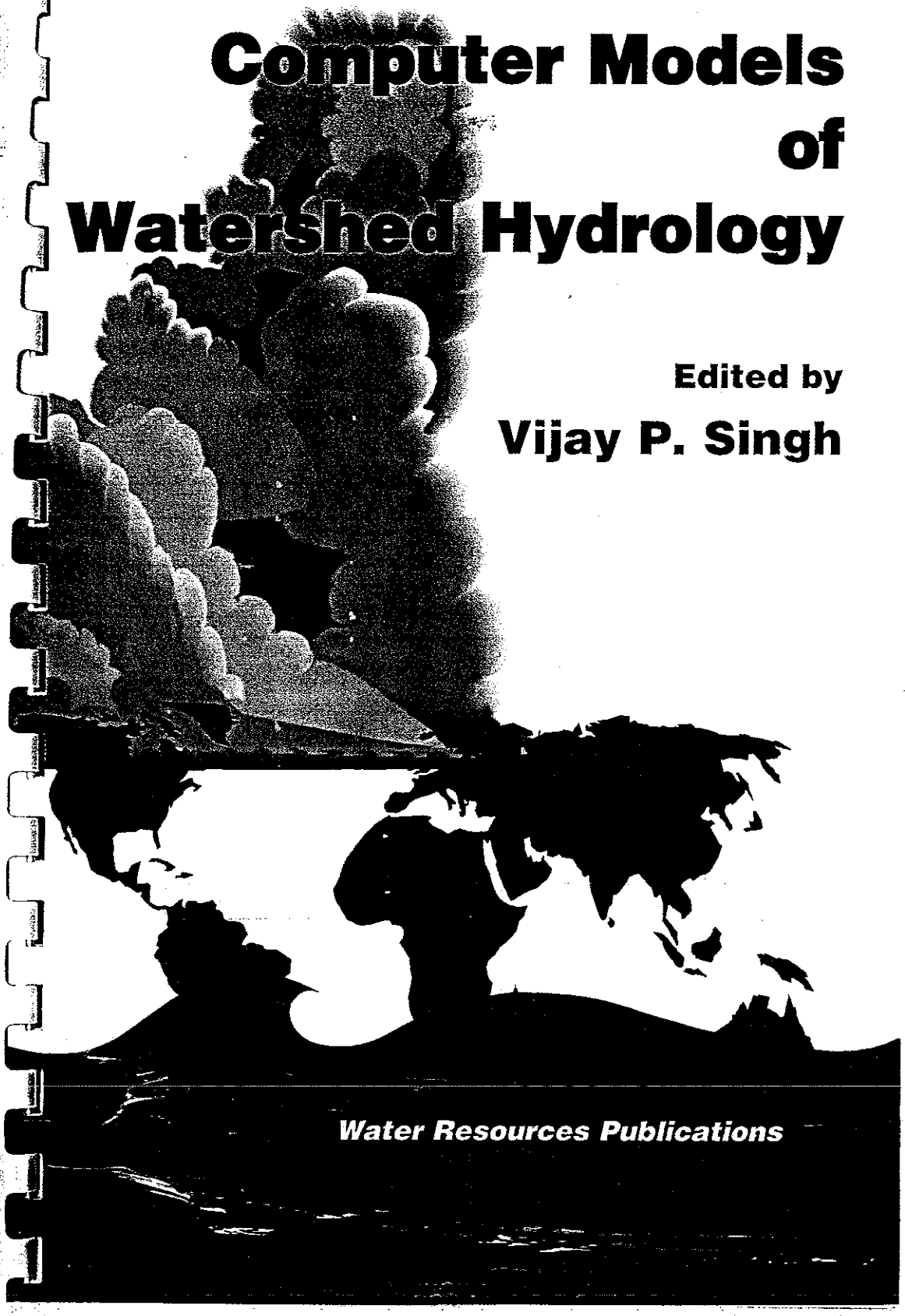
Computer Models of Watershed Hydrology



Edited by

Vijay P. Singh

Water Resources Publications

Chapter 9

THE PRECIPITATION-RUNOFF MODELING SYSTEM - PRMS

G. H Leavesley and L. G. Stannard

9.1. INTRODUCTION

The Precipitation-Runoff Modeling System (PRMS) (Leavesley et al., 1983) is a modular-design, distributed-parameter, physical-process watershed model that was developed to evaluate the effects of various combinations of precipitation, climate, and land use on watershed response. Watershed response to normal and extreme rainfall and snowmelt can be simulated to evaluate changes in water-balance relations, flow regimes, flood peaks and volumes, soil-water relations, sediment yields, and ground-water recharge. Parameter-optimization and sensitivity-analysis capabilities are provided to fit selected model parameters and to evaluate their individual and joint effects on model output.

Selected model enhancements have been made and support programs have been coupled with PRMS to produce a watershed-modeling and data-management system that provides a range of hydrologic-simulation and data-analysis capabilities. Data-management and analysis