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SPUR— Simulation of Production and Utilization of Rangelands:

A Rangeland Model for Management and Research

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SPUR HYDROLOGY COMPONENT: UPLAND PHASES

By K. G. Renard, E. D. Shirley, J. R. Williams, and A. D. Nicks[1]

INTRODUCTION

The hydrology component of the model is designed to use inputs from the climate component and produce outputs unto its own (for example, runoff and sediment yield) or inputs for other components of the model (for example, estimates of available soil moisture for forage production and water for grazing animals). The hydrology component is divided into three parts: an upland phase, a snowmelt phase, and a channel phase. This portion of the report is for the upland phase.

In the streams draining the rangeland areas of the western United States, extreme spatial and temporal variability in physiographic and climatic conditions require that a hydrologic model consider such conditions. For example, it is very possible to have an individual storm event occurring as rain at low elevations and snow at high elevations. Airmass thunderstorms dominating the rainfall-runoff process in the semiarid Southwest have extreme variations in precipitaton depth in short distances (1 in/mi is not rare).

A hydrologic model component should be capable of simulating the effects of management changes on streamflow for streams that may have influent or effluent characteristics; have flow conditions that are subcritical or supercritical; and have a wide variety of slopes up to steep, rocky, pool-riffle systems.

The objectives of the upland phase of the hydrologic model are: 1) be capable of predicting changes in water quantity and quality resulting from management changes; 2) be physically based, so that model parameters can be evaluated from available data for ungaged areas; 3) have sufficient detail to allow simulation on subdivided watersheds to coincide more or less with ranch and pasture boundaries; 4) be computationally efficient to enable long-term simulation for frequency analyses; 5) be capable of providing input to other SPUR model components, such as soil moisture for plant forage yield estimates and water for domestic animals and wildlife; and 6) be used for environmental impact analysis, nonpoint pollution assessment, and other types of resource utilization and environmental protection problem solutions.

Although these objectives may seem overly ambitious, there have been significant improvements of water resource models in recent years (Crawford and Donigian 1976;[2] Williams and LaSeur 1976; Beasley et al. 1977; Simons et al. 1977; Knisel 1980a, 1980b), which facilitate such a development.

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[2]The author's name followed by the year underlined, refers to Literature Cited, p. 38.

The upland phases of the hydrology model for SPUR draw heavily from a model called SWRRB (Simulation for Water Resources in Rural Basins),[3] which has been modified and improved to consider the essential features known to affect the hydrologic response from rangelands. The SWRRB model includes the major processes of surface runoff, percolation, return flow, evapotranspiration, pond and reservoir storage, and erosion and sedimentation. The well known curve number technique (USDA 1972) is used to predict surface runoff for any given precipitation event because 1) many years of use have given confidence in its validity, 2) it relates runoff, soil type, vegetation, land use, and management, and 3) it is computationally efficient. The use of rainfall data for short time increments (minutes and/or hours), which are required with infiltration equations to compute precipitation excess, are not generally available for most areas of the United States, and especially not on the rangelands with the orographic precipitation effects, sparsity of recording rain gages, etc. Finally, daily rainfall estimates are computationally more efficient than similar operations with shorter time increments.

MODEL DESCRIPTION

Water Balance

The SPUR model maintains a continuous water balance on a daily computational basis using the equation

$$SW = SW_o + P - Q - ET - PL - QR \qquad [1]$$

where

SW = current soil water content (in),

SW_o = initial soil water content (in),

P = cumulative rainfall (in),

Q = cumulative amount of surface runoff (in),

ET = cumulative amount of evapotranspiration (in),

PL = cumulative amount of percolation loss to ground water storage (in), and

QR = cumulative amount of return flow (in).

In maintaining the continuous water balance, complex watersheds are subdivided to reflect different vegetation or soils, topography, stream morphology, etc. In other words, runoff is computed for each subarea and the water routed to the outlet of the basin to obtain the total runoff. This accounting allows changing management practices of only a portion of the area and should improve the model's accuracy, yet provide a more detailed physical preservation of the watershed details.

Soil Layer Water Storage

The soil in each subarea of the watershed is divided into layers (user specified number of layers (up to eight) and layer thickness for each subarea). Water balance

[3] Mimeograph handout by J. R. Williams and A. D. Nicks.

is done on a daily basis using rainfall excess, evapotranspiration, percolation, and return flow as described in equation [1]. Total storage, field capacity, and initial water storage in the various layers are computed from input parameters as follows:

$$Ul_i = (SMO_i - SM15_i)THK_i \quad [2]$$

$$FC_i = (SM3_i - SM15_i)THK_i \quad [3]$$

$$SW_{oi} = (FC_i)(STF) \quad [4]$$

where

Ul_i = upper limit of water storage in layer i (in),

FC_i = field capacity in layer i (in),

SW_{oi} = initial soil water in layer i (in),

SMO_i = soil porosity for layer i (in/in),

$SM3_i$ = 0.3 bar water content for layer i (in/in),

$SM15_i$ = 15 bar water content for layer i (in/in),

THK_i = soil layer thickness for layer i (in), and

STF = initial soil water content as a fraction of field capacity.

Runoff

The traditional three antecedent moisture levels (I -dry, II - normal, III - wet), as used by SCS, have been modified in the model by allowing soil moisture to be updated daily and by computing daily curve numbers based on soil water storage rather than using the three curve numbers associated with their moisture classes. Thus, each day has a curve number (Williams and LaSeur 1976), and the soil moisture changes between runoff events with estimates of evapotranspiration and percolation, using routines very similar to those used in CREAMS (Knisel 1980b). From the curve number method, surface runoff is estimated on a daily basis from

$$Q = \frac{(P - I_a)^2}{P + s - I_a} = \frac{(P - 0.2s)^2}{P + 0.8s} \quad [5]$$

where

Q = daily runoff (in),

P = daily rainfall (in),

s = a retention parameter (in), and

I_a = 0.2s = initial abstraction.

The maximum value, s_{mx}, for the retention parameter, s, is computed with the following SCS curve number relationship (USDA 1972):

$$s_{mx} = \left(\frac{1000}{CN_I} - 10\right) \qquad [6]$$

where CN_I is the dry antecedent moisture condition curve number. If handbook curve numbers are available for the normal moisture condition, CN_{II}, the following polynomial may be used to estimate CN_I.

$$CN_I = -16.91 + 1.348(CN_{II}) - 0.01379(CN_{II})^2 + 0.0001177(CN_{II})^3 \qquad [7]$$

The soil retention parameter is computed daily as a weighted average of the unused storage in the various soil layers scale from 0 to s_{mx}.

$$s = s_{mx}\left[\sum_{i=1}^{N} W_i\left(\frac{Ul_i - SW_i}{Ul_i}\right)\right] \qquad [8]$$

where

N = number of soil layers,

SW_i = current water storage in layer i (updated daily) (in), and

W_i = weighting factor.

The weighting factors decrease exponentially to give greater dependence of s on the upper soil layers.

$$W_i = a\exp(-4.16d_i) \qquad [9]$$

where

d_i = (depth to bottom of layer i)/(depth to bottom of last layer), and

a = constant adjusted so $\sum_{i=1}^{N} W_i = 1$.

Peak Flow Calculation

Peak discharge for daily runoff events is calculated using some relationships discussed in the channel routing process (SPUR Hydrology Component: Water Routing and Sedimentation) for SPUR

$$Q_p = C_5 Q/D \qquad [10]$$

where

Q_p = peak flow rate (in/hr),

Q = daily runoff volume (in),

D = duration of runoff (hr), and

C_5 = a constant.

Runoff duration (D is in hr) is obtained from

$$D = C_1 A^{C_2} \text{ (hr)} \qquad [11]$$

where

A = watershed area (acres); and C_1 and C_2 are constants.

Combining equations and converting units gives

$$Q_p = (1.00833) \frac{C_5}{C_1} QA^{(1-C_2)} \qquad [12]$$

where the constant (1.00833) allows conversion to give Q_p in cubic feet per second. The constants C_1, C_2, and C_5 are data input to the program.

Percolation

The percolation component of SPUR uses a storage routing model combined with a crack-flow model to predict flow through the root zone. These models are similar to those used in CREAMS (Knisel 1980b) and SWRRB. Water moving below the root zone becomes ground water, or appears as return flow that is routed into the channel network.

