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CELMOD5 — a semi-distributed cell model for conversion of rainfall into runoff in semi-arid watersheds

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CELMOD5 — a semi-distributed cell model for conversion of rainfall into runoff in semi-arid watersheds

Arnon M. Karnieli*[,a], Mordechai H. Diskin[b], Leonard J. Lane[c]

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Abstract

This paper introduces the general outline of CELMOD5, a parametric, semi-distributed, quasi-linear model, for conversion of rainfall into surface runoff. The model considers the watershed as a series of interconnected cell units, each representing a specific portion of the area of the watershed. In contrast to grid models using a large number of rectangular elements or cells, the number of cells in CELMOD is relatively small and their boundaries are chosen according to the watershed topography. For each cell, the program computes the surface runoff hydrograph at the cell outlet, related to a specified record of total rainfall data at a number of rain gages. If measured surface runoff data are available for some locations in the watersheds, the program can compare these data with computed values of surface runoff at the corresponding points of the cell model. Detailed descriptions are provided for the main model procedures — computation of rainfall excess, conversion of rainfall excess into surface outflow, routing the channel inflow and subtraction of channel losses. Special attention is given in this model to the specific conditions of arid or semi-arid watersheds.

This paper is also concerned with a technique for calibrating and testing a forecasting model of storm hydrographs with emphasis on two objective functions — runoff volume and peak discharge. A method for evaluation procedure is presented based on the following five steps: trial and error calibration; sensitivity analysis; bilinear interpolation optimization; testing the model on different storm events; testing the model on a different watershed. Results are presented for all the larger storm events with reliable data during 12 years in two subwatersheds of the Walnut Gulch Experimental Watershed in southeastern Arizona. The evaluation procedure is demonstrated for one particular rainfall–runoff event.

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1. Introduction

Semi-distributed models in surface hydrology were developed in response to dissatisfaction with the lumped models used to represent the relationship between rainfall and runoff. The main criticism of the lumped models is that they require an assumption that the rainfall input is uniformly distributed over the area of the watershed or, at least, that the spatial distribution is constant (Diskin, 1964). This condition is rarely realized. It may be speculated that some of the non-linearities in the rainfall–runoff response reported in technical papers are actually due to the non-uniform or non-constant distribution of the rainfall or rainfall excess inputs.

Cell models are not new in hydrology. The first semi-distributed model was probably that proposed by Dooge (1959), although he suggested that the model should be used to derive a unit hydrograph rather than using it directly as a rainfall–runoff model. Other early contributors to the development of semi-distributed models were Laurenson (1964), Mein et al. (1974) and Boyd (1978, 1981). In all models, including CELMOD, the cells are large and their number is relatively small. The cells are, in fact, subwatershed units represented by lumped model elements. This is in contrast to grid models, in which the equations of motion are applied to very small rectangular elements or cells. The purpose of this paper is not to review the various models proposed over the years, but to describe the structure of one semi-distributed model and results obtained with it. The model considered is CELMOD, originally developed while the first author was visiting the University of Arizona at Tucson. Tracing the development of the CELMOD, it is possible to identify five versions of the model. These are described briefly in the following section.

2. Development of the cell model

CELMOD was originally proposed as a semi-distributed, parametric, routing model for conversion of rainfall excess into direct surface runoff (Diskin and Simpson, 1978). In that form, it consisted of a system of interconnected cell units, each representing a definite portion, or area unit, in the watershed (Fig. 1). The boundaries of these area units were chosen along water divides between adjacent internal streams so that if there is inflow to a cell from adjacent upstream cells, this inflow occurs only through a small number (0–4) of channels crossing the cell boundaries. The outflow from each cell unit is by a single outlet channel.

All cell units receive rainfall excess input which is variable with time but assumed to be uniform across each area unit. Two types of cells are recognized, exterior or interior. Exterior cells, defined as cells without any channel inflow, have only one input, namely the rainfall excess input. Interior cells are cells receiving channel inflow from upstream cells in addition to the rainfall excess input. The interconnections between the cells form a branching tree-like structure, which reflects the main drainage pattern of the watershed (Fig. 1).

In the first cell model, called herein CELMOD1, the two inputs of each cell were summed to form a single input which was routed through the model cell to form the

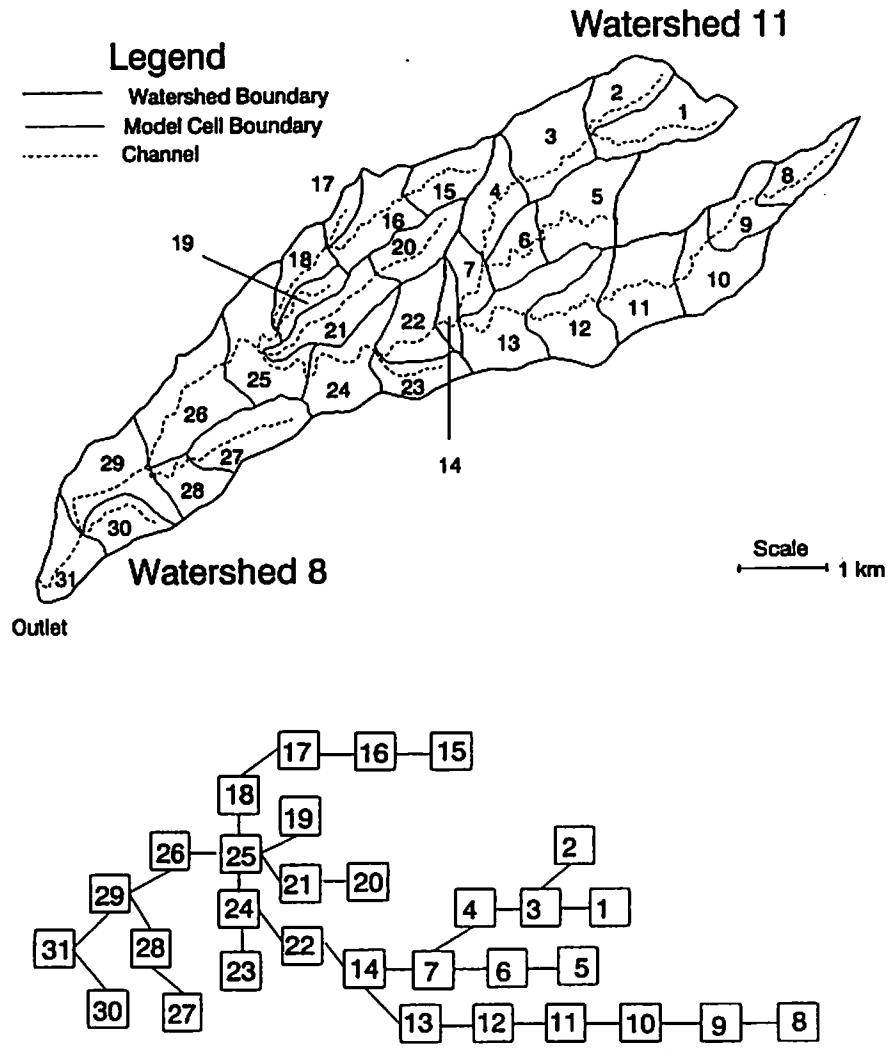


Fig. 1. Subwatersheds 8 and 11 of the Walnut Gulch Experimental Watershed, divided into model cells.

output of the cell. This, in turn, formed part of the channel input for the next cell downstream of the cell considered. In CELMOD1 the cell model was assumed to be a single linear reservoir. Furthermore, all cells were taken to have the same value of reservoir constant, which is the only parameter of the model. By performing the routing computations for each cell in a sequence, starting with upstream exterior cells and continuing with downstream cells to the output of the watershed, a surface runoff hydrograph is obtained. A detailed description of CELMOD1, with some examples of its applications, has been given by Diskin and Simpson (1978) and by Diskin (1979).

Despite the crudeness of the model and the use of only one parameter, the cell model produced fairly good results, both for stationary and for moving rainfall patterns. It was realized, however, that better results could be obtained if the two inputs were treated differently in the cell model and if each cell was assigned its own set of parameters according to its size and shape. The result of these improvements is a model that will be called herein CELMOD2, described by Diskin et al. (1984).