components are provided by the U.S. Geological Survey's data-management program ANNIE (not an acronym) (Lumb et al., 1990). ANNIE is an interactive program that accepts and reformats data from a variety of sources for model application. ANNIE also includes a number of statistical, graphical, and data-management functions for analysis of measured and simulated data.

Hydrologic forecasting capabilities are provided by a modified version of the U.S. Weather Service's Extended Streamflow Prediction (ESP) program (Day, 1985). ESP applications include short-term and seasonal forecasting for floods and water supply, and evaluation of the effects of land-use and climate changes on hydrologic response.

The purpose of this paper is to provide an overview of PRMS, its system components, and system applications. More detailed discussions regarding individual components can be found in cited references for each component. The major components of the system, and the data transfers among these components, are depicted schematically in Fig. 9.1 and described hereafter.

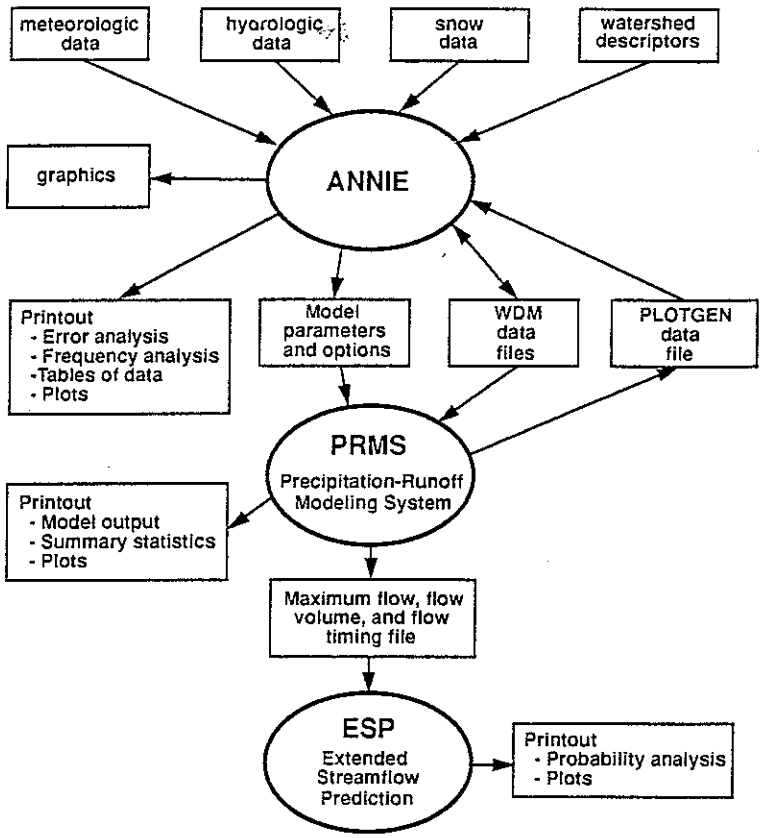


Fig. 9.1. Schematic diagram of the data transfers among PRMS.ANNIE. and ESP.