In the following, $PL1_i$ is percolation flow out of the bottom of layer i from the storage routing model. $PL2_i$ is the crack flow out of the same layer. $PL_i = PL1_i + PL2_i$ is the total flow out of layer i (ignoring return flow). PL_o is computed as equal to precipitation minus rainfall excess--the amount of water flowing into the first layer.

Flow through a soil layer may be restricted by a lower layer which is saturated or nearly saturated. PL_i, as subsequently computed, may exceed the projected available storage in the next layer ($UL_{i+1} - SW_{i+1}$ + projected evapotranspiration losses from layer i + 1), in which case, PL_i is set to this projected value. There is no "succeeding" layer to the bottom layer. Crack-flow computations use bottom layer values where the bottom layer needs succeeding layer values. PL2 is not limited by the succeeding layer.

Storage Routing

The storage routing model uses an exponential function with the percolation computed by subtracting the soil water in excess of field capacity at the end of the day from that at the beginning of the day

$$PL1_i = \begin{cases} (SW_i - FC_i)\left(1 - \exp(-\Delta t/T_i)\right), & SW_i > FC_i \\ 0, & SW_i \leq FC_i \end{cases} \qquad [13]$$

where

$PL1_i$ = amount of percolate (in),

SW_i = the soil water content at the beginning of the day for layer i (in),

Δt = time interval (24 hr),

T_i = travel time through a particular layer (hr),

FC_i = the field capacity water content for layer i, (in), and

i = soil layer number increasing with depth.

The travel time through each soil layer is computed with the linear storage equation

$$T_i = \frac{SW_i - FC_i}{H_i} \qquad [14]$$

where

H_i = the hydraulic conductivity of layer (in/hr).

Hydraulic conductivity is varied from the specified saturated conductivity value by

$$H_i = SC_i \left(\frac{SW_i}{UL_i}\right)^{\beta_i} \qquad [15]$$

where

SC_i = saturated conductivity for layer i (in/hr), and

β_i = parameter that causes $H_i \rightarrow 0.0022$ SC as $SW_i \rightarrow FC_i$.

The equation for estimating β_i is

$$\beta_i = \frac{-2.655}{\log(FC_i/UL_i)} \qquad [16]$$

where the constant (-2.655) assures that $H_i = 0.0022\ SC_i$ at field capacity.

Crack Flow

The crack-flow routine is used in the model to allow percolation of infiltrated precipitation even though the soil water content may be less than field capacity. Given a dry soil with cracks, infiltration can move through the cracks of a layer without becoming part of the soil water in the layer, while the portion that becomes part of a layer's stored water cannot percolate by the storage routing model until the storage exceeds field capacity. Crack flow percolation uses the equation

$$PL2_i = (d_c)(PL_{i-1})\left(1 - \frac{SW_{i+1}}{UL_{i+1}}\right)^2 \qquad [17]$$

where d_c is a soil parameter that expresses degree of cracking. Crack flow occurs only on days when water enters the layer (PL_{i-1}) and is greatest when the next layer down is dry.

Since the daily time increment is relatively long for routing the flow through soils, it is desirable to route the water in volume increments. The increments to be routed are variable and are a function of the difference between the $UL_i - FC_i$ and the total amount to be routed. By dividing the layer inflow into several "slugs," each slug may be routed through the layer, thus allowing SW_i to be updated during the calculation.

Return Flow

Return flow is calculated as coming from the bottom soil layer, N. The return flow function used for SWRRB is also used in SPUR (note the similarity to eq. [13])

$$QR = \left(SW_N - FC_N\right)\left(1 - \exp(1/T_R)\right) \qquad [18]$$

where

QR = return flow (in),

T_R = return flow travel time (days), and

N = last soil layer.

Return flow time, T_R, is the time required for subsurface flow from the centroid of the basin to the basin outlet. The value of T_R is input for each subarea by the SPUR user instead of being calculated form soil hydraulic properties. Experienced hydrologists familiar with the base flow characteristics of watersheds within a region should have little problem in assigning reasonable values to T_R.

Evapotranspiration

The evapotranspiration (ET) component in SPUR is the same as is used in CREAMS and SWRRB, and is based on work by Ritchie (1972). Potential evaporation is computed with the equation

$$E_o = \frac{0.0504\ \Delta\ H_o}{\Delta + \gamma} \qquad [19]$$

where

E_o = potential evaporation (in),

Δ = slope of the saturation vapor pressure curve at the mean air temperature,

H_o = net solar radiation (Langleys), and

γ = a psychometric constant.

is computed with the equation

$$\Delta = \frac{5304}{(T_K)^2} \exp(21.255 - 5304/T_k) \qquad [20]$$

where

T_k = daily temperature (degrees Kelvin).

H_o is calculated with the equation

$$H_o = \frac{(1 - \lambda)\ R}{58.3} \qquad [21]$$

where

R = daily solar radiation (Langleys), and

λ = albedo.

Soil Evaporation

The model computes soil evaporation and plant transpiration separately. Potential soil evaporation is computed with the equation

$$E_{so} = \min \begin{cases} E_o \exp(-0.4\ LAI) \\ E_o\ GR \end{cases} \qquad [22]$$

where

E_{so} = potential evaporation at the soil surface (in),

LAI = leaf area index defined as the area of plant leaves relative to the soil surface, and

GR = mulch (residue) cover factor (in).

Actual soil evaporation (E_s) is computed in two stages based on the soil moisture status in the upper soil profile. In stage 1, soil evaporation is limited only by the energy available at the surface, and thus, is equal to the potential (equation [22]). When the accumulated soil evaporation exceeds the first stage upper limit, the stage 2 evaporation begins. (The reader is referred to Ritchie (1972) for additional explanation of the procedure). The first stage upper limit is estimated from

$$U = 1.38\ (\alpha - .118)^{.42} \qquad [23]$$

where

U = stage 1 upper limit (in), and

α = soil evaporation parameter dependent on soil water transmission characteristics (ranges from 0.13 to 0.22 in/day$^{1/2}$).

Ritchie (1972) suggests using α = 0.14 for clay soils, 0.18 for loamy soils, and 0.13 for sandy soils. Similar values were obtained for data from Jackson et al. (1976).

Stage 2 soil evaporation is predicted by

$$E_s = \alpha \left(t^{1/2} - (t-1)^{1/2}\right) \qquad [24]$$

where

E_s = soil evaporation for day t (in), and

t = days since stage 2 evaporation began.

Plant Transpiration

Potential transpiration (E_{po}) from plants is computed with the equations

$$E_{po} = \frac{(E_o)(LAI)}{3}, \qquad 0 \leq LAI \leq 3 \qquad [25]$$

$$Epo = E_o - E_s, \qquad LAI > 3 \qquad [26]$$

(NOTE: If $E_{po} + E_s > E_o$, E_s is reduced so $E_{po} + E_s = E_o$.)

Because the LAI is generally considerably less than 3 in rangeland plant communities, such as SPUR is intended to consider, equation [25] will be used most of the time. If soil water is limited, plant transpiration is reduced with the equation

$$E_p = \frac{(E_{po})(SW)}{0.25(UL)}, \qquad SW \leq 0.25(UL) \qquad [27]$$

(NOTE: If $E_p + E_s$ exceeds available water, E_s is reduced so $E_p + E_s$ = available water.)

where

E_p = plant transpiration reduced by limited soil moisture (in),

SW = current soil water in the root zone (in), and

UL = total soil water storage capacity (in).

Evapotranspiration (ET) then is the sum of plant transpiration (equation [25], [26], or [27]) plus soil evaporation (equation [23] and/or [24]), and cannot exceed available soil water.

Distribution of ET in Soil Profile

Given the computed ET for a particular day, it must be distributed properly in the soil layers based on the rooting depth. The rate of soil water use by evapotranspiration as a function of root depth is computed with the equation

$$v = v_o \exp(-v_1 D) \qquad [28]$$

where

v = water use rate by crop at depth D (in/day),

v_o = water use rate at the surface (in/day).

v_1 = 3.065, and

D = Soil depth/depth to bottom of last soil layer with roots.

The total water use within any depth can be computed by integrating equation [28]. The value of v_o is determined for the root depth each day, and the water use in each soil layer is computed with the equation

$$v_o = (v_1)\ (ET)/(1.0 - \exp(-v_1)) \qquad [29]$$

$$UW_i = \frac{v_o}{v_1} \left(\exp(-v_1\ D_{i-1}) - \exp(-v_1\ D_i)\right) \qquad [30]$$

where

UW_i = water use in layer i (in), and

D_{i-1} and D_i = the fractional depths at the top and bottom of layer i.

