcting eddy-covariance flux underestimates over a grassland. *Agric. For. Meteor.* 103, 279-300

USGS, 2005: US Geological Survey surface-water data for the nation. (Accessed online at http://waterdata.usgs.gov/nwis/sw, August 2005)

Zhang, L., W.R. Dawes, and G.R. Walker, 2001: Response of mean annual evapotranspiration to vegetation changes at catchment scale. *Wat. Resour. Res*., 37 (3), 701-708