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Development of a simple, catchment-scale, rainfall-evapotranspiration-runoff model

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Abstract

Representation of the hydrological interaction between the land surface and the atmosphere requires considerable improvement, particularly for predicting evapotranspiration feedbacks for use in models of the general circulation (GCMs) of the atmosphere. The predictive model developed here attempts to use a water balance approach that extracts information from the masses of catchment-scale time series data available on precipitation, energy-related variables and stream discharge. It begins with a few simple assumptions in order to seek some synthesis of the climate and landscape controls on evapotranspiration and soil moisture feedbacks, and catchment water yields. The model adopts the hydrograph identification approach used in the linear module of the rainfall-runoff model IHACRES but replaces the previous statistically based non-linear evapotranspiration loss module by a catchment moisture deficit accounting scheme. One advantage of this more conceptual approach is that evapotranspiration can be output on the same time step at which precipitation and energy variables are available (such as from GCMs), and this time step can be shorter (e.g. half hourly) than the discharge time step (e.g. daily) used to calibrate the model parameters.

Keywords: Water balance; Hydrograph; Identification; Land surface; Atmosphere

1. Introduction

1.1. Data requirements

Data requirements of existing rainfall-evapotranspiration-runoff models, whether physically based or of a more conceptual lumped parameter type, are usually too demanding for them to be applied outside experimental areas. Yet these models have been applied in catchments where the data required to determine all their parameters are unavailable. This is often the case for Soil-Vegetation-Atmosphere Transfer (SVAT) schemes used in General Circulation Models (GCMs) of the atmosphere (e.g. Sellers et al., 1986; Dickinson et al., 1986). When applying models under these circumstances parameter identifiability problems become apparent.

The Project for Intercomparison of Land-Surface Parameterization Schemes (PILPS) is currently in the process of evaluating the performance of 23 SVATs (Timbal et al., 1997). In phase 2 of the project two experimental sites (Cabauw, The Netherlands, and the HAPEX-Mobilhy site in the south of France) are being used to evaluate the performance of SVAT schemes. The use of these experimental sites reduces the parameter identifiability problem, allowing some evaluation of the process representations in the models. But just as has been argued for physically-based hydrological models (Beven, 1989), one cannot easily measure parameters that must effectively characterize a heterogeneous landscape. Since these SVAT schemes are designed to be used as the land-surface components within GCMs they must be applied over the entire land surface of the planet which makes fulfilling large data requirements impossible at this point in time (Wood et al., 1992). This leads to the use of such models with parameter values in a highly ambiguous state because of a combination of the following: parameter identifiability problems (model insensitive to changes in parameter values); problems in obtaining feasible parameter values; and the problems associated with estimating 'effective' parameters that

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cannot be obtained by direct physical measurement at appropriate space and/or time scales.

The most widely available data used in this sort of modeling are time series of precipitation, temperature and streamflow. With only these data sets the danger of over-parameterization has been recognized (Beven, 1989; Hornberger et al., 1985; Jakeman and Hornberger, 1993; van Genuchten, 1991). It is apparent that, when confined to the above data sets, models need to be kept simple in their process representation. Too many parameters quickly leads to identifiability problems when estimating parameter values and questionable physical characteristics being attributed to the catchment.

1.2. Estimation of evapotranspiration

Estimation of evapotranspiration (ET) can be done in several ways. Two rather obvious ways are to consider it the unknown term in either the water budget equation or the energy budget equation. In any given system at the earth surface, ET is the connecting link between the water budget and the energy budget. The water budget equation, which expresses the conservation of mass in a lumped or averaged hydrological system, can be written as

$$(P - E)A - Q_s - Q_g = \frac{dS}{dt} \tag{1}$$

where P is the mean rate of precipitation on the system, E the rate of evaporation, A the surface area, Q_s the net outflow of surface water, Q_g the net outflow of ground water and S the water volume stored in the system. Solving Eq. (1) for E is not generally practical as unavoidable errors in measuring precipitation and runoff, and the difficulty in measuring groundwater contributions, can often produce large errors in the resulting evaporation.