(NOTE: The UW_i are the initial estimates of ET to subtract from the various soil layers. If a layer has insufficient water, the excess ET is taken out of the first layer containing available water and having roots present.)

Water Balance for Ponds

Water for grazing animals in rangeland watersheds is often supplied by small earth dams, which create small ponds. These ponds can hold a considerable portion of the runoff from the contributing watershed, depending upon how full the pond is when runoff begins. In addition, the retention of water in such ponds can result in a significant delay or reduction in the downstream runoff and a distortion of the time-flow rate relationship. The SPUR model uses a component of SWRRB which was designed to account for the effects of farm/ranch ponds on water yield. The water balance equation is

$$VM = VM_o + QI - QO - EV - SP \qquad [31]$$

where

VM = volume of water stored in pond at end of day (ac-ft),

VM_o = volume of water in pond at beginning of day (ac-ft),

QI = inflow to the pond during the day (ac-ft),

QO = outflow from the pond during the day (ac-ft),

EV = evaporation from pond (ac-ft), and

SP = seepage from pond (ac-ft).

(Note: The amount of water consumed by grazing animals is assumed to be negligible compared to seepage and evaporation losses.)

Inflow, QI, is considered to be surface runoff from the watershed area draining into the pond plus precipitation on the pond's water surface. Outflow from the pond occurs from either an emergency spillway or a principle spillway and occurs when the permanent pool storage capacity is exceeded.

Evaporation from the pond is computed with the equation

$$EV = \frac{1}{12} (\alpha) (E_o) (SA) \qquad [32]$$

where

α = evaporation coefficient (≈ 0.6), and

SA = surface area of the pond (acres).

Seepage from the pond is computed with the equation

$$SP = 2(SC)\ (SA) \qquad [33]$$

where

SC = saturated conductivity of the pond bottom (in/hr).

No effort was made to make SC vary with water depth in the pond and other factors, like soil stratification or sediment distribution, in the pond. These modifications were not felt to be warranted because of the need for additional detailed information to implement them.

Since pond surface area is required for computing evaporation (equation [32]) and seepage (equation [33]), a relationship between pond volume and surface area is necessary. Data from a large number of stock ponds and small reservoirs in Texas and Oklahoma (USDA 1957) indicate that surface area can be calculated with the equation

$$SA = SA_{max}\ (VM/VM_{max})^{\delta} \qquad [34]$$

where

δ = a parameter determined to be 0.9,

VM_{max} = maximum pond volume (ac-ft), and

SA_{max} = maximum pond surface area (ac).

Other research by Hanson et al. (1975) has indicated that, in Montana and South Dakota, the exponent δ should be about 0.7.

Sediment Yield

Estimating soil loss from the upland areas of rangelands is a difficult problem (Renard 1980) because most of the technology currently in use was developed for cultivated cropland areas. The Universal Soil Loss Equation (USLE) (Wischmeier and Smith 1978) and the modification to this equation (MUSLE) (Williams and Berndt 1977) are used in SPUR. The equation used is

$$Y = \eta(Q \bullet Q_p)^{0.56}\ (K)\ (C)\ (P)\ (LS) \qquad [35]$$

where

Y = sediment yield from upland area (tons/ac),

η = coefficient = 95

Q = upland runoff volume (in),

Q_p = peak flow rate (cfs),

K = soil erodibility factor,

C = cover/management factor,

P = erosion control practice factor, and

LS = slope length and steepness factor.

The determination of the LS factor in this equation is quite critical to the sediment yield calculation. Care must be taken when selecting the model elements to describe prototype configuration. As the model is used to describe larger and larger elements, some detail is lost. Thus, the way the LS term is evaluated may change with the size of the area to be simulated. The average land slope of any subarea or subwatershed can be estimated by field measurements or by measurements from a topographic map with the Grid-Contour Method (Williams and Berndt 1976) using the equations

$$S_d = N_d H/D_d \qquad [36]$$

$$S = (S_\ell^2 + S_w^2)^{1/2} \qquad [37]$$

where

S_d = slope in one grid direction,

S = average land slope of a subarea or subwatershed,

N_d = total number of contour crossings from all grid lines in direction d,

H = contour interval,

D_d = total length of all grid lines within the subarea in direction d,

S_ℓ = slope in the length grid direction obtained from equation [36] and,

S_w = slope in the width direction obtained from equation [36].

The average slope length can be estimated for each subarea or subwatershed by field measurements or with the Contour-Extreme Point Method (Williams and Berndt 1976) by using the equation

$$L = \frac{LC}{2EP} \qquad [38]$$

where

EP = number of extreme points (channel crossings) on the contours of a topographic map,

LC = total length of all contours within the subarea or subwatershed, and,

L = average slope length (ft).

The LS factor is computed with the equation

$$LS = \left(\frac{L}{72.6}\right)^M (65.41 \sin^2 \theta + 4.56 \sin \theta + 0.065) \qquad [39]$$

where

θ = angle of slope (note S is often substituted for sin θ), and

M = exponent proportional to steepness.

The exponent, M, varies with slope and is computed with the equation

$$M = 0.6\left(1-\exp\left(-35.835(S)\right)\right) . \qquad [40]$$

The value of the C factor for each crop is determined from the tables in Agriculture Handbook 537 (Wischmeier and Smith 1978). In many rangeland areas, erosion pavement (rocks larger than one-half in) on the surface are very effective in absorbing the kinetic energy of rainfall. We recommend including an estimate of the percentage of the soil surface covered by the erosion pavement and including it with the plant basal area to arrive at a C factor (for example, by using table 10 of Handbook 537). Values of K and P can be obtained for each subwatershed using Handbook 537 or using the conservation reports of SCS for each State.

Sediment Routing in Ponds

The SPUR model assumes that the sediment coming into the pond with the inflow is retained there. Thus, the outflow from the pond is assumed to be clear, and any water leaving the pond thus picks up sediment again from the channel boundaries below the pond.

APPLICATION OF SPUR UPLAND HYDROLOGY MODEL

The hydrology portion of the SPUR model is designed to operate with the climatic portion of the SPUR model providing the input and with the channel routing portions for both the runoff and sediment transport. Thus, the user of the technology must be familiar with considerations in this part of the program as well.

Figure 1 is a flow chart of the upland hydrology model in SPUR. Examination of the chart shows that the main program consists of a series of loops to handle individual computations as well as those for each month and year. Loops are also used to handle the channel routing calculations. Finally, depending upon the needs of the user, summaries of the calculations can be made on a daily or monthly (tables for year) basis.

The conceptual configuration of a surface topography for input to the model is given in figure 2. In this conceptualization, there were 4 channel reaches (C1. . . .C4), 11 lateral inputs (L1, L2. . . .L11), 2 upland regions (U1 and U2), plus 1 pond (P1). The constraints shown at the bottom of the figure illustrate requirements for the computer model. These constraints allow simulation of almost any topographic or land use variation patterns into a fairly rigorous reproduction of the prototype.