The structure of each cell model in CELMOD2 consists of two procedures in parallel. The conversion of rainfall excess into surface runoff at the outlet of each cell is performed by routing the rainfall excess hyetograph through a pair of unequal linear reservoirs in series. The conversion of the channel inflow to channel outflow is carried out by routing the inflow hydrograph through a combination of a linear channel and a linear reservoir in series.

The next version of the cell model, CELMOD3, was based on the same basic model cell structure as used in CELMOD2, with the addition of an option of using, as input to the model, the total rainfall hyetographs for a number of rain gage locations, instead of the rainfall excess input needed for each cell in CELMOD2. This required the addition of a procedure for separating rainfall excess from total rainfall.

CELMOD4 was a new version of the cell model developed at the Watershed Research Center (US Department of Agriculture (USDA)) in Tucson, Arizona, to take into account the specific conditions of arid or semi-arid watersheds. The main feature included in this version (Diskin, 1984) is a procedure for simulating the channel infiltration losses in the main channel of the area represented by each cell. The channel loss procedure is empirical, but it is based on results reported by Lane et al. (1971) and Murphy et al. (1972). The channel loss operation consists of subtracting two components from the routed channel hydrograph (see Fig. 3 below) using equations similar to Horton's infiltration equation. The output produced by the channel loss operation is the part of the routed hydrograph which is in excess of the sum of the two components of the channel loss.

CELMOD4 served as a basis for a new version of the semi-distributed cell model named herein CELMOD5 (Karnieli, 1988). The main feature included in the new version is an advanced procedure for computing the rainfall excess input for each cell from the total rainfall data for one or several rain gages. The input for the new version model is the total rainfall hyetographs for a number of rain gages, inside and near the watershed boundaries, and the output is the direct surface runoff at the outlet of the watershed, or at intermediate points in it. A detailed description of CELMOD5 and the procedures used in every step of the calculations are given in the next section.

3. Outline of CELMOD5

3.1. Structure of the model

The model consists of a series of interconnected cell units, each representing a specific portion of the area of the entire watershed. The interconnections between the cells form a tree-like structure which reflects the main drainage pattern and the

topography of the basin. Each cell may be connected to and be receiving channel runoff from one or more upstream cells, but it is connected and drains to only one cell downstream. An example of the branching structure of a certain watershed showing the boundaries of sub-units is given in Fig. 1. As noted above, two types of cells can be distinguished in the model: exterior cells and interior cells. An exterior cell has no channel inflow and is usually, but not necessarily, located on the watershed divide. Cell 20, for example, is located internally in the watershed, but with absence of channel inflow it is classified as an exterior cell. As there is no channel inflow into an exterior cell, the rainfall hyetograph over the cell area is the only input to this cell. An interior cell has one or more inflow channels crossing its boundary and supplying channel inflow to the cell. The interior cell receives as inputs both the rainfall hyetograph over the cell area and the channel inflow hydrographs from upstream channels.

Each cell in the model is composed of a number of elements that transform the cell's inputs into a single output in the form of a cell outflow hydrograph. This outflow hydrograph is the sum of the surface outflow hydrograph and the channel output hydrograph. The program starts with the upstream exterior cells, and proceeds to the downstream cells. In an interior cell with multiple inflows, all the inflow hydrographs are added together to form a single input hydrograph which is routed through the cell. The computation ends in the most downstream cell having its outlet at the outlet of the watershed.

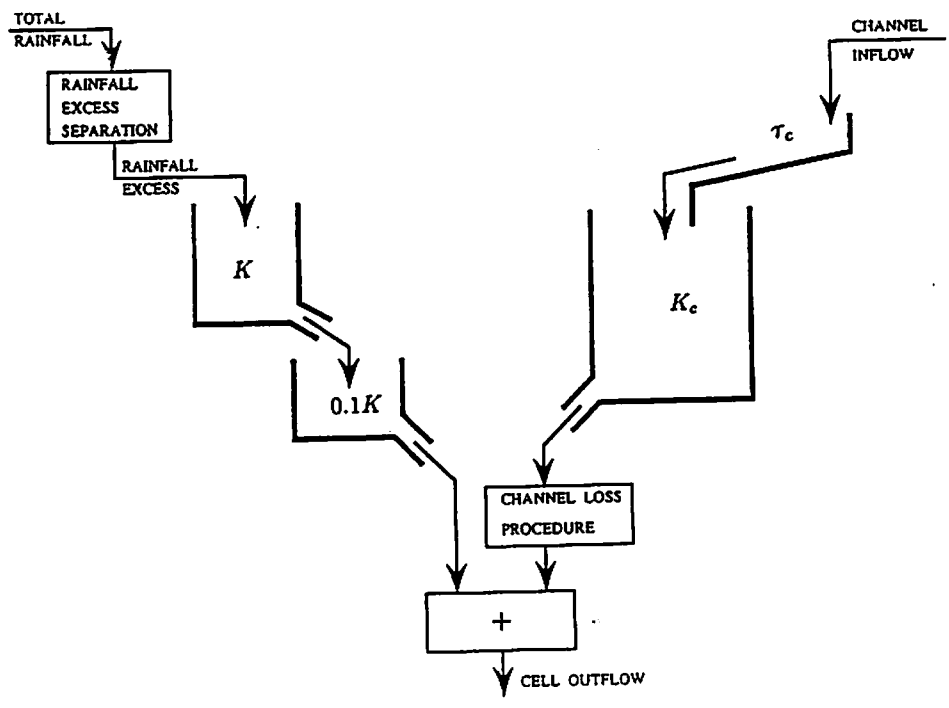


Fig. 2. Internal structure of a cell in CELMOD4 and CELMOD5.

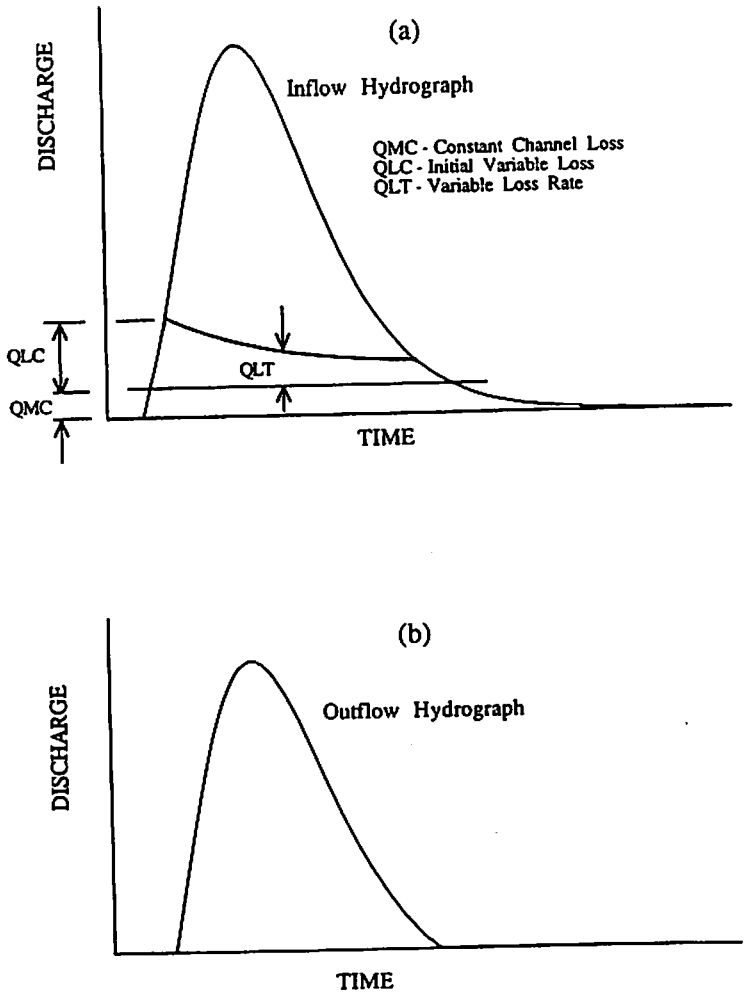


Fig. 3. Procedure used for subtracting channel losses and resulting outflow hydrograph for a given cell.