For a simple lumped system, when effects of unsteadiness in the atmosphere, ice melt, photosynthesis and lateral advection can be neglected, the energy budget is defined by

$$R_n = L_eE + H + G \tag{2}$$

where R_n is the net incoming radiation, L_e is the latent heat of evaporation, E the rate of evaporation, H the specific flux of sensible heat into the atmosphere and G the specific flux of heat conducted into the earth. The energy of the incoming radiation is partitioned at the earth's surface into long wave back radiation, upward thermal conduction and convection of sensible heat, evaporation of water, and downward conduction of heat into the earth Brutsaert, 1982. Evaporation as a latent heat flux plays a crucial role in governing the weather and climate. Unfortunately using Eq. (2) to determine E directly is not sensible due to the assumptions made above, and measurement of the other terms involves significant problems. For example, G depends on land cover, soil type and albedo, and the extrapolation from point measurements to areal values is very difficult.

Another ET estimation technique was first introduced by Dalton (1802). The evaporation rate is determined only by the vapour pressure deficit of the air and the ability of the air to carry the water vapour away from the surface, given as a function of wind speed:

$$E = f_D(\bar{u})(e_s^* - e_a) \tag{3}$$

where E is the rate of evaporation, $e_s{}^*$ the saturation vapour pressure at the temperature of the water surface, e_a the vapour pressure in the air, and $f_D(\bar{u})$ a function of the mean wind speed $\bar{u}$.

Penman (1948) presented a new approach to estimate evaporation by combining ideas from Dalton and energy budget methods. Estimation of ET using techniques such as those of Penman, or amendments of this technique by Monteith (1965) or Priestley and Taylor (1972), can require knowledge of variables such as wind speed, specific humidity, vapor pressure deficit, net radiation, soil heat flux and various vegetation related characteristics, which for this paper we are assuming are not generally available. Clearly the data required to explicitly use these techniques are not available extensively. Discussion of ET estimation methods can be found in Garratt (1992); Linacre (1992); Brutsaert (1982). Various more empirical techniques, which do not require masses of data, were developed by Thornthwaite (1948); Hamon (1961) and others. However, these techniques were generally designed to estimate ET at time scales of a month or longer.

1.3. Measurement of evapotranspiration

Measuring ET is itself a difficult problem with the currently best accepted methods (eddy correlation, Bowen ratio energy balance) being both expensive and labor intensive. Because of this problem these data sets tend to have only small temporal extent. The eddy correlation method allows the direct measurement of latent heat flux (and hence ET) by the correlation of fluctuations of vertical wind speed with fluctuations of vapour density. The Bowen ratio approach relies on the *similarity* principle. This holds that the atmospheric diffusion coefficients for heat, water vapour, momentum and carbon dioxide are equivalent. The Bowen ratio approach then requires an accurate determination of temperature and absolute humidity at two levels in the atmosphere in order to determine ET (Tapper, 1996).

When attempting to verify ET models, one is therefore forced to use either the short term data given by the methods above (e.g. days of hourly data) or longer term data obtained from a catchment scale water balance approach (e.g. decades of yearly data). The type of data used for verification is usually determined by the time scale of interest, though it is worth noting that methods such as eddy correlation are point measurements, whereas the longer term water balance methods obtained using streamflow measurements are an integration of outputs from spatially varying runoff generation and transmission processes within a catchment.

2. Description of the model

The rainfall-ET-runoff model proposed here is based on the structure of the IHACRES metric/conceptual rainfall-runoff model. This model undertakes identification of hydrographs and component flows purely from rainfall, temperature and streamflow data (Jakeman and Hornberger, 1993; Jakeman et al., 1990, 1994). The IHACRES module structure consists of a non-linear loss module, which converts observed rainfall to effective rainfall or rainfall excess, and a linear streamflow routing module, which extends the concept from unit hydrograph theory that the relationship between rainfall excess and total streamflow (not just quick flow) is conservative and linear.

The linear module allows any configuration of stores in parallel or series. From the application of IHACRES to many catchments it has been found that the best configuration is generally two stores in parallel, except in semi-arid regions or for ephemeral streams where often one store is sufficient (Ye et al., 1997). In the two-store configuration, at time step k, quickflow, $x_k^{(q)}$, and slowflow, $x_k^{(s)}$, combine additively to yield streamflow (discharge), q_k:

$$q_k = x_k^{(q)} + x_k^{(s)} \tag{4}$$

$$x_k^{(q)} = -\alpha_q x_{k-1}^{(q)} + \beta_q U_k \tag{5}$$

$$x_k^{(s)} = -\alpha_s x_{k-1}^{(s)} + \beta_s U_k \tag{6}$$

where U_k is the effective rainfall. The parameters α_q and α_s, can be expressed as time constants for the quick and slow flow stores, respectively:

$$\tau_q = -\Delta/\ln(-\alpha_q) \tag{7}$$

$$\tau_s = -\Delta/\ln(-\alpha_s) \tag{8}$$

where Δ is the time step (daily here).