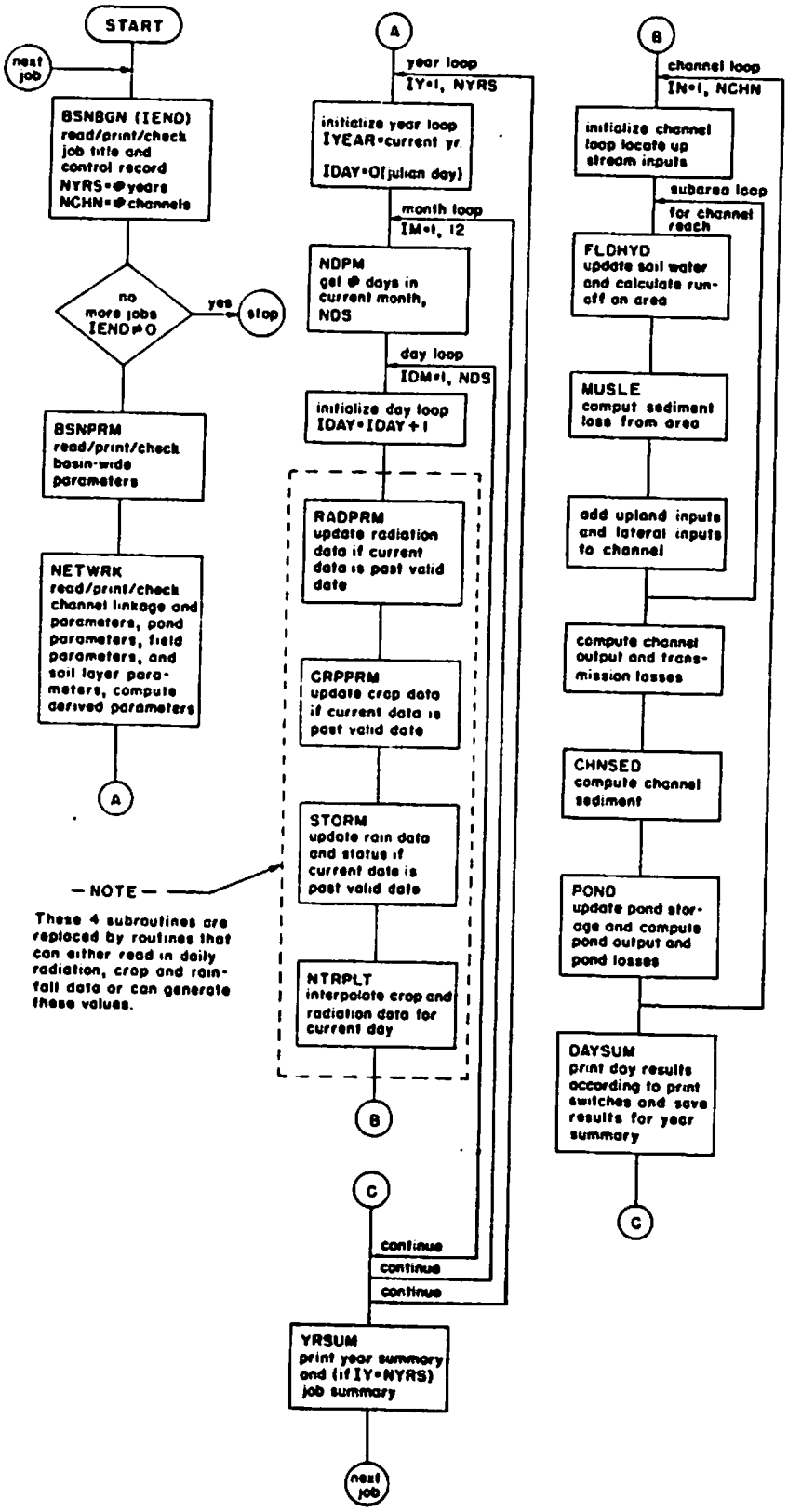
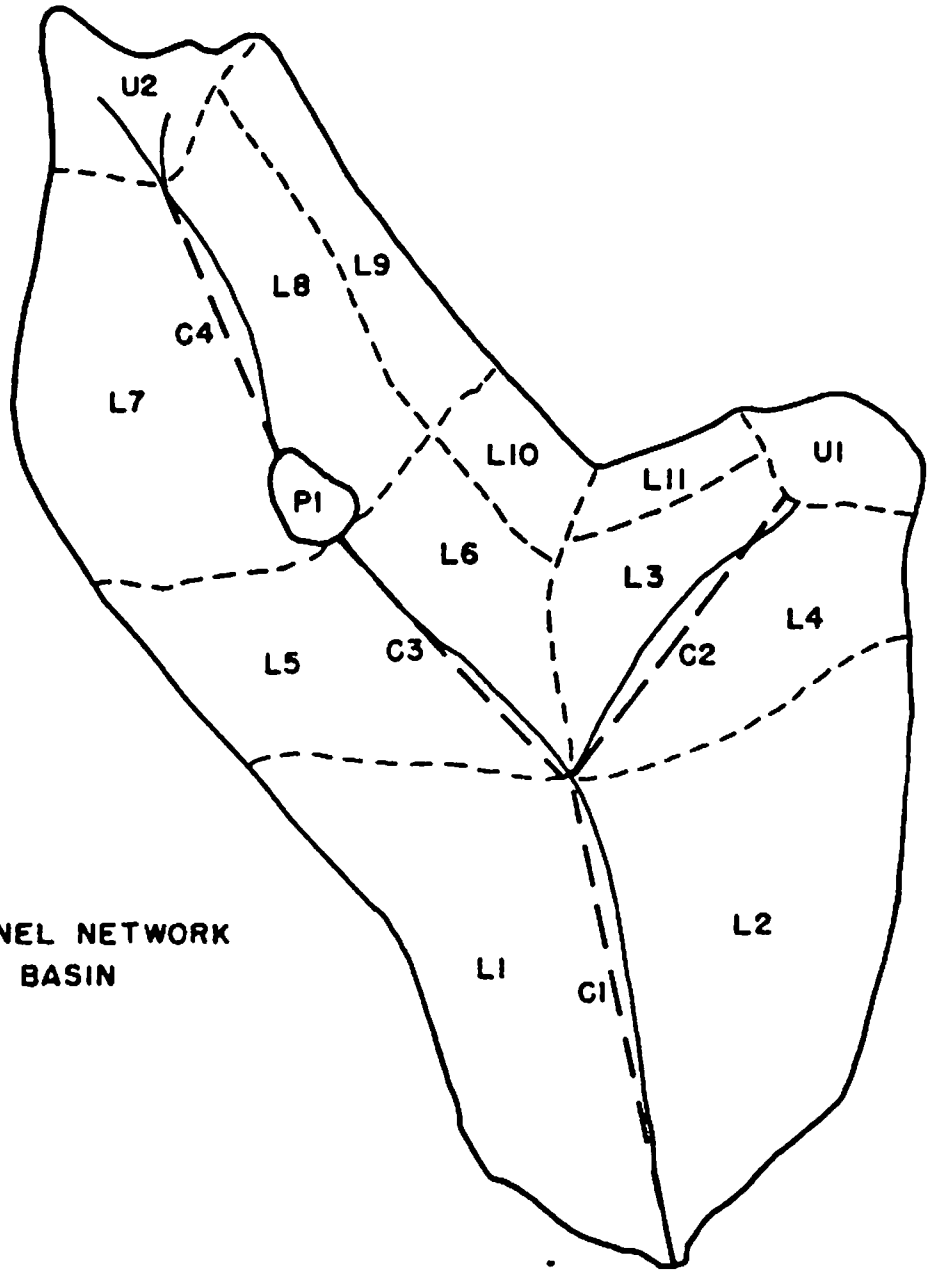


Figure 1.--Hydrology component. This flowchart of the upland and channel phases includes everything but the snowmelt.



SAMPLE CHANNEL NETWORK
FOR A BASIN

MODEL CONSTRAINTS

1. EACH CHANNEL MUST HAVE AN INPUT ; EITHER AN UPLAND REGION OR UP TO TWO CHANNELS.
2. EACH CHANNEL MUST HAVE ONE OR MORE LATERAL INPUTS.
3. EACH CHANNEL MAY OUTPUT THROUGH A POND.

Figure 2.—Conceptualization of a watershed into upland ares (U1-U2), lateral areas (L1-L11), stream channel reaches (C1-C4), and ponds (P1).

Illustrations of the model application to a small watershed on Walnut Gulch follow. Walnut Gulch is an ephemeral tributary of the San Pedro River in southeastern Arizona. The watershed is an intermountain alluvial basin typical of mixed grass-brush areas encountered in Major Land Resource Area 41, the Southeastern Arizona Basin and Range. Figure 3 illustrates the features of stock pond watershed 23 (known locally as the Lucky Hills Watersheds) on Walnut Gulch. The watershed was conceptualized for the model as one 9.1-acre upland area discharging to a 4,000-ft long channel (C1 and C2) having lateral contributing areas L1 (49.2 acres) and L2 (49.7 acres) or a total drainage of 108.0 acres into the pond (P1).

Tables 1, 2, and 3 contain the input data used in the upland hydrology portion of the SPUR model for the 108-acre watershed used in the test application for the hydrology component only. The 100-day return flow travel time was used to ensure that there was no base flow. Similarly, the use of 0 for the crack-flow factor means that the model in the test application did not consider this type of flow situation (table 1).

The soils data in Table 2 are for a Rillito-Laveen gravelly loam soil. Gelderman (1970) described this association as occurring on moderately sloping ridges formed by the deep dissection of old alluvial fans and valley plains.