All cells have the same internal structure in terms of the elements that make up the cell and their interaction in producing the cell output (Fig. 2). They differ only in the size, topographic features and soil properties of the area units represented by them. This enables the user to define, work with and optimize only one set of parameters that represent all the cells in the model. This set of parameters is termed the model parameters. Using these model parameters, each cell is assigned individual values of the parameters, which are termed cell parameters. The values of the cell parameters are influenced by the area of the cell or by the length of the channel within the cell relative to the average cell area or average channel length. This unique feature of the cell model is of great help in the optimization and calibration procedures. However, if

the spatial distribution of one or more parameters is known, the model can accept individual values of the parameters for each cell.

3.2. Computation of rainfall excess

The equation adopted for the computation of rainfall excess in CELMOD5 is the Green and Ampt (1911) equation:

$$\frac{dF}{dt} = f = K_s + \frac{K_s \cdot SM}{F} \tag{1}$$

where F is the cumulative infiltration $[L]$ in time t, f is the infiltration rate $[L\,T^{-1}]$, K_s is the saturated hydraulic conductivity at the transmission zone $[L\,T^{-1}]$, and SM is the effective matric potential $[L]$. The equation was modified by Mein and Larson (1971, 1973) and Chu (1978). These modifications allow the calculation of either the infiltration or the rainfall excess under unsteady rainfall conditions.

The complete procedure for calculating the rainfall excess under variable rainfall has been discussed by Karnieli (1988). Eq. (1) has two parameters:

(1) the saturated hydraulic conductivity, K_s, which is a function of the watershed characteristics, such as hydraulic conductivity, the canopy cover and the ground cover of the area represented by the rain gage. It should be noted that these features vary in the space domain but are assumed to be constant in the time domain.

(2) The effective matric potential, SM, which is calculated from the soil texture and the residual soil moisture.

In CELMOD5, it is preferable to break the SM parameter into two separate parameters: S, the difference in the average effective capillary potential across the wetting front $[L]$; M, the difference in average soil moisture across the wetting front $[L^3\,L^{-3}]$. S depends on the soil texture, whereas M is a function of both the soil texture and the residual soil moisture.

The residual moisture is the only parameter in CELMOD5 which is time dependent. It is proposed that its value be calculated outside the model either by an API-type procedure (Karnieli, 1988) or by soil water balance calculations (Karnieli and Ben-Asher, 1993). Both procedures are based on daily rainfall data for days before a particular event and the inter-arrival time between storms. Only the final values from these procedures, the residual soil moisture values for each area represented by a rain gage, are transferred to CELMOD5 and are used for the rainfall excess calculations.

The rainfall excess is computed separately for each rain gage for which data are available. Therefore, the rainfall excess data are initially applied to areas defined by a Thiessen weighting procedure. The transformation of the rainfall excess from the Thiessen polygons to appropriate model cells is done by GIS methods such as overlaying the Thiessen polygons on the model cells. This procedure, explained in detail in Karnieli (1991), is applied to each time interval for which rainfall data are available, and results in an individual rainfall excess hydrograph for each cell in the model.

3.3. Conversion of rainfall excess to surface runoff

The rainfall excess function for each cell is converted to a surface runoff hydrograph by routing it through a pair of unequal linear reservoirs in series (Fig. 2). The two reservoirs represent the combined action of overland flow and local channel flow within the area represented by the cell. To keep the number of model parameters to a minimum, two assumptions were made:

(1) The storage coefficient of one reservoir, K, is taken arbitrarily to be 0.1 of the value of the coefficient for the second reservoir; thus

$$K_2 = 0.1K \tag{2}$$

(2) The value of the storage coefficient for each cell is assumed to be proportional to the square root of the area represented by the cell; thus

$$K = A_{kc}\left[\frac{A_c(j)}{A_m}\right]^{1/2} \tag{3}$$

where A_m is the average of the areas of all cells $[L^2]$, $A_c(j)$ represents the area of individual cell j $[L^2]$, A_{kc} is the storage coefficient model parameter for the cell of average size $[T]$ and K is the corresponding storage coefficient cell parameter.

The actual operation of conversion of rainfall excess to surface outflow for each cell is carried out by a numerical convolution procedure involving a unit hydrograph $U(t)$ derived for a pair of unequal linear reservoirs in series (Dooge, 1973):

$$U(t) = \int_0^t \frac{1}{K} e^{-\tau/K} \frac{1}{K_2} \exp[-(t-\tau)/K_2] d\tau = \frac{e^{-t/K} - \exp(-t/K_2)}{K - K_2} \tag{4}$$

Introducing Eq. (2) into Eq. (4) yields

$$U(t) = \frac{e^{-t/K} - e^{-10t/K}}{0.9K} \tag{5}$$

It should be noted that the hydrograph described by Eq. (4) or by Eq. (5) has a more realistic shape than that derived for a single linear reservoir (Dooge, 1973). This is because the former includes a sloping rising limb, whereas the hydrograph describing the single reservoir has a vertical rise at the origin. The end result of the numerical convolution procedure is a surface runoff hydrograph for each cell, specified by its ordinates at the end of all time intervals.

The surface runoff hydrograph for each interior cell is combined with the channel outflow hydrograph of that cell by adding the ordinates of the two hydrographs at corresponding ends of the time intervals. The resulting curve is the total cell outflow hydrograph. This hydrograph becomes the channel inflow hydrograph of the next cell immediately downstream of the cell considered.

3.4. Routing the channel inflow

The total channel inflow hydrograph for an interior cell is equal to the total cell outflow hydrograph or the sum of such hydrographs for the contributing cells immediately upstream of the cell considered. The channel inflow hydrograph is routed through a pair of elements representing a linear channel and a linear reservoir in series (Fig. 2). The routed hydrograph is then subjected to a channel loss element which subtracts the channel losses from the routed hydrograph (Fig. 3). The remaining part of the routed hydrograph, after subtracting the channel losses, forms the channel outflow component of the interior cell.

Routing is accomplished by a linear reservoir routing subroutine using a storage coefficient K_c $[T]$, followed by a forward time shift, τ_c $[T]$, of the linear reservoir. The values of the two parameters of the routing operation are assumed to be proportional to the length of the channel in the interior cell considered. The values are derived from the corresponding model parameters by the following expressions:

$$K_c = \frac{T_{kf} \cdot L_c(j)}{L_m} \tag{6}$$

and

$$\tau_c = \frac{T_{df} \cdot L_c(j)}{L_m} \tag{7}$$

where $L_c(j)$ and L_m are, respectively, the individual channel length of each cell and the mean channel length for all cells. The storage coefficient and the time shift for a cell with an average channel length are given by the model parameters T_{kf} and T_{df}, respectively.

The outflow hydrograph has a lower peak and a longer time base in comparison with the peak and duration of the inflow hydrograph. In addition, it is shifted forward in time by the time interval τ_c. It should be noted that the time shift does not have to be an integer multiple of the time step used in the computation for rainfall increments. The program includes an interpolation procedure that calculates, by linear interpolation, the ordinates of the outflow hydrograph at the equally spaced time intervals, starting at the time origin used for all computations.

The combined operation of the reservoir routing and the time shifting described above for each of the cells is equivalent to convolution of the channel inflow hydrograph of that cell and a unit impulse response function, $G(t)$, expressed by

$$G(t) = \begin{cases} \dfrac{1}{K_c} \exp[-(t - \tau_c)/K_c] & \text{for } t > \tau_c \\ 0 & \text{for } t \leqslant \tau_c \end{cases} \tag{8}$$

3.5. Subtraction of channel losses

Channel losses are computed individually for each interior cell; they are subtracted from the channel hydrograph after it is routed through the linear channel and linear

reservoir combination. The channel loss operation for each cell is thus assumed to be lumped near the outlet of the cell, after the routing operation and before the channel outflow is combined with the surface outflow to form the total cell output. The channel loss element thus receives the routed channel hydrograph as input and produces as output a channel outflow hydrograph having smaller peak discharge and total outflow volume than the corresponding inflow hydrograph.