Parameters expressing the relative volumes of quick and slow flow can also be calculated:

$$V_q = 1 - V_s = \frac{\beta_q}{1+\alpha_q} = 1 - \frac{\beta_s}{1+\alpha_s} \tag{9}$$

The previous IHACRES loss module used a statistical approach to account for antecedent soil moisture conditions and ET losses. Here this module is replaced by our more physically based catchment moisture store accounting scheme which uses rainfall and temperature as inputs and provides ET and rainfall excess as outputs, creating the overall model structure shown in Fig. 1.

The catchment moisture store accounting scheme calculates Catchment Moisture Deficit at time step k, CMD_k, according to

$$CMD_k = CMD_{k-1} - P_k + E_k + D_k \tag{10}$$

where P is the precipitation, E is the ET loss and D is the drainage. Such a scheme has been used previously by Littlewood (in preparation). CMD is zero when the catchment is saturated and increases as the catchment becomes progressively drier.

Effective rainfall is calculated from

$$U_k = \begin{cases} D_k & CMD_k \geq 0 \\ D_k - CMD_k & CMD_k < 0 \end{cases} \tag{11}$$

Suitable parameterizations for both E_k and D_k were sought that minimized the number of parameters needed and for which the only data requirements are temperature, rainfall and streamflow. Several parameterizations were tried and the results for the simplest successful model are presented below.

For ET modeling, techniques vary considerably in the postulated relationship between ET and temperature, T. The effects of vegetation on ET have been represented in various ways, most notably by the incorporation of a 'surface resistance' in the Penman–Monteith equation. This surface resistance has itself been estimated in many ways, commonly with a dependence on the available soil moisture such as that given by Stewart (1989) where $E_T = \exp[K(\delta\theta - \delta\theta_{\max}]$ in which $\delta\theta$ is the soil moisture and K is a constant. The parameterization here is defined by

$$E_k = c_1 T_k \exp(-c_2 CMD_k) \tag{12}$$

Eq. (12) has ET directly proportional to temperature and decreasing exponentially as CMD increases. So higher

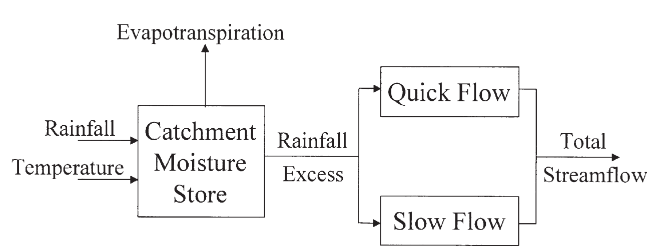

Fig. 1. Structure of the rainfall-ET-runoff model.

temperatures will result in larger ET losses provided sufficient soil moisture is available. Eventually, however, the authors wish to enhance the accuracy of the E_k term by utilizing other energy-related variables in its forcing (e.g., humidity). It has been noted that "natural" controls on evaporation may constrain the modelled values to be reasonable even for fairly simplistic evaporation functions (Wood et al., 1992).

It was assumed that drainage is not temperature dependent. As with the modeling of ET, several parameterizations of drainage were investigated. One of the simplest investigated was

$$D_k = \begin{cases} \dfrac{-c_3}{c_4} CMD_k + c_3 & CMD_k < c_4 \\ 0 & CMD_k \geq c_4 \end{cases} \tag{13}$$

where c_3 and c_4 are non-negative constants. The drainage equation allows water to escape to the stream even when a moisture deficit exists within the catchment. Parameter c_3 represents the maximum drainage that can occur whilst a moisture deficit exists. Parameter c_4 represents the maximum CMD that can occur before water ceases draining to the stream. To maximize simplicity and minimize computing time required, $[CMD_{k-1} - P_k]$ was used in place of CMD_k in the ET and drainage equations, along with the extra requirement that if $[CMD_{k-1} - P_k] < 0$ then it is considered as equal to zero.