Table 1.--Parametric values input for upland areas in the SPUR hydrology model

Parameter	Units	Field identification		
		1	2	3
Field type		Upland	Lateral	Lateral
Soil layers	number	8	8	8
Field area	acres	9.1	49.2	49.7
Curve number		86	86	86
Return flow time	days	100	100	100
MUSLE Parameters				
K		0.10	0.10	0.10
C		0.10	0.13	0.13
P		1.00	1.00	1.00
LS		1.30	1.30	1.30
Soil evaporation	$in/(day)^{1/2}$	0.122	0.122	0.122
Crack-flow factor		0	0	0

These soils generally consist of deep, well-drained, medium and moderately coarse-textured gravelly soils. Because the same soil occurred in each of the three field elements simulated in the model, only one data set is included in table 2. The seventh layer of the model was assumed to have zero saturated hydraulic conductivity to simulate the caliche layer, which persists throughout the area. This layer is synonymous with the limit of the most active root layers. In our experience, using a

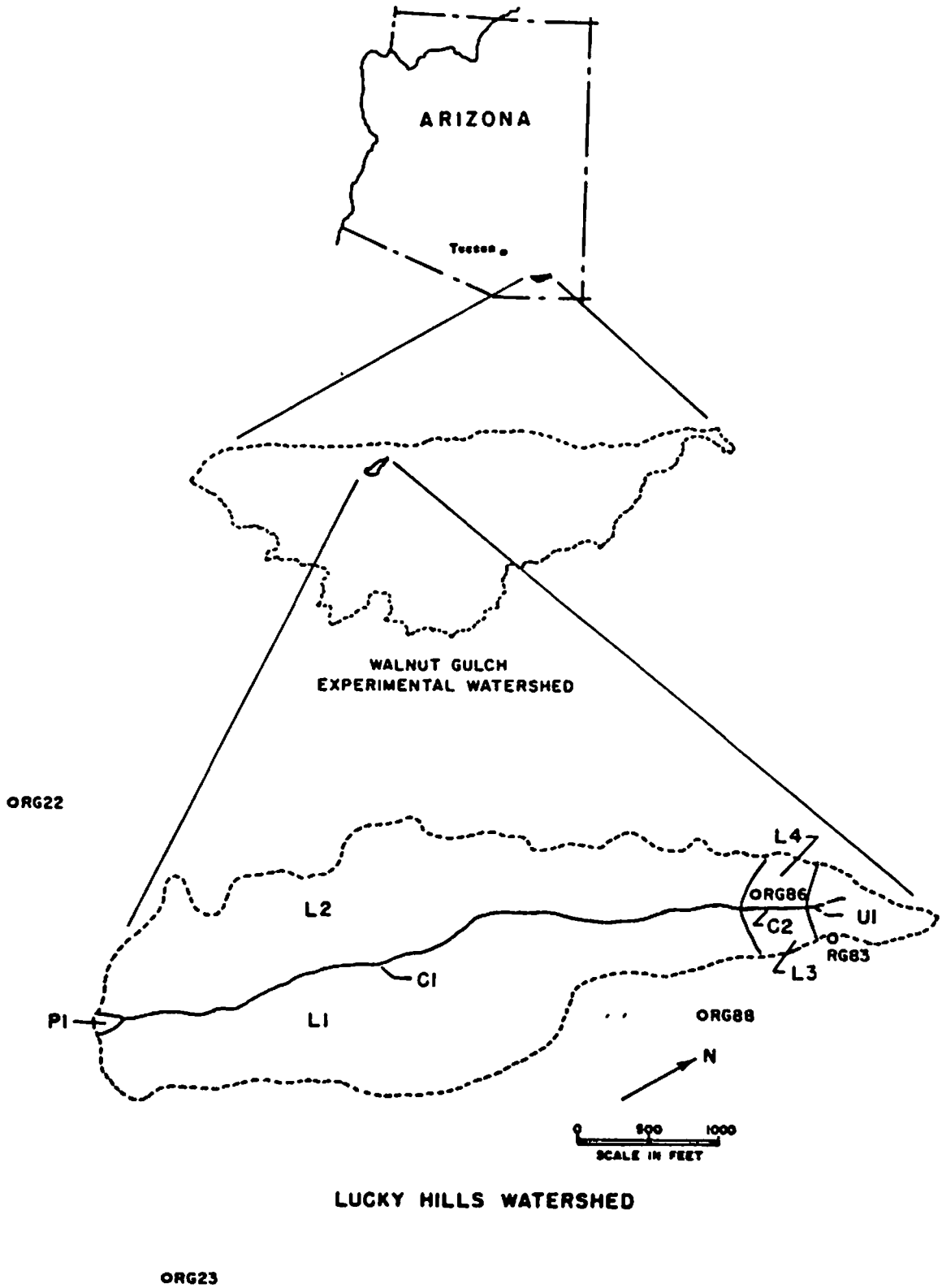


Figure 3.--Location map of the Lucky Hills watershed used in the model evaluation. There are two lateral areas (L1-L2), one upland area (U1), one pond (P1), and a single channel reach.

Table 2.--Soil data for upland areas in SPUR hydrology model

	Soil layer parameters							
	Layer 1	Layer 2	Layer 3	Layer 4	Layer 5	Layer 6	Layer 7	Layer 8
Soil porosity (in/in)	0.440	0.440	0.440	0.400	0.400	0.400	0.400	0.400
Water at .3 bar (in/in)	.120	.120	.120	.120	.120	.120	.120	.120
Water at 15 bar (in/in)	.045	.045	.045	.056	.056	.056	.056	.056
Saturated condition (in/hr)	.500	.450	.300	.300	.300	.300	.000	.300
Soil depth, accumulative (in)	.500	5.000	10.000	15.000	20.000	22.500	25.000	27.000
Field capacity (in)	.037	.337	.375	.320	.320	.160	.160	.128
Maximum storage (in)	.197	1.777	1.975	1.720	1.720	.860	.860	.688

greater soil depth results in the creation of an artificially large soil moisture reservoir, and, in turn, a low curve number which, therefore, simulates lower runoff than the prototype records indicate.

A sample of the output from the hydrology portion of the SPUR model is given in table 4 for 1965. The 11.39 inches of precipitation is very near the average annual for the period of record, but below the normal for the long-term record at the Tombstone, Arizona gage about three miles from the watershed. Monthly values of infiltration, evaporation, and plant transpiration are very representative of those for normal conditions in this environment. The table summarizes what the model predicts will happen from the fields (upland and lateral areas), from the soil profile, in the channels, and, finally, the net yield of sediment from the fields as well as the fine material (silt and clay) and coarse material (bedload) from the channels.

Table 3.--Input data for calculating potential evapotranspiration in Spur hydrology model

Julian date	Temperature (°F)	Radiation (L)
1	46.9	327
12	46.0	341
22	45.8	359
35	46.1	390
46	47.0	420
56	48.5	451
66	50.1	484
110	61.3	628
175	76.7	714
185	77.8	707
195	78.5	694
205	78.7	676
215	78.7	653
225	78.0	626
235	76.7	596
245	75.1	564
300	61.4	388
350	49.2	319
360	47.8	322
365	47.0	325

Table 4.--Sample output from the simulation with the SPUR hydrology model for 1965 on the 108-acre Lucky Hills watershed using measured daily precipitation

	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Year
						FIELDS							
Rainfall	0.580	0.078	0.231	0.032	0.039	0.241	3.299	1.904	1.134	0.000	0.247	3.609	11.392
Infiltration	.580	.078	.231	.032	.039	.241	3.069	1.480	1.109	.000	.247	3.237	10.340
Runoff	.000	.000	.005	.000	.000	.000	.231	.424	.025	.000	.000	.372	1.052
						SOIL							
Return flow	.000	.000	.000	.000	.000	.000	.000	.000	.000	.000	.000	.000	.000
Soil evap	.913	.077	.224	.032	.039	.031	1.794	1.292	1.066	.000	.154	1.382	7.204
Plant evap	.009	.000	.007	.000	.000	.010	.587	.696	.222	.000	.004	.105	1.640
Deep perc	.000	.000	.000	.000	.000	.000	.000	.000	.000	.000	.000	.000	.000
Storage	.000	.000	.000	.000	.000	.000	.688	.179	.000	.000	.089	1.838	1.838
						CHANNEL							
Losses	.000	.000	.005	.000	.000	.000	.057	.058	.016	.000	.000	.077	.208
Runoff	.000	.000	.000	.000	.000	.000	.173	.366	.009	.000	.000	.295	.844
Peak	.0	.0	.0	.0	.0	.0	32.6	76.2	2.0	.0	.0	31.6	76.2
Minimum CN*	100.00	100.00	100.00	100.00	100.00	100.00	84.07	84.65	84.25	100.00	100.00	85.09	84.07
Average CN*	.00	.00	.00	.00	.00	.00	84.73	84.65	84.25	.00	.00	85.88	85.02
Maximum CN*	.00	.00	.00	.00	.00	.00	85.38	84.65	84.25	.00	.00	85.67	86.67
						SEDIMENT							
Field Sediment	.00	.00	.03	.00	.00	.00	19.37	40.66	1.75	.00	.00	32.07	93.85
Silt-clay	.00	.00	.00	.00	.00	.00	4.06	12.36	.08	.00	.00	7.17	23.68
Bedload	.00	.00	.00	.00	.00	.00	15.75	42.73	.29	.00	.00	27.89	86.66

Note: Water = inches peak flow = cfs; sedimentation = tons. 1 acre-ft of water is 0.1111 inches over the watershed.