The channel loss operation, shown schematically in Fig. 3, consists of subtracting two components from the routed channel hydrograph (Fig. 3(a)). One part is a constant rate of loss for the cell considered (Q_{mc}) and the second is a variable rate (Q_{lt}) which decreases exponentially from an initial value (Q_{lc}) throughout the occurrence of runoff in the routed hydrograph. The part of the routed hydrograph which is in excess of the sum of the two loss components is the output of the channel loss operation.

The constant rate of channel loss for each cell is taken to be proportional to the length of the channel in the cell considered. The value of this cell parameter is computed from the corresponding model parameter (Q_{ca}) by

$$Q_{mc} = \frac{Q_{ca} \cdot L_c(j)}{L_m} \tag{9}$$

where $L_c(j)$ and L_m are, respectively, the individual channel length of each cell and the mean channel length for all cells. The model parameter Q_{ca} represents the constant channel loss for a cell with average channel length.

The variable part of the channel losses is taken to have an initial value, Q_{lc}, which is a fixed proportion (P_{ls}) of the magnitude of the peak flow less the constant channel loss of the routed channel flow hydrograph. The magnitude of this initial loss is given by

$$Q_{lc} = P_{ls} \cdot (Q_{mx} - Q_{mc}) \tag{10}$$

where Q_{mx} is the maximum value of the routed channel hydrograph $[L^3 T^{-1}]$. The model parameter P_{ls} is a dimensionless quantity restricted to be between zero and one ($0.0 \leqslant P_{ls} \leqslant 1.0$).

The diminishing portion of the non-constant channel loss component is obtained by multiplying the variable loss at the beginning of each time interval by a constant multiplier Q_{ks} to produce the value of the variable at the end of the time interval. The value of the multiplier is restricted to the range between zero and one ($0.0 \leqslant Q_{ks} \leqslant 1.0$). The variable loss rate, Q_{lt}, is thus given by

$$Q_{lt}(i) = Q_{ks} \cdot Q_{lt}(i-1) \tag{11}$$

where the initial value of the variable loss rate (for $i = 1$) is computed from the initial value (Q_{lc}), computed above, by

$$Q_{lt}(1) = Q_{ks} \cdot Q_{lc} \tag{12}$$

The total channel losses at the end of each time increment are equal to the sum of the constant loss (Q_{mc}) and the variable channel loss (Q_{lt}) for that time increment. The channel outflow hydrograph for the cell considered (Fig. 3(b)) is the remaining part of the routed channel hydrograph after subtracting the two channel loss components.

4. The study area

The watershed chosen for this study is part of the Walnut Gulch Experimental Watershed, operated by the Agricultural Research Service of the United States Department of Agriculture (USDA-ARS). It is located around the town of Tombstone, about 100 km southeast of Tucson, Arizona. The watershed is representative of the semi-arid rangelands in southeastern Arizona, southwestern New Mexico and northern Sonora, Mexico (Renard, 1970).

Elevations on the watershed range from 1280 to 1830 m above mean sea-level (MSL). The watershed dissects a high foothill alluvial fan with soil texture material ranging from clays and silts to well-cemented boulder conglomerates. The climax vegetation of the Walnut Gulch area is desert plain grassland. About 60% of the watershed supports desert shrubs and 40% is grass covered with a few scattered shrubs.

The annual mean precipitation in this area is about 300 mm. The precipitation distribution is bimodal, with slow-moving cold fronts providing lift for winter precipitation and convective heating of moist tropical air producing summer rainfall. Nearly all the streamflow occurs between July and early October, and results from intense, convective thunderstorms of short duration and limited areal extent. During an average year, 5–10 individual runoff events are recorded at the gaging stations, with the channel being dry about 99% of the time.

The 150 km^2 Walnut Gulch watershed is divided into 12 gaged subdrainage basins of various sizes ranging upward from 2 km^2 to the entire watershed. The watershed is equipped with a network (1.6 km × 1.6 km grid) of 95, 24 h, weighting-type, recording rain gages, which have been in continuous operation since 1956.

The study described herein is limited to two subwatersheds: Watershed 8 (14.70 km^2) and Watershed 11 (6.28 km^2) (Fig. 1). Watershed 8 includes Watershed 11 as part of its area. Watershed 8 is covered by 17 rain-gage stations (10 in Watershed 11 alone). Fig. 1 illustrates 31 model cells which were obtained from topographic maps and aerial photographs for the two watersheds (Karnieli, 1991). The first 14 cells cover the area of Watershed 11.

Approximately 30% of the two watersheds' area is dominated by desert shrubs, with a crown cover of approximately 30% and an understory of grasses with a basal area of less than 1%. Five soil series have been defined by the Soil Conservation Service (SCS) soil survey in 1986; all of them are sandy loam in texture (Karnieli, 1991).

5. Evaluation procedures

5.1. Methods

The aim of the evaluation procedure is to fit the predicted results, as closely as possible, to the observed data and then to analyze the degree of error or goodness of fit. In this study, the evaluation procedure includes the following steps: (1) trial and

error calibration; (2) sensitivity analysis; (3) bilinear interpolation optimization; (4) testing the model on different storm events; (5) testing the model on a different watershed. The evaluation procedure is demonstrated in detail on a single rainfall–runoff event which occurred on 12 August 1971 in Watershed 11. In addition, results of the calibration, test and forecasting operations are presented for 21 events in Watershed 11 and eight events in Watershed 8. These are the largest storm events with reliable data which occurred in the area during 12 years.

CELMOD5 is presented here as a forecasting model, which means that all the model parameters are either constant in the time domain or their temporal variations are known. In its forecasting mode, CELMOD5 requires information about the soil water deficit before each rainfall event. Karnieli and Ben-Asher (1993) described a water accounting model for predicting water excess and soil water content values, based on water balance equation calculations. The water accounting model requires as input only the daily rainfall depths and their interarrival days. This model was optimized for Watershed 11 by using a long record of rainfall and runoff data (468 rainfall events and 100 runoff events during 12 years). As a result, the model allows simulations of the continuous dynamic behavior of the soil water content, which, on desired days, was used as part of the input data to CELMOD5.

5.2. Objective functions

The evaluation procedure is carried out with reference to a specified objective function. The objective function is designed to reduce the differences between the predicted and the observed data during a calibration period to a minimum, or, as an alternative, to calculate the maximization of a likelihood function. To avoid bias, Diskin and Simon (1977) recommended judging the model performance with reference to a number of objective functions rather than to a single function.

Of a large number of objective functions that have been defined in the literature (e.g. Diskin and Simon, 1977), only two are represented below:

(1) predicted to observed total runoff ratio:

$$P_{vat} = \left(\frac{V_{qt} \cdot 100}{V_{at}}\right) - 100 \tag{13}$$

where P_{vat} is the per cent difference of the predicted runoff volume, V_{at}, from the observed runoff volume, V_{qt} (expressed in mm or m^3); P_{vat} can be positive when the observed value is greater than predicted, zero in the case of no deviation, or negative;

(2) predicted to observed peak discharge ratio:

$$P_{ymx} = \left(\frac{Q_{mx} \cdot 100}{Y_{mx}}\right) - 100 \tag{14}$$

where P_{ymx} is the per cent difference of the predicted peak discharge, Y_{mx}, from the observed peak discharge, Q_{mx} (expressed in $m^3\,s^{-1}$).

Table 1
Model parameters for Watershed 11

	Notation	Value	Unit
Routing parameters			
Storage coefficient	A_{kc}	0.27	h
Routing coefficient	T_{kf}	0.10	h
Time delay factor	T_{df}	0.03	h
Infiltration parameters			
Matric potential	SM	Varies	mm
Saturated conductivity	K_s	10.0	$mm\,h^{-1}$
Channel loss parameters			
Constant channel loss	Q_{ca}	0.15	$m^3\,s^{-1}$
Initial loss ratio	P_{ls}	0.10	$0 \leqslant P_{ls} \leqslant 1$
Loss decay factor	Q_{ks}	0.85	$0 \leqslant Q_{ks} \leqslant 1$

5.3. Trial and error calibration

The trial and error calibration procedure is a visual method of adjusting the predicted hydrograph to match or replicate the observed hydrograph. As discussed above, the model has eight operative parameters for the evaluation procedure (Table 1). There are two different types of parameter. The first type includes the physically based parameters — the saturated hydraulic conductivity, K_s, and the constant channel loss parameter, Q_{ca}. The values of these parameters are either directly measured or characterized by physically based functions. It should be noted that, to simplify the evaluation procedure, weighted average values of the parameters K_s and Q_{ca} were used and their spatial variability has not yet been taken into account.