To measure the performance of the model estimate of streamflow, $\hat{q}_i$, two performance statistics are used: the bias (B) and the observed streamflow variance explained (R^2). These are defined as

$$B = \frac{1}{n} \sum_{i=1}^{n} (q_i - \hat{q}_i) \tag{14}$$

$$R^2 = 1 - \alpha_e^2 / \alpha_q^2 \tag{15}$$

where α_e^2 and α_q^2 are the variance of the model residuals $(q_i - \hat{q}_i)$ and of the observed streamflow respectively.

Unfortunately, adequate ET data for the catchments investigated were not available so no direct performance measure of the ET estimate could be made. It should be noted though that low bias in predicting streamflow indicates that losses from the catchment were well accounted for overall. The distribution of these losses from day-to-day remains in question.

3. Description of catchments

To investigate the applicability of the new non-linear module for ET loss and discharge prediction, it was applied on a daily time step to three catchments covering a range of space scales and climatic conditions. The hydrometeorological characteristics of the selected catchments are given in Table 1.

Two small catchments were selected from the Coweeta Hydrological Laboratory in the United States. Coweeta is located in the Nantahala Mountains of western North Carolina. Coweeta is a relatively humid area and both streams are perennial. Rainfall occurs throughout the year. Watershed 36 is a high-elevation, steeply sloping catchment with shallow soils, a high annual yield and a large proportion of quick flow (Swift et al., 1988). Watershed 34 is a mid-elevation catchment with somewhat deeper soils and, consequently, substantially more delayed flow. The Coweeta site is covered predominantly by hardwood forest. Details of the physical characteristics of the Coweeta catchments are given by Swank and Crossley (1988).

The final catchment used in this investigation is Scott Creek at Scotts Bottom. It is situated near the city of Adelaide in South Australia. Scott Creek is an ephemeral catchment, and has a warm climate with hot, dry summers. The catchment is dominated by winter rainfall and the streamflow can cease during summer. Soils in the catchment are made up of a sandy loam top layer with a clay subsoil. More than 80% of the catchment is covered by grass. It is part of a network of benchmark catchments in Australia. More details can be found in Chiew and McMahon (1992).

4. Results

The model was calibrated on a daily time step over periods of two years on all three catchments and the results are given in Table 2. The performance in terms of goodness of fit criteria, R^2 and B, is more than reasonable. The models were then used to simulate the daily streamflow over the subsequent 5 years and the results are given in Table 3. Fig. 2a-c compares the modeled streamflow with observed streamflow for the five year simulation periods. Fig. 2a indicates that the model had some difficulty reproducing hydrograph recessions when simulating streamflow in Watershed 34 even though the performance statistics remain quite good. Model residuals for Watershed 34 show a slowly varying component due to under-prediction of baseflow at times. Fig. 3 shows the modelled catchment moisture deficit changing over the simulation period in the Scott Creek catchment. It clearly demonstrates the expected seasonal nature of the catchment moisture store. Each year CMD builds over summer, a period of little rain and higher temperatures, and then it is reduced quickly when the rains return in winter. Comparison of Fig. 3 and Fig. 2c demonstrates the strong relationship between CMD and streamflow. Fig. 4 shows the ET predicted for the Scott Creek catchment during the simulation period. Fig. 5

Table 1
Hydrometeorological characteristics of catchments investigated

Catchment	Area, km^2	Precipitation, mm/yr.	Average daily maximum Temperature, °C	Annual Yield, %
Watershed 34 (Coweeta, North Carolina)	0.33	2012	19.6	46
Watershed 36 (Coweeta, North Carolina)	0.49	2012	19.6	64
Scott Creek @ Scotts Bottom	27	1090	14.8	20.5

Table 2
Calibration results for the three catchments

Catchment	Period	c1	c2	c3	c4	τ_q (days)	τ_s (days)	V_s	R^2	B (mm/d)
Watershed 34, (Coweeta, North Carolina)	1/1/82–31/12/83	0.2	0.00	4	55	2.16	85.75	0.857	0.90	0.00
Watershed 36, (Coweeta, North Carolina)	1/1/82–31/12/83	0.15	0.00	5	95	1.59	43.79	0.324	0.87	−0.05
Scott Creek @ Scotts Bottom	24/2/70–23/2/72	0.24	0.01	2	8	0.71	53.07	0.475	0.88	0.00