9.2. ANNIE

ANNIE is a modular-design interactive program that provides the data-management and analysis functions for PRMS. ANNIE is composed of software modules designed to help a user interactively create, check, and update input to hydrologic models, and to provide statistical and graphical tools to assist in the analysis of model input and output. Data from a variety of sources are reformatted by ANNIE to a system-compatible file structure. Meteorologic, hydrologic, and snow data are placed in a direct-access file structure called the Watershed Data Management (WDM) file. Data in the WDM file can be analyzed using a number of statistical, graphical, and data-management functions. Model- and watershed-specific data can be entered

interactively into ANNIE to provide attribute information for selected data sets.

A WDM file is composed of one or more data sets, each of which is identified by a user-specified data-set number. Stored with each data set are a variety of user-selectable attributes that can be used to describe the type, geographic location, and other characteristics of the data set.

New data sets can be generated from existing data sets using a number of selectable procedures. One procedure generates a new data set by applying a mathematical operation to one, or selected combinations of two, data sets. Mathematical functions include log transformations, power transformations, weighted sums, and the addition or subtraction of, and multiplication or division by, constants. A second procedure generates a new data set by changing the time step of an existing data set. A disaggregate transformation expands a time series into shorter time steps. An aggregate transformation condenses a time series into longer time steps.

A plot component provides the graphic capabilities in ANNIE. Data sets in the WDM file, results of selected statistical analyses, and PRMS results placed in a PLOTGEN file can be displayed graphically on a monitor or on a pen plotter. Graph formats include time-series plots of one to five data sets, x-y plots of selected data sets, frequency plots, flow-duration plots, and scatter diagrams.

9.3. TIME AND SPACE IN PRMS

Watershed response can be simulated at both a daily and a storm time scale in PRMS. The daily mode simulates hydrologic components as daily average or total values. Streamflow is computed as a mean daily flow. The storm mode simulates selected hydrologic components at time intervals shorter than 1 day. The minimum time interval is 1 minute. The storm mode computes storm hydrographs and sediment yields for selected rainstorms. Sediment-modeling capabilities are provided only in the storm mode.

Distributed-parameter capabilities are provided by partitioning a watershed into units, using characteristics such as slope, aspect, elevation, vegetation type, soil type, and precipitation distribution. Each unit is assumed to be homogeneous with respect to its hydrologic response and to the characteristics listed above; each unit is called a hydrologic response unit (HRU). A water balance and an energy balance are computed daily for each HRU. The sum of the responses of all HRUs, weighted on a unit-area basis, produces the daily watershed response.

A second level of spatial partitioning is used for storm-mode computations. The watershed is conceptualized as a series of interconnected flow-plane and channel segments. An HRU can be considered the equivalent of a single flow plane, or it can be delineated into a number of flow planes.

9.4. PRMS DAILY-MODE COMPONENTS

The daily-mode components simulate the daily accretion, depletion, storage, and movement of water in a HRU. The rates and volumes of these processes are a function of HRU physical, hydrologic, and climatic characteristics. The conceptualized watershed system and its inputs are schematically depicted in Fig. 9.2. Brief descriptions of the components that comprise major hydrologic-process subsets of the watershed system and those that provide model-support capabilities are presented hereafter.