The second type of parameter has no physical meaning, and can be termed a fitting parameter. In this stage of the evaluation, only the effective matric potential, SM, is treated as a fitted parameter. The constant values of all the other parameters are presented in Table 1. The trial and error calibration was limited only to the fitted parameters.

5.4. Sensitivity analysis

The relative change approach is used in the current study for sensitivity analysis. Using this technique, the values of the parameters (which were obtained by the trial and error calibration) are changed by incrementing each of the parameters in turn by a small amount of equal relative magnitude, +5%, +10%, −5% and −10%, while the other parameters are held constant at their original values. The resulting change in the value of the objective function will be a measure of the model sensitivity to changes in the parameter's values. By comparing the sensitivity of the model with respect to the various parameters, one can indicate how the parameters influence the values of the objective functions (Diskin, 1970).

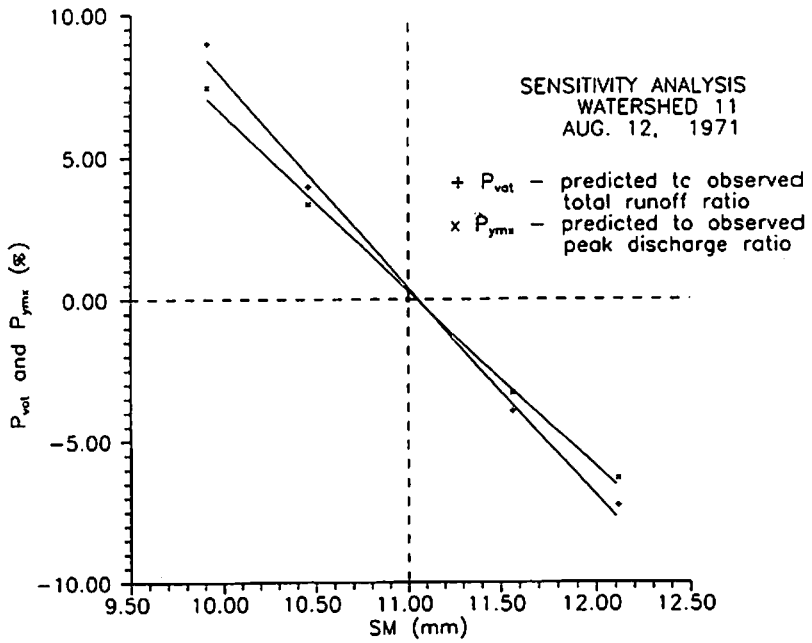


Fig. 4. Sensitivity analysis for the storm of 12 August 1971 on Watershed 11.

Fig. 4 demonstrates the linear relationships obtained from an example of the sensitivity analysis of the effective matric potential parameter, *SM*, for the storm of 12 August 1971 and for the objective functions which describe the goodness of fit of the predicted volume and of the predicted peak discharge. Values from this analysis will be used later for the model optimization.

5.5. Bilinear optimization

Before continuing to the next step (model calibration with optimized parameters), one of the above objective functions should be selected to be used in the optimization scheme. As mentioned above, it is recommended to use more than one objective function, to avoid bias. In the present study, the optimization procedure is based on two objective functions: (1) predicted to observed total runoff ratio (Eq. (13)); (2) predicted to observed peak discharge ratio (Eq. (14)). These objective functions were selected because the flow volume and the peak discharge are widely used for many engineering applications.

Once the objective functions have been selected, it is necessary to minimize simultaneously the differences between the predicted and observed runoff volume and between the predicted and observed peak discharge, to fulfill the criterion of efficiency for the optimization scheme. Combining these two objective functions yields a new function

$$F_{unc} = |P_{vat}| + |P_{ymx}| \tag{15}$$

Table 2
Calibration-optimization procedure; comparison of results for the storm of 12 August 1971 on Watershed 11

	Trial and error calibration	Prediction by optimization	Prediction by API model
Given			
Observed runoff (m^3)	21 579.13	21 579.13	21 579.13
Observed peak discharge ($m^3 s^{-1}$)	9.7850	9.7850	9.7850
Effective matric potential SM (mm)	11.000	9.989	8.376
Results			
Predicted runoff (m^3)	19 895.43	21 543.49	24 071.28
Predicted peak discharge ($m^3 s^{-1}$)	9.2327	9.8669	10.7543
Evaluation			
Predicted to observed runoff ratio, P_{vat} (%)	8.463	0.165	−10.353
Predicted to observed peak discharge, P_{ymx} (%)	5.982	0.830	−9.013
Combined objective function, F_{unc} (%)	14.445	0.995	19.366

The optimal values of the model parameters are those which give the minimum value of the criterion of efficiency defined by Eq. (15).

This efficiency criterion can be fulfilled by using a bilinear interpolation technique. The procedure described below makes use of the linear relationships between the effective matric potential parameter (SM) and the objective functions defined by the ratio between the predicted and observed volume (P_{vat}) and the ratio between predicted and observed peak discharge (P_{ymx}) (Fig. 4).

The following steps are proposed for the optimization procedure, demonstrated with the example of the storm of 12 August 1971:

(1) From the trial and error procedure it was found that $SM = 11.00$ mm is close to fitting the predicted hydrograph to the observed but it is not the optimal value. Results of the trial and error calibration are presented in the first column of Table 2 and in Fig. 5. It can be seen that both the observed runoff volume and the observed peak discharge are larger than the respective predicted values. To obtain a perfectly matched runoff volume, one should increase the predicted value by 8.46%; similarly, to obtain a perfectly matched peak discharge, one should increase the predicted value by 5.98%. The combined objective function defined in Eq. (15) yields

$$F_{unc} = |8.46| + |5.98| = 14.44\% \tag{16}$$

(2) From the sensitivity analysis results for the storm of 12 August 1971 (Fig. 2), it can be seen that there are linear relationships between the percentage of change of P_{vat} and P_{ymx} and the percentage of change of the product SM.

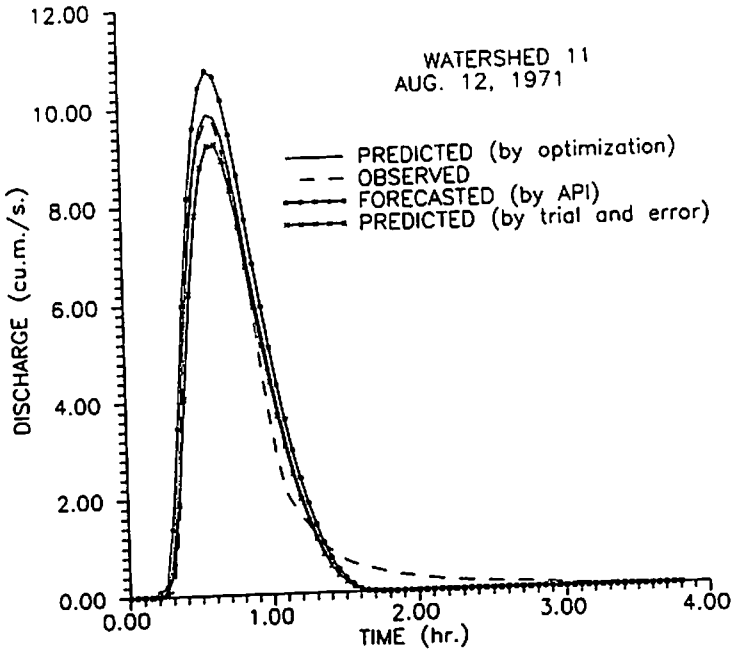


Fig. 5. Hydrographs for the various stages of evaluation, for the storm of 12 August 1971 on Watershed 11.

(3) To obtain optimal results both for the runoff volume and the peak discharge, the bilinear interpolation algorithm was developed (Table 3). This algorithm calculates separately the regression lines for the runoff volume points and for the peak discharge points which are obtained from the sensitivity analysis results. Next, it calculates separately the new SM values for each of these cases, given the percentages of change from Table 2 ($P_{vat} = 8.46$ and $P_{ymx} = 5.98$). Finally, the algorithm calculates the mean SM value from the new SM values.