*When there is no runoff for the month in question, the computer program produces the indicated values.

The output from the channel routing is that documented by L. J. Lane in the subsequent section titled, "SPUR Hydrology Component: Water Routing and Sedimentation."

A 17-year simulation with the SPUR hydrology component was performed to compare with actual data from the Lucky Hills watershed for 1965-81. Figures 4 and 5 illustrate the agreement between the predicted and observed runoff for the upland area and that of the entire area. The relatively poor agreement between the observed and predicted data, as evidenced by the regression statistics in figure 4b, results largely from the 1975 data where the 2.10-inch simulation seriously underestimates the 2.96 inches of observed runoff. Without this one year, the slope of the regression line is much closer to unity.

In figure 6, the cumulative observed and predicted runoff are shown for the annual runoff as predicted with two different curve numbers. Again, the problem of the 1975 data shows with the large departure from the one-to-one line. With the curve number equal to 87, the cumulative runoff at the end of the 17 years overpredicted the observed results. The sensitivity of the curve number model is illustrated with this figure.

Figure 7 illustrates the annual variability of precipitation, evapotranspiration, and transmission losses from the upland area and the entire 108 acre Lucky Hills watershed. As would be expected, the ET follows the precipitation fairly closely, with some noticeable exceptions like that in 1966. In 1966, the computed ET actually exceeds the precipitation because of some soil moisture carryover from the fall of 1965. In addition, the underestimation of the runoff meant there was additional soil moisture for evaporation and transpiration in 1966. Transmission losses are notably larger on the larger watershed as would be expected.

To test agreement of simulated and actual sediment yield with the MUSLE relationship in SPUR, data was available from the upland area (9.1 acres) (fig. 3) for 1965-81. Correlation coefficient of 0.92 and an intercept near zero with a slope of 1.1 indicates a close relationship between field-measured and simulated values.

CONCLUSIONS

A model has been developed which facilitates describing the spatial variability of soils, vegetation, and topography. By allowing such spatial physiographic variability, differences in hydrologic process magnitudes can be accommodated, including those which are restricted to the upland areas as contrasted from those that happen in stream channels. Although testing of the model must be expected to continue, the fundamental precepts behind the development are felt to be in sufficient detail to facilitate describing the heterogeneity encountered in most rangeland conditions.

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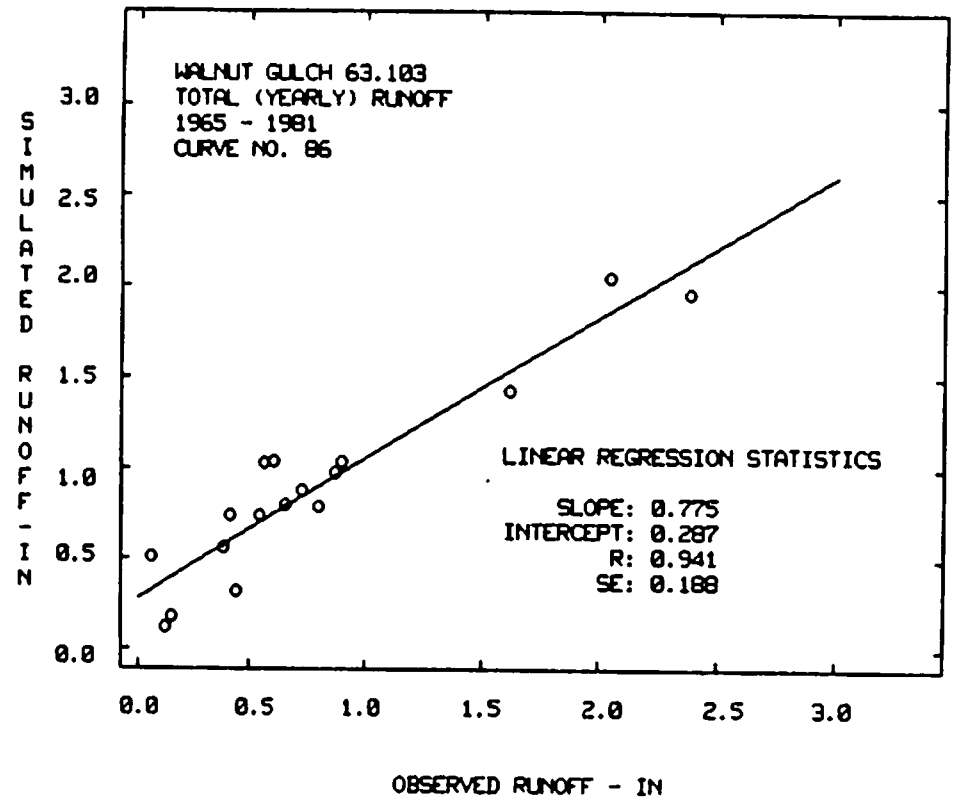


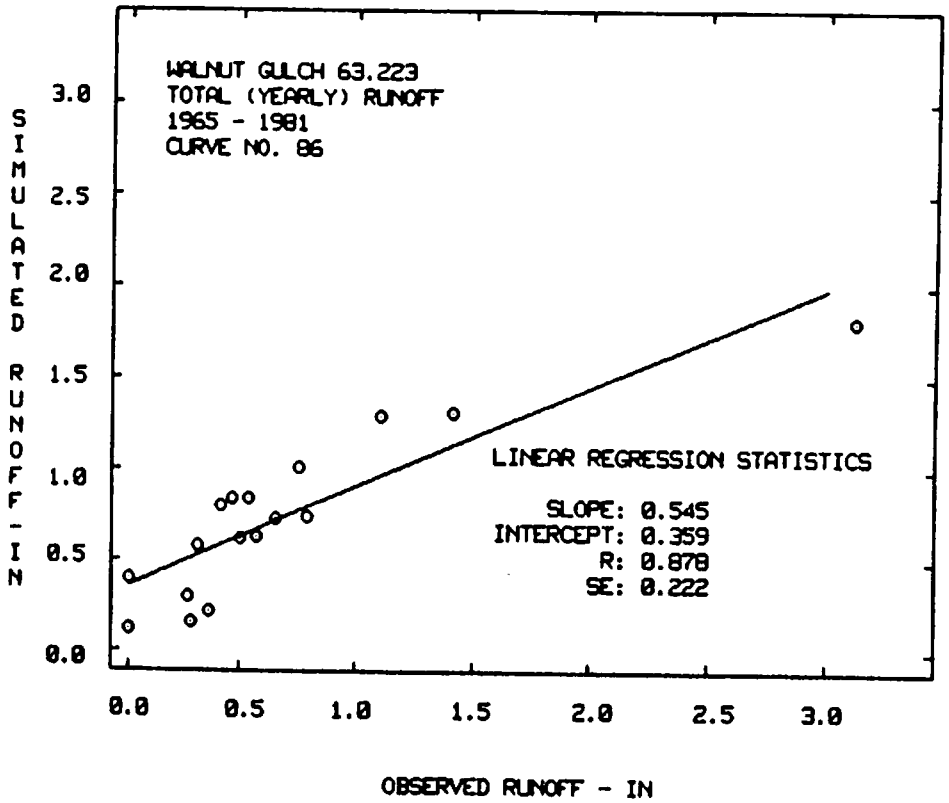
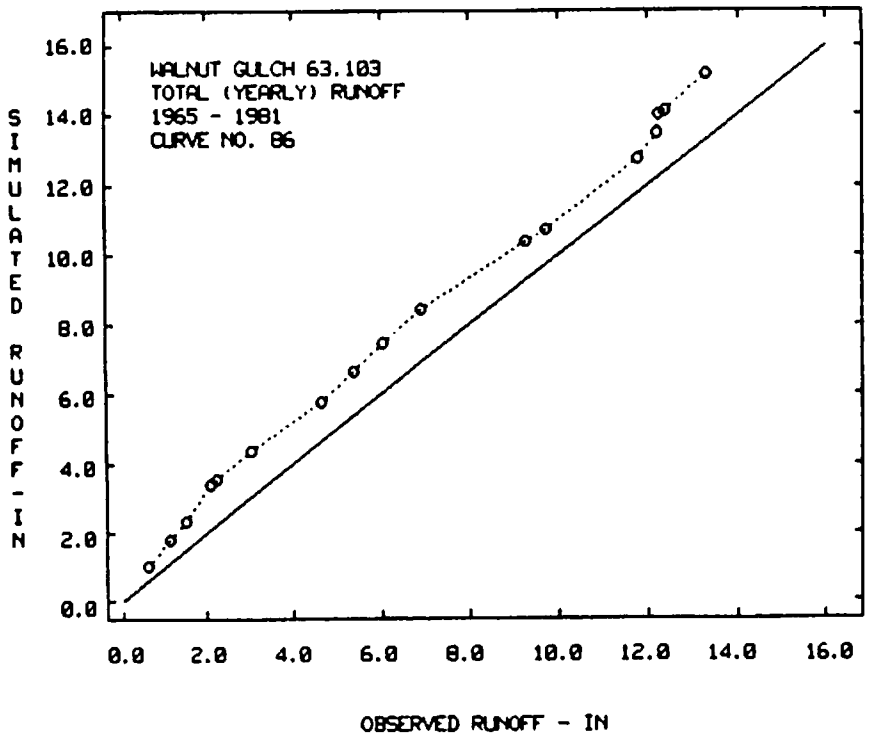