Table 3
Simulation results for the three catchments

Catchment	Period	R^2	B (mm/d)
Watershed 34, Coweeta	1/1/84–31/12/88	0.84	0.23
Watershed 36, Coweeta	1/1/84–31/12/88	0.76	−0.26
Scott Creek	24/2/72–23/2/77	0.79	0.00

compares the mean monthly rainfall and ET from 1961 until 1990 for Coweeta Watersheds 34 and 36 as calculated by our model and that calculated at a representative site within the Coweeta Hydrological Laboratory using the Thornthwaite (1948) method (Swift, 1998). This provides a broadscale check on the ET predicted by our model.

5. Discussion and conclusions

The parameterizations used for both ET and drainage here are the simplest of those tested so far which have been judged to perform well. Indeed the model tested here performed as well if not better than several other more complex parameterizations investigated, indicating that higher levels of parameterization are not warranted on these data sets. However, selection of the most appropriate parameterization in the future will require more extensive testing on longer data sets and across different hydroclimatologies.

Table 2 demonstrates the model's ability to fit the streamflow characteristics of each catchment when calibrated on only two years of data, with R^2 ranging from 0.87 to 0.90. It appears to perform equally well in humid regions as in drier regions. Notably, the bias remains low in all cases. The IHACRES models for Coweeta Watershed 34 and Scott Creek produce zero bias, and Coweeta Watershed 36 yielded a bias of − 0.05 mm/day. These low biases indicate that the total ET losses have been estimated well. Measured ET, preferably on a daily basis, is required however if one is to have confidence in the estimated ET losses per day. The parameter c_4 was the only parameter to demonstrate significant identifiability problems in the calibration procedure. This emphasizes the difficulty in establishing a clear value for the CMD at which water ceases to drain to the stream, especially in humid catchments where modelled moisture never ceases to drain to the stream during the calibration period. A longer calibration period may reduce this uncertainty.

Clearly from Table 3 the model performs well under simulation conditions, with R^2 ranging from 0.76 to 0.84. As expected the R^2 values are lower than during the calibration period but still remain quite high. For both Coweeta catchments the bias has increased significantly but remains small relative to the streamflow. The simulation period bias for the Scott catchment remains zero, indicating once again that the ET losses have been estimated well overall. Fig. 2 demonstrates the simulated model fit for the three catchments. Both Fig. 2b and c show consistently good fits while Fig. 2a shows the model had some trouble fitting the hydrograph recession for Watershed 34.