(4) Running the model again with the optimized value $SM = 9.99$ produces the best-fitted predicted hydrograph, as presented in the second column of Table 2 and in Fig. 5. From the table, it can be seen that the combined objective function gives

$$F_{unc} = |0.16| + |-0.83| = 0.99\% \tag{17}$$

which is a relatively small value. Any further change in SM will introduce a larger combined error for the runoff volume and the peak discharge.

5.6. Forecasting procedure

It should be noted that the calibration–optimization procedure described above helps in refining the predicted hydrograph to match the observed one. However, in the

Table 3
Bilinear algorithm for model optimization; example for the storm of 12 August 1971 on Watershed 11

Step 1: Runoff volume

Change (%)	P_{vat} (%)	SM (mm)
+10	−7.280	12.113
+5	−3.937	11.563
0	0.000	11.000
−5	3.960	10.461
−10	8.994	9.911

Given new P_{vat} value is 8.463%.
Predicted new SM value is 9.911 mm.

Step 2: Peak discharge

Change (%)	P_{ymx} (%)	SM (mm)
+10	−6.327	12.113
+5	−3.285	11.563
0	0.000	11.000
−5	3.345	10.461
−10	7.454	9.911

Given new P_{ymx} value is 5.982%.
Predicted new SM value is 10.087 mm.

Step 3: Mean predicted new SM value is 9.999 mm.

current forecasting version of CELMOD5, an objective procedure for evaluating the soil moisture condition before the storm event is needed. It has been proposed that the residual moisture, which is the only time-dependent parameter in CELMOD5, should be calculated outside the model either by an API-type procedure (Karnieli, 1988) or by soil water balance calculations (Karnieli and Ben-Asher, 1993). Both procedures are based on daily rainfall data before a particular event and the interarrival time between storms. Only the final product of these procedures, the residual soil moisture values for each area represented by a rain gage, is transferred to CELMOD5 and used for the rainfall excess calculations. Consequently, the performance of the API model greatly influences the differences between the observed and predicted hydrographs.

Based on the above models (API or water balance), which eventually produced similar results, it was calculated that the effective matric potential for the storm event of 12 August 1971 in Watershed 11 is $SM = 8.38$. Running CELMOD5 with this parameter value produces the results presented in the third column of Table 2 and in Fig. 5. It can be seen that, according to the objective functions, these are the worst results among the three cases.

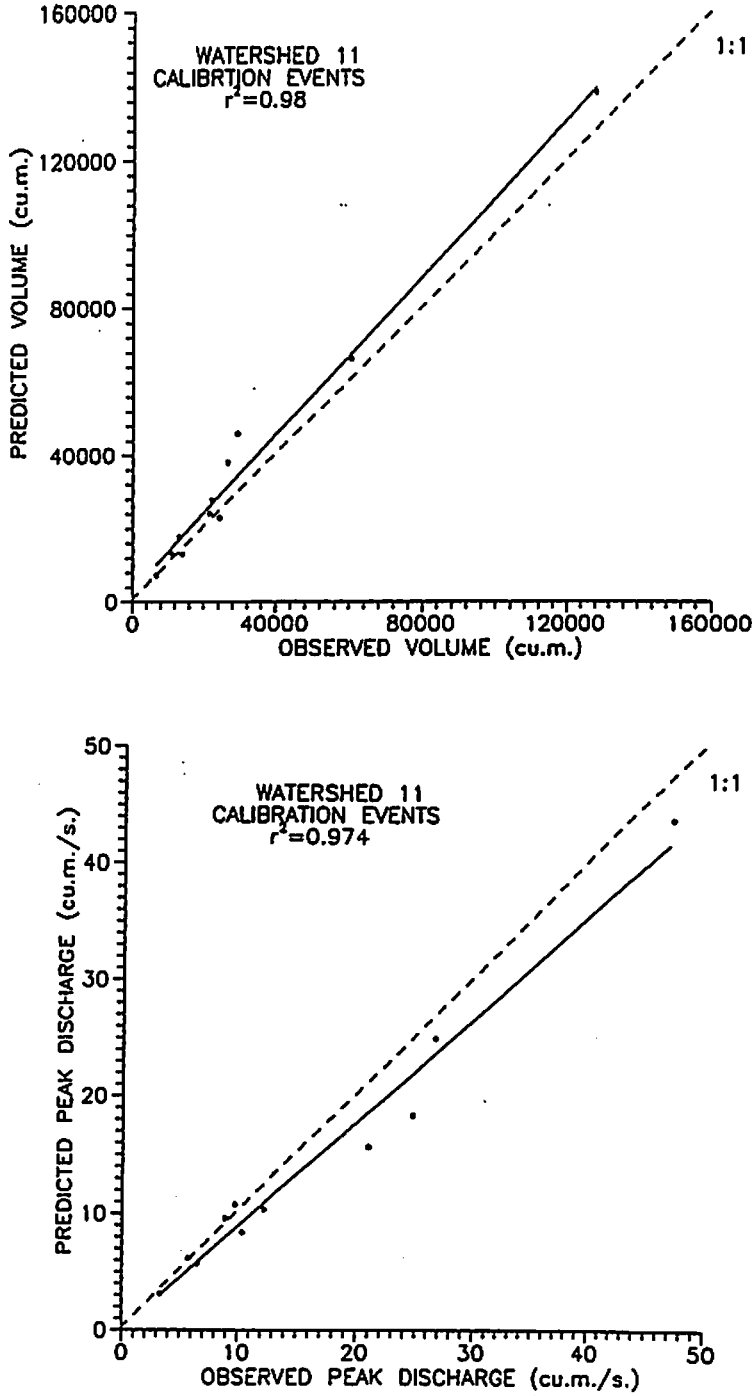


Fig. 6. Observed vs. predicted runoff volume and peak discharge for Watershed 11, calibration events.

5.7. Results

Twenty-one storm event data sets (rainfall and runoff) are available for Watershed 11 between the years 1966 and 1982. These include all the larger events with reliable data. Seven of these storm events contain data for Watershed 8 (as mentioned above, Watershed 11 is a part of Watershed 8).

Eleven of the storm events of Watershed 11 (Events 1–11) were selected for the model calibration. In addition, two types of model testing were used: (1) testing the model on other storm events on the same watershed (Events 12–21); (2) testing the model on another watershed with seven data sets of Watershed 8 (Events 4, 13–15 and 18–20).

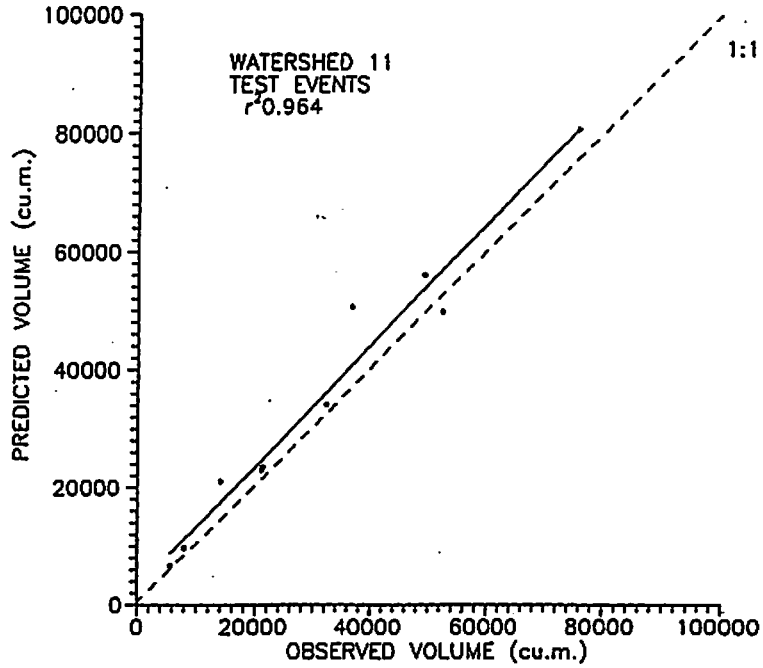
There are three types of results: (1) results of the calibration events; (2) results of the tested events; (3) results of the forecasting procedure. It was found that, according to the main objective function (F_{unc}; Eq. (15)), the mean is 2.23% for all the events. Very high correlations (r^2 above 0.96) are obtained between the observed and predicted volume and between the observed and predicted peak discharges (Figs. 6 and 7) both for the calibration and the test events.