Figure 4.--Simulated versus observed runoff for the upland area and the entire 108-acre Lucky Hills watershed.



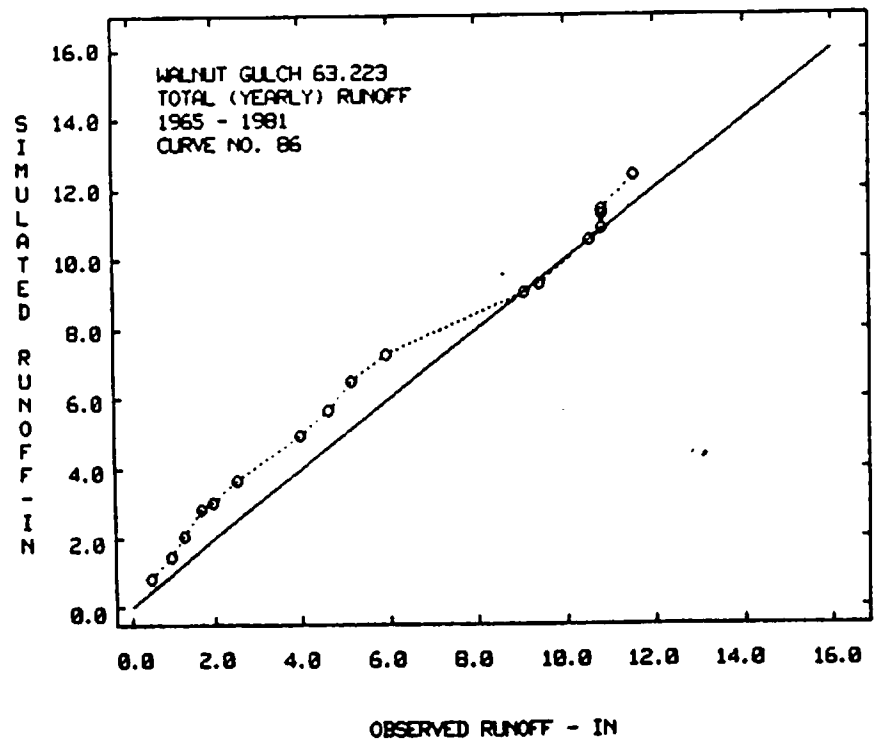


Figure 5.—Cumulative predicted versus actual runoff from the 9.1-acre upland area and for the entire 108-acre Lucky Hills watershed.

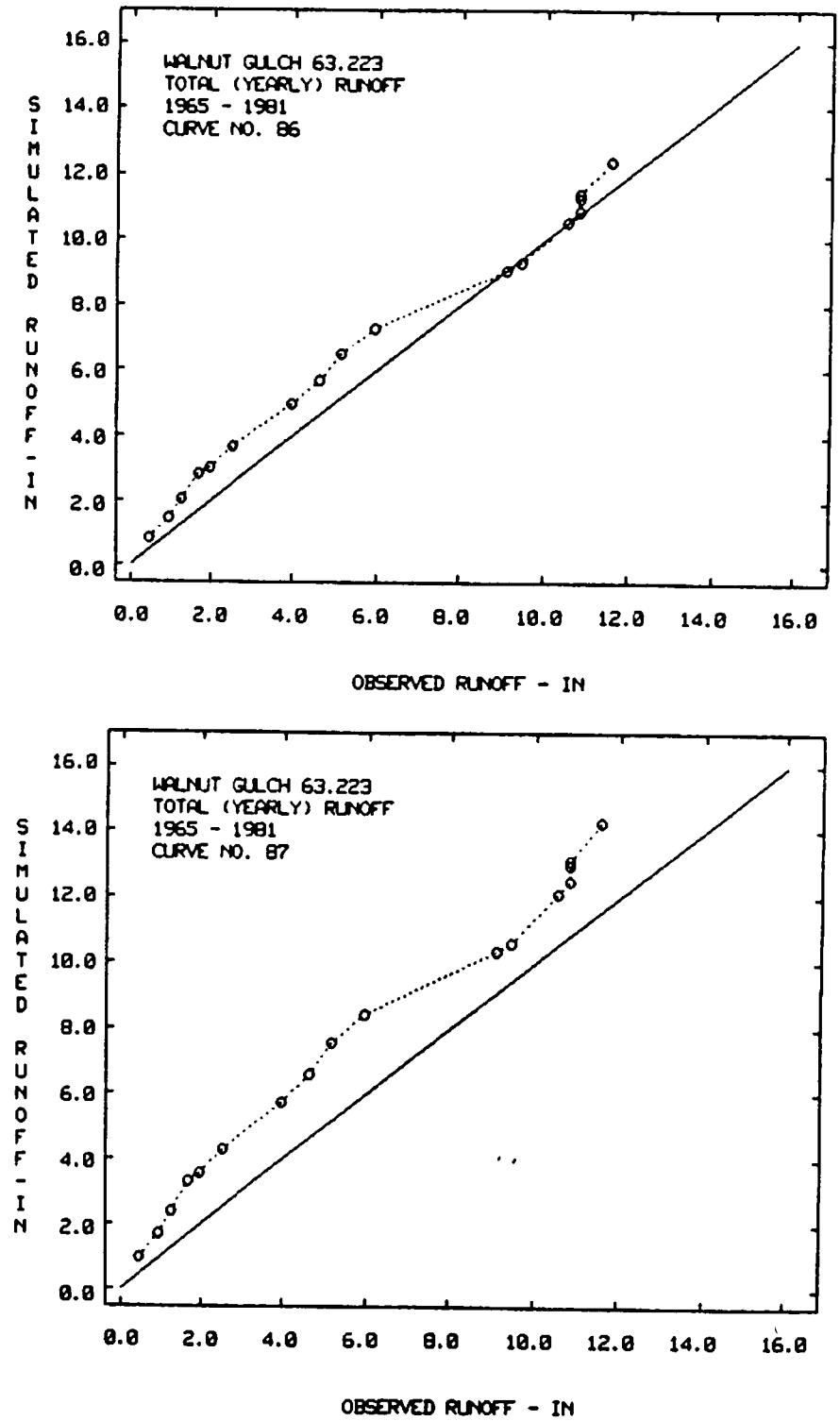
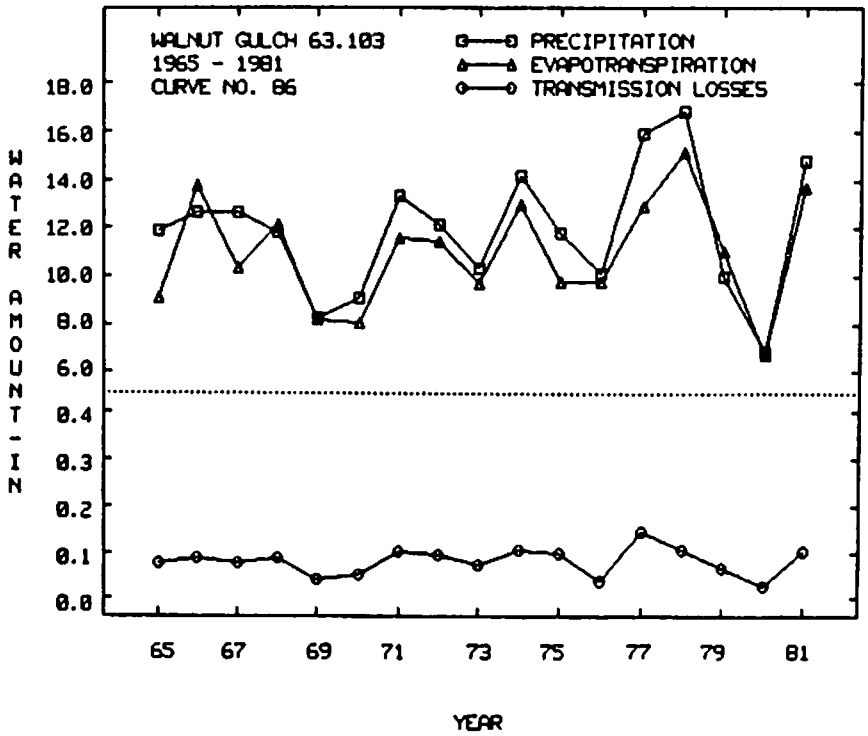