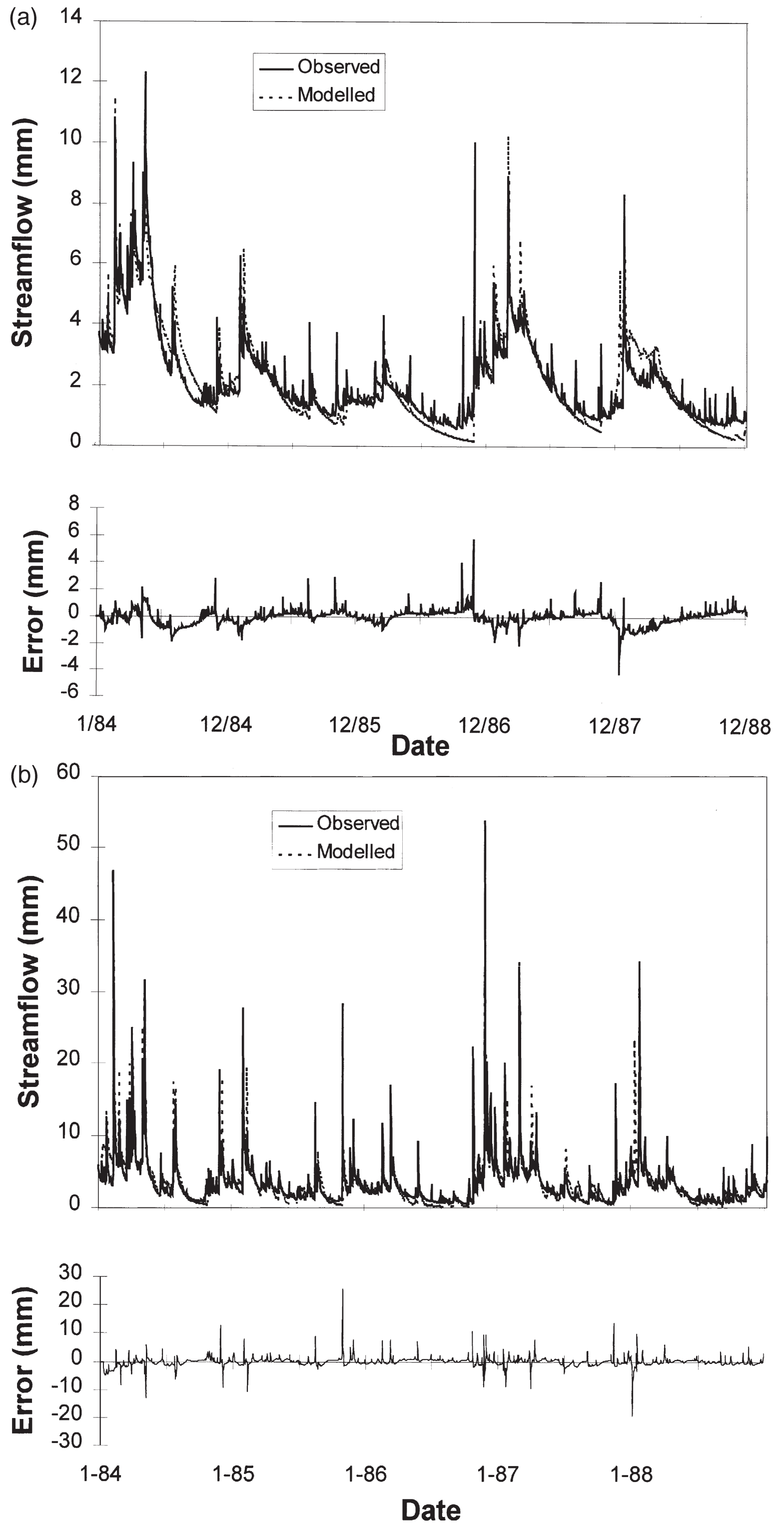

Fig. 2. (a) Simulation model fit for Coweeta Watershed 34. (b) Simulation model fit for Coweeta Watershed 36.

Fig. 4 shows an example of the daily ET losses predicted by the model for the simulation period in the Scott catchment. Comparison with Fig. 3 shows the influence of CMD on the evaporation. The seasonal nature of ET is not as obvious as for CMD. The lowest values occur towards the end of summer when CMD is still high and temperatures are falling, demonstrating the trade off between available soil moisture and available energy which drives the evaporative process.