Larger errors were obtained for the forecasting mode than for calibrated and tested modes for all the objective functions. From the correlation plots (Fig. 8), it can be seen that the model produced lower correlations ($r^2 = 0.806$ and $r^2 = 0.771$ for volume and peak discharge, respectively) for Watershed 11 than for Watershed 8. For the forecasting mode of Watershed 8, relatively high correlations were obtained ($r^2 = 0.976$ and $r^2 = 0.956$ for volume and peak discharge, respectively; Fig. 8); however, significant bias was obtained in the volume forecasting case.

6. Discussion and conclusions

CELMOD5 is a semi-distributed, single-event rainfall–runoff model developed for the special hydrologic conditions that prevail in semi-arid locations such as the Southwest in the USA or the Negev in Israel. The model represents the watershed as a set of interconnected cells reflecting the main drainage pattern of the watershed. The model accepts as input non-uniform and time-varying rainfall data for a number of rain gages, derives rainfall excess hyetographs for each cell, and routes the rainfall excess through the cell to produce a surface runoff hydrograph. Any channel inflow to a cell is routed through model elements that modify the shape of the inflow hydrograph and then subtract channel losses. The remaining channel outflow hydrograph is added to the surface runoff hydrograph to produce the total cell outflow hydrograph.

Adopting the assumption that the values of the routing and channel loss parameters are proportional either to the size of the area represented by the cell or to the length of the main stream in this area reduces the number of parameters that are evaluated in the calibration of the model. The rainfall excess separation model is based on the modified Green and Ampt infiltration equation coupled with a simple moisture accounting procedure. This requires the estimation of some additional



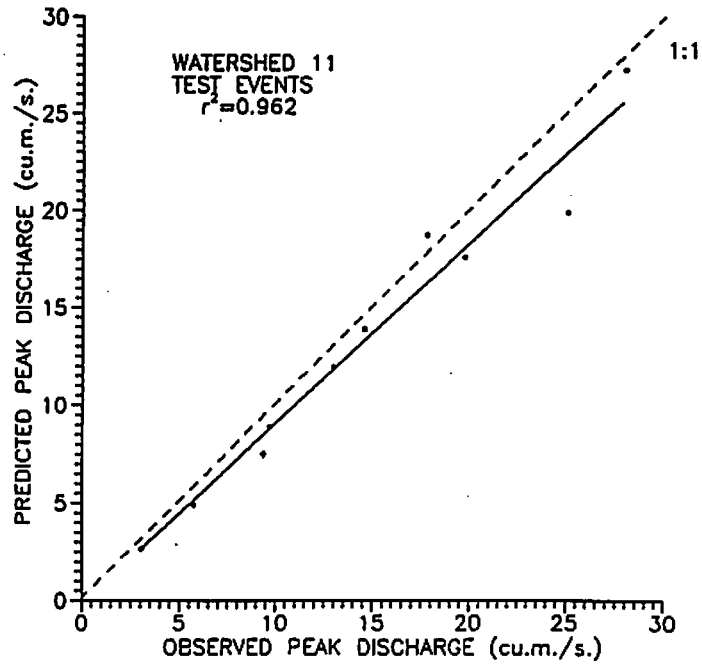
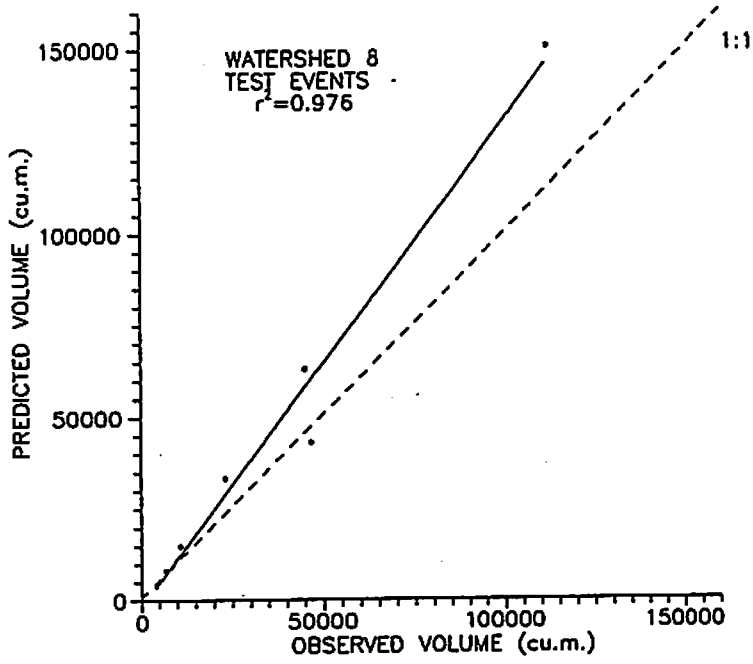


Fig. 7. Observed vs. predicted runoff volume and peak discharge for Watersheds 11 and 8, test events.



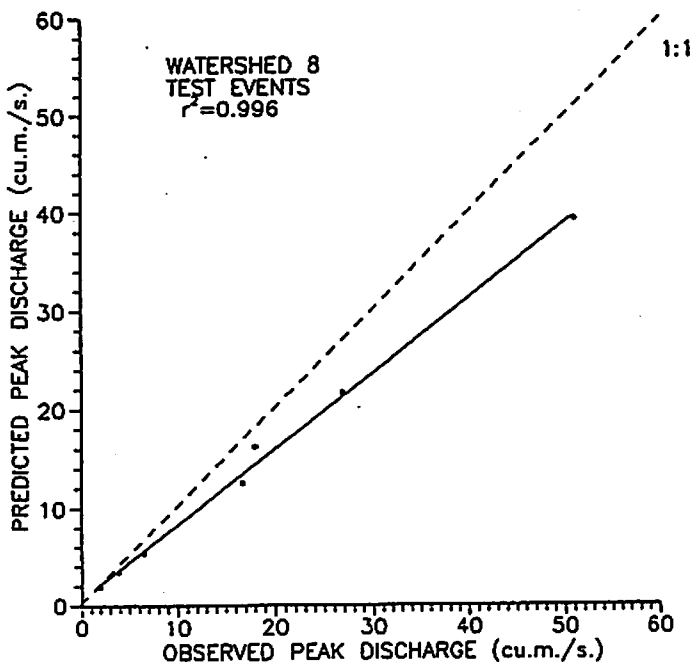


Fig. 7. Continued.