Figure 6.--Cumulative predicted versus actual runoff for the 108-acre Lucky Hills watershed for two different curve numbers.



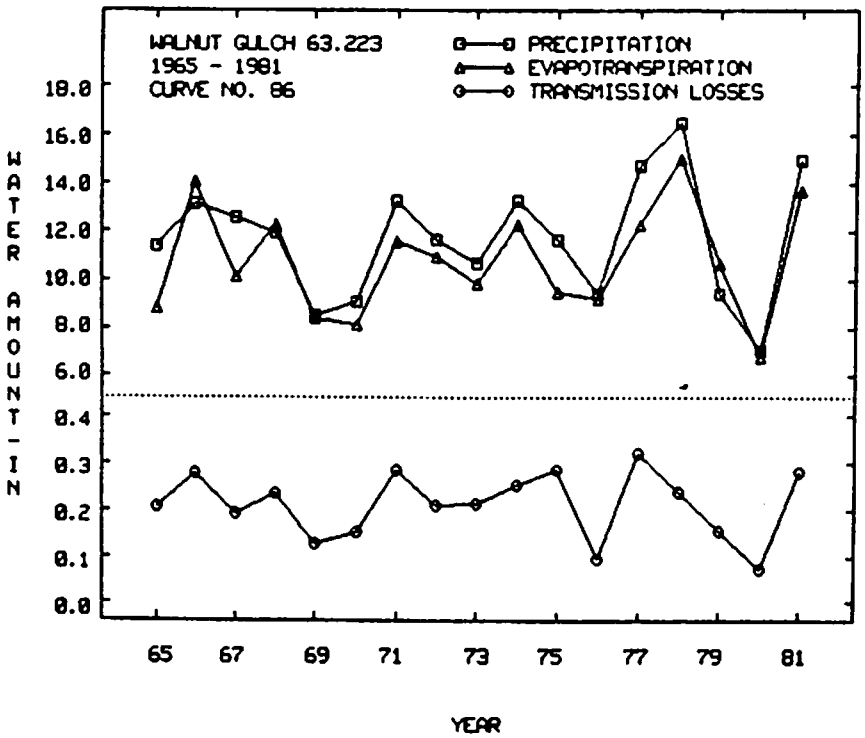


Figure 7.--Precipitation, evapotranspiration, and transmission losses are simulated daily and summed annually. The variability is appreciable from year to year and the transmission losses are much larger for the entire watershed (bottom) than for the 9.1-acre upland area (top).

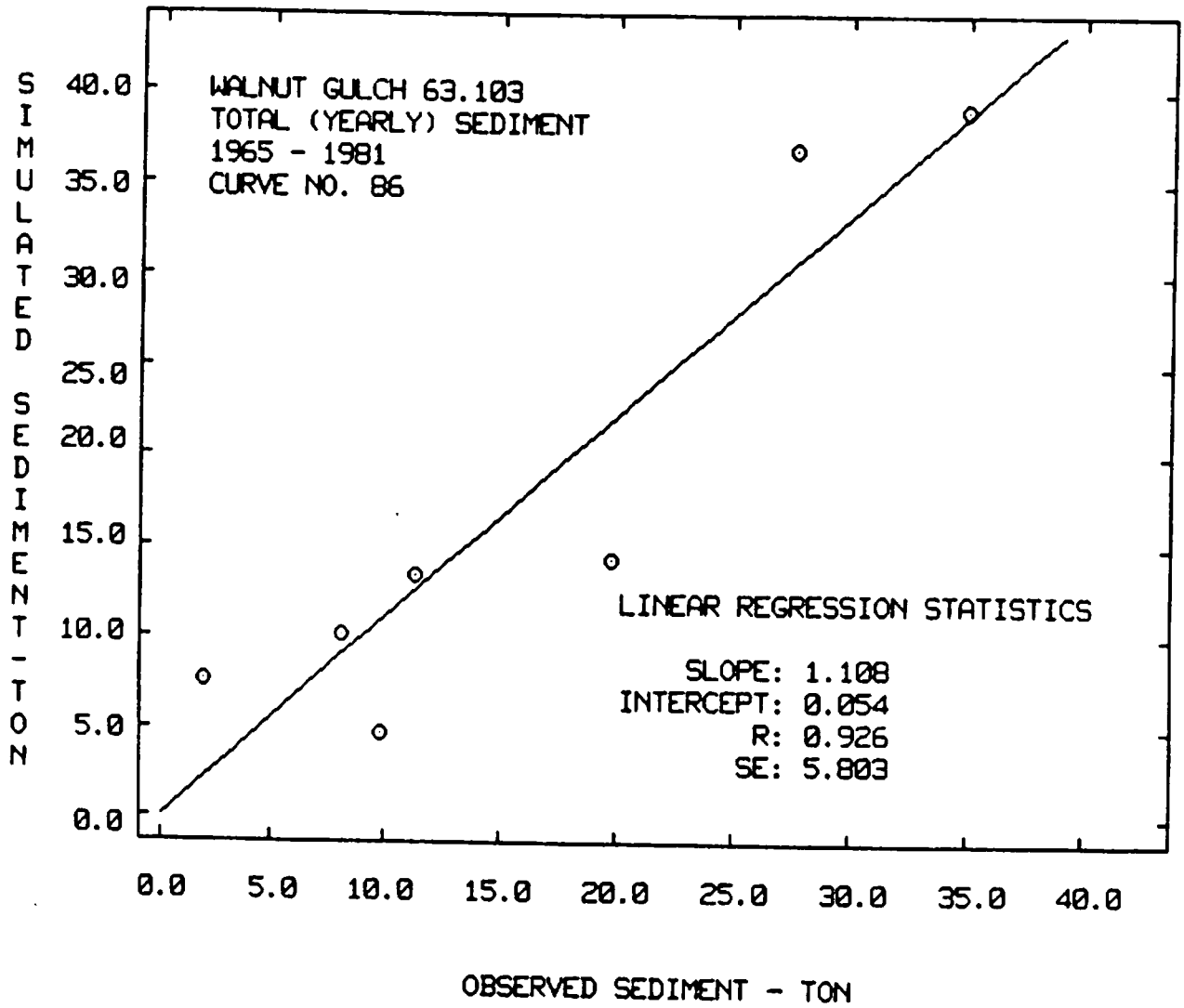


Figure 8.--Predicted versus simulated annual sediment yield using MUSLE for the 9.1-acre upland area of the Lucky Hills watershed.

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