Fig. 5 shows that the rainfall observed for Watersheds

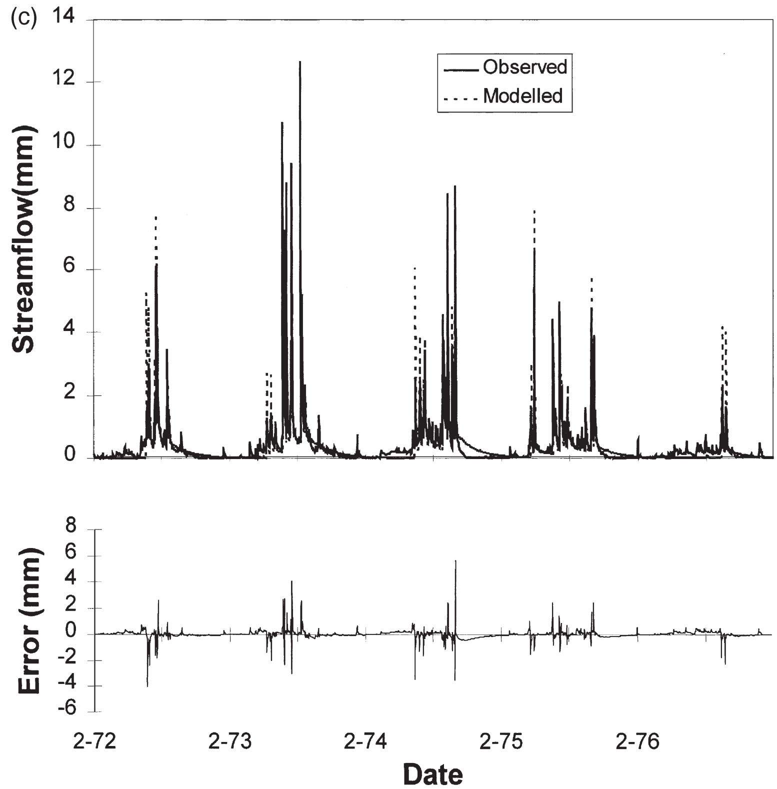

Fig. 2. (c) Simulation model fit for Scott Creek.

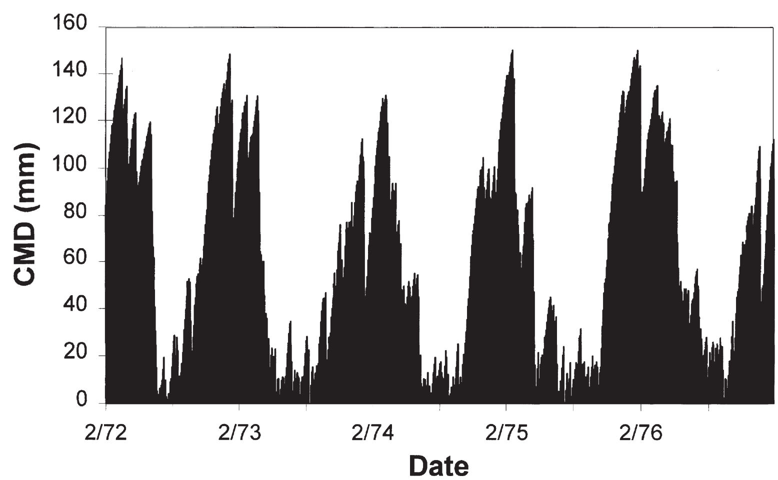

Fig. 3. Catchment moisture deficit in the Scott Catchment.

34 and 36 is higher than that for the representative site within the Coweeta Hydrological Laboratory, with Watershed 36 expectedly having the most rainfall since it is at the highest elevation. The representative site rainfall and ET almost coincide in July suggesting that in some years soil moisture may be low enough to affect transpiration of the vegetation. Similarly Watershed 34 rainfall and ET almost coincide in August, again suggesting transpiration of the vegetation may be affected in some years. Watershed 36 maintains a much larger gap between rainfall and ET, so the vegetation is less likely to be stressed in this Watershed. ET estimated by our model follows a very similar pattern to that estimated by the Thomthwaite method on a monthly time step. For Watershed 36 there is a much larger gap between rainfall and ET losses which is consistent with its annual yield of 64% being almost 50% higher than the annual yield of Watershed 34 (46%).

From Eq. (13) we can see that if parameter c_3 were zero then effective rainfall only occurs when CMD is forced below zero by the rainfall. The non-linear module of IHACRES is then acting as a threshold mechanism

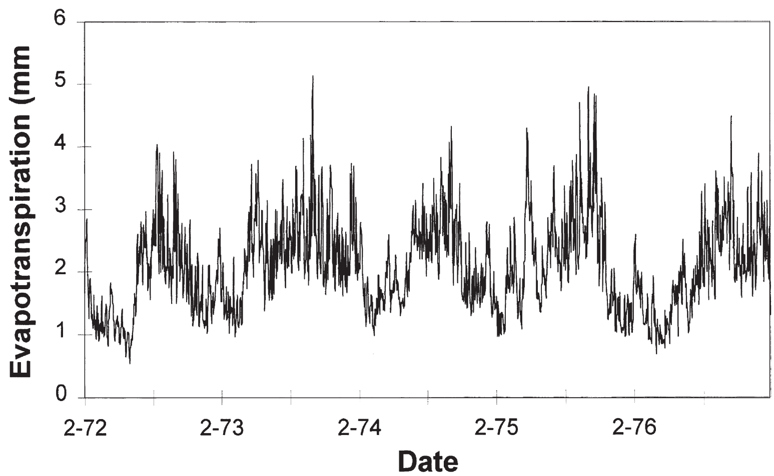

Fig. 4. Evapotranspiration in the Scott Catchment.