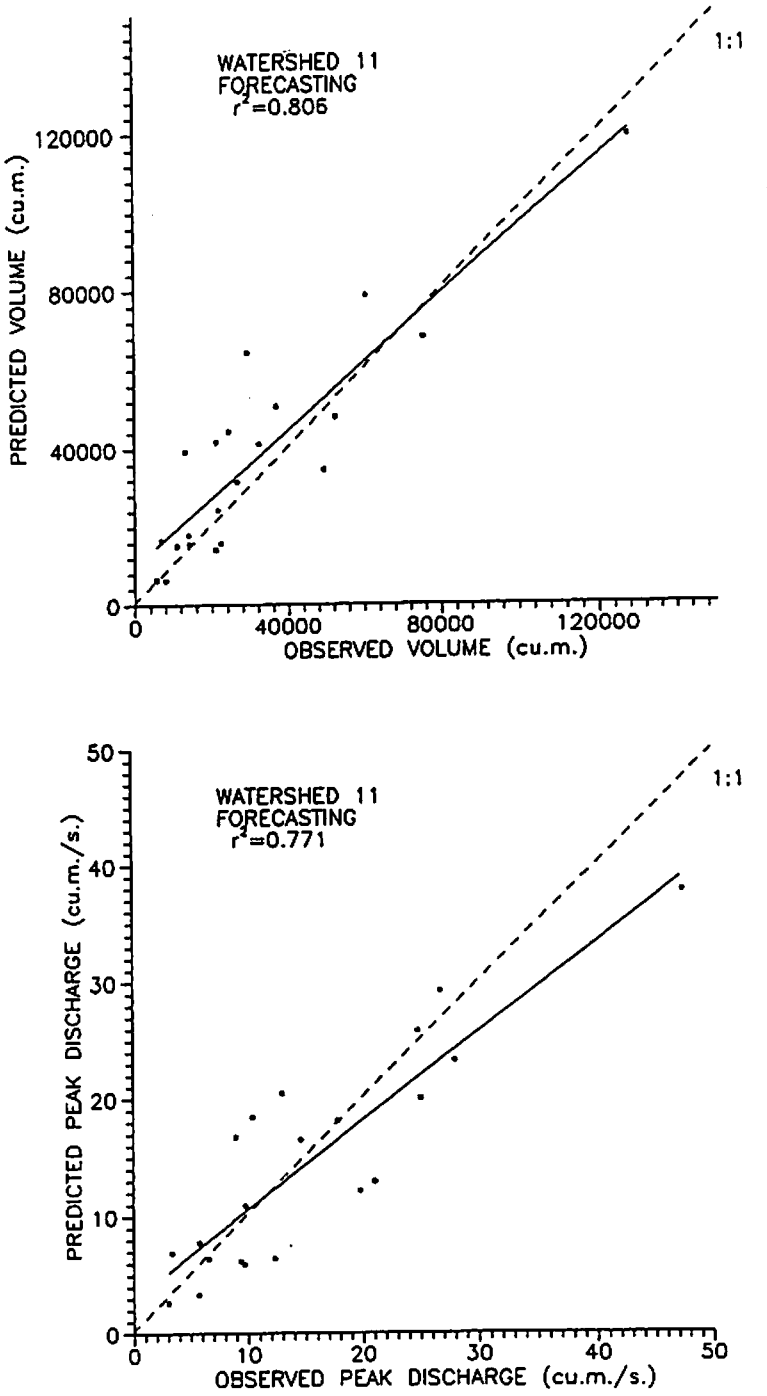


Fig. 8. Observed vs. predicted runoff volume and peak discharge for Watersheds 11 and 8, for forecasting events.

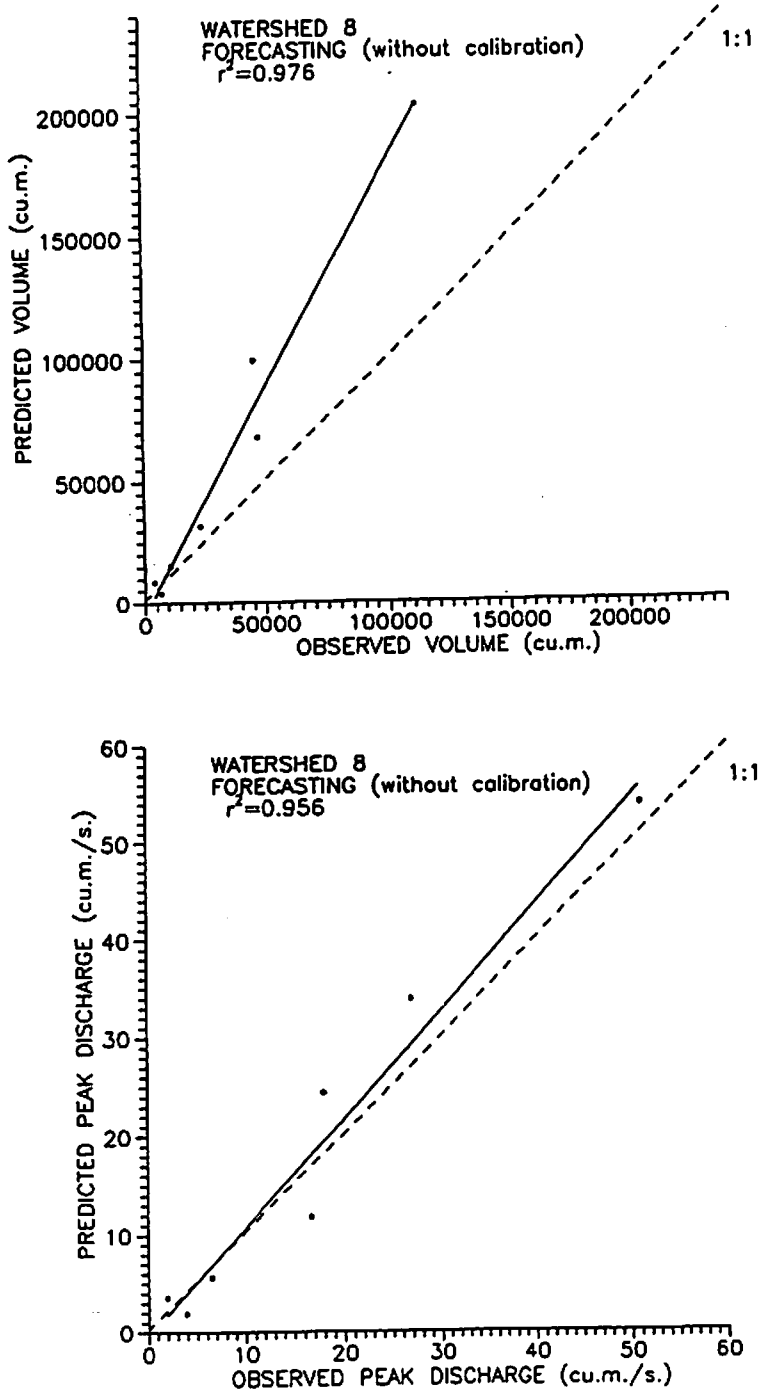


Fig. 8. Continued.

parameters for the entire watershed or for a number of different soil types in the watershed.

This paper presents an approach for calibrating and testing a forecasting storm runoff model. The fitting parameters together with the physically based parameters were used to calibrate the model on one subwatershed in the Walnut Gulch Experimental Watershed (Watershed 11). Because the physically based parameters were deduced from the soil characteristics, they remained constant during the calibration procedure. On the other hand, by changing the fitting parameters' values, a trial and error calibration procedure was used to match as closely as possible the observed and calculated hydrographs.

Two objective functions which are widely used for many engineering applications were selected for the optimization procedure. These two expressed the goodness of fit between the calculated hydrograph volume and peak discharge and the respective values of the observed hydrograph.

By nominating two objective functions rather than a single one, the user is able to utilize the model for more applications. However, greater error is introduced into the results. Taking advantage of the linear relationships between the effective matric potential parameter and the two objective functions obtained from the sensitivity analyses made it possible to develop a bilinear interpolation algorithm to minimize, simultaneously, the differences between the calculated and observed volume, and peak discharge. It was found that this algorithm produces a minimum error with reference to the combined objective function, and any further change in the parameter will introduce a larger combined error. Satisfactory results were achieved from testing the prediction mode of the model with data of other storm events of Watershed 11 and also of another watershed (Watershed 8). In the prediction mode, the effective matric potential parameter was allowed to vary from storm to storm.

In the forecasting mode of the model, the values of the effective matric potential parameter were estimated from the API model or from a water accounting procedure. Larger errors were obtained in the forecasting mode than in the prediction mode. These errors were in part caused by the errors in estimating the soil moisture deficit.

The overestimated runoff volume values in the forecasting mode of Watershed 8 can be explained by higher transmission losses as a result of larger and deeper channels. These errors highlight some of the weaknesses of the model, which was tested with one rainfall pattern (thunderstorms) on watersheds with ephemeral stream channels. It can be concluded that, when applying the model to other watersheds, it should be evaluated for the new conditions and the fitting parameters should be a basis for a new calibration procedure.

In summary, the main contribution of the model is the technique which provides significant assistance in terms of design runoff forecasting. The proposed forecasting procedure can be used for various applications in the planning, design and management of water resources and other engineering systems. The design forecasting concept is 'one-step-ahead' prediction, and it is believed that by applying the model with weather radar data or a telemetric rain-gage system it can be used in real time for flood forecasting and warning.

7. Acknowledgments

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