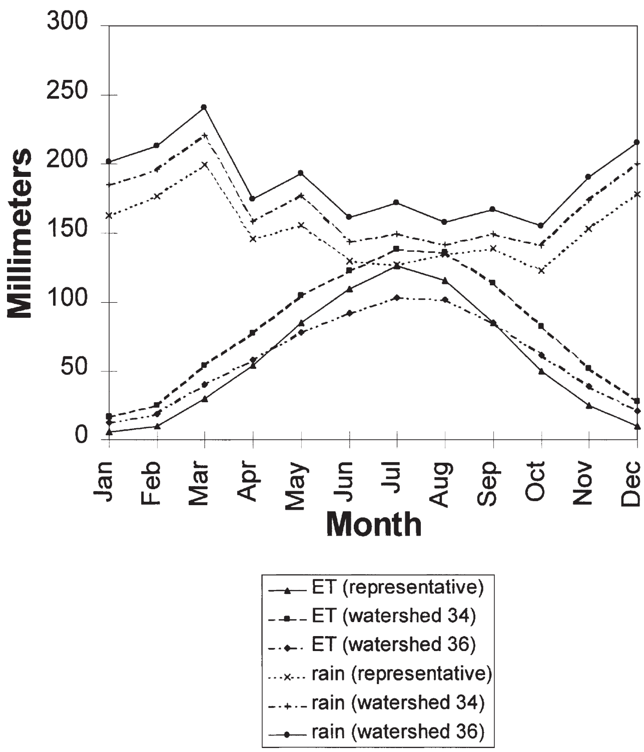

Fig. 5. Mean monthly precipitation and evapotranspiration: 1961–1990.

adjusted by the antecedent moisture conditions. In that case the model collapses down to five parameters and the non-linear module acts similarly to the bucket model of Manabe et al. (1965).

Table 2 also shows that the best calibration results for both Coweeta Watersheds occur when c_2 is zero. This means that the modelled ET has no dependence on CMD. For the period of calibration this is not entirely surprising as Coweeta has relatively high rainfall and so the CMD may rarely get high enough to affect the transpiration of the vegetation and therefore ET.

By using this semi-physical approach as described in the accounting scheme in (10) the physical interpretation of the parameters will hopefully be made easier than with the previously used statistical approach in IHACRES. It is hoped that, by studying catchments with gauged discharge data and possibly measured ET data, improved relationships can be constructed between the model parameters and landscape attributes, as has been attempted with some success by Post and Jakeman (1996), using the statistical type of non-linear module. These relationships would permit the simulation of streamflow and ET for changes in land use and for ungauged catchments, at least within a similar region (see Jakeman et al., 1994). The construction of these relationships is only made possible by the minimal parameterization of catchment hydrologic response, with the model having at most 7 parameters. Three of these parameters are the routing parameters in (8) and (9).

Only temperature and catchment moisture are investigated as forcing variables for ET here. However it is envisaged that other important variables, such as humidity, net radiation and wind speed, will be considered as the methodology is developed further.

When refined, this approach to modeling rainfall-ET-runoff could be used to provide the land surface feedbacks for climate models. It will be necessary to spatially disaggregate climate forcing variables from the grid scale down to the relevant catchment scale (e.g. Bates et al., 1998; Bogardi et al., 1993; Hughes and Guttorp, 1994), and to temporally disaggregate daily ET and energy feedbacks from the land surface to the atmosphere down to a sub-daily time step of the order of an hour. These scale issues are recognized problems (Avissar, 1995; Beven, 1991; Kalma and Sivapalan, 1995; Raupach and Finnigan, 1995; Wood et al., 1990) and are the focus of continuing research. The accounting scheme in (10) potentially permits the temporal disaggregation. Provided the ET expression, such as (12), is applicable for the sub-daily time step, the model can be calibrated by fitting average daily or temporally higher—average discharge data.

In this approach, parameterization of all gauged catchments must be undertaken off-line, but once inserted in a GCM the overall computational complexity is little greater than that of a bucket model. These parameterizations are valid at least for historically tested climate conditions, vegetation and land use status. With the construction of relationships between the model parameters and landscape and vegetation attributes the model parameterizations would gain a much wider applicability.

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