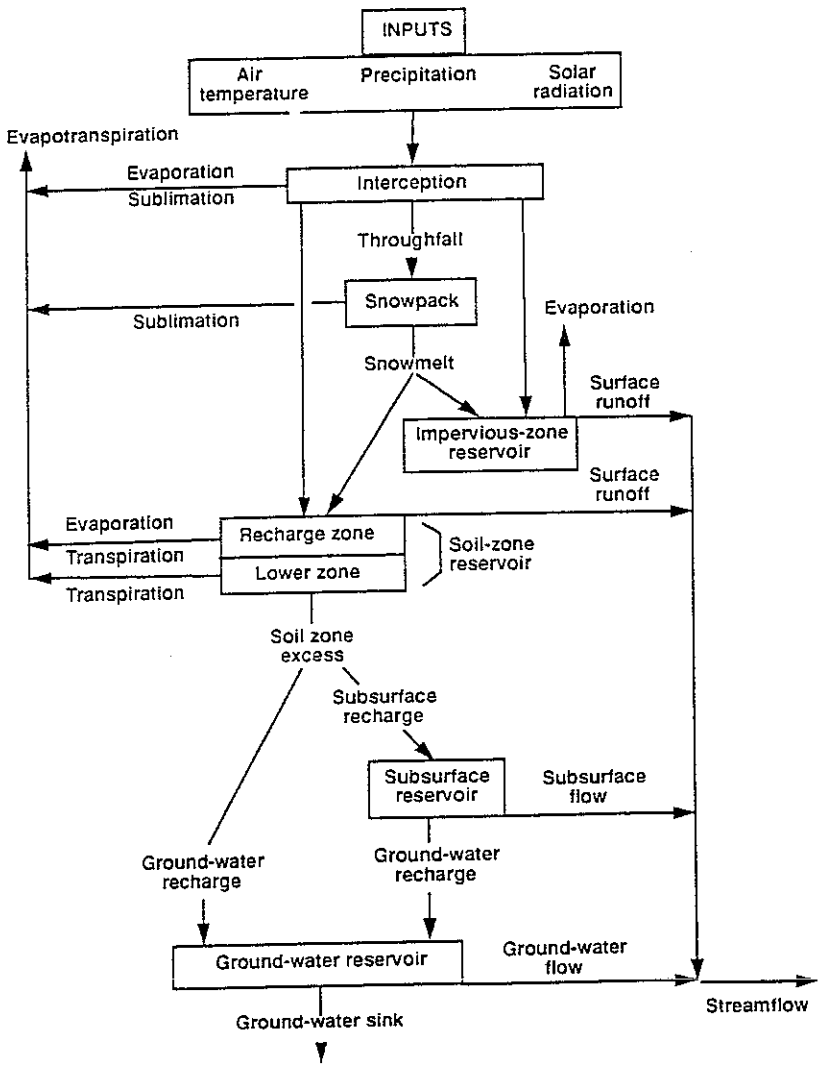


Fig. 9.2. Schematic diagram of the conceptual watershed system and its inputs (modified from Leavesley et al., 1983).

9.4.1. Climate Components

Model inputs are daily precipitation, maximum and minimum air temperature, and solar radiation. Precipitation in the form of rain, snow, or a mixture of both is decreased by interception and becomes net precipitation delivered to the watershed surface. The energy inputs of air temperature and solar radiation are used in the computation of evaporation, transpiration, sublimation, and snowmelt. Daily climate data are extrapolated to each HRU using a set of user-defined adjustment coefficients developed from regional climate data.

Temperature

Measured maximum (TMX) and minimum (TMN) daily air-temperature data are adjusted using monthly or daily lapse rates and the elevation difference between a climate station and each HRU. Up to five climate stations can be used. When using monthly lapse rates, a correction factor (TCRX) for adjusting TMX for each HRU is computed for each month (MO) by

$$TCRX(MO) = [TLX(MO) \bullet ELCR] - TXAJ \tag{9.1}$$

where

TLX is the lapse rate for maximum daily air temperature, in degrees per 1000 ft change in elevation for month MO,

ELCR is mean elevation of an HRU minus the elevation of the climate station associated with the HRU, in 1000's of ft, and

TXAJ is an average difference in maximum air temperature between a horizontal surface and the slope-aspect of an HRU.

A correction factor [TCRN(MO)] for adjusting TMN for each HRU is computed monthly with an equation of the same form as (9.1), using the monthly lapse rate for minimum daily air temperature [TLN(MO)] and a minimum temperature slope-aspect correction (TNAJ). An option is provided to compute TLX and TLN daily using the maximum and minimum daily air temperatures from a high-elevation and a low-elevation temperature station.

Adjusted daily maximum air temperature (TM) for each HRU is computed by

$$TM = TMX - TCRX(MO) \tag{9.2}$$

Adjusted daily minimum air temperature (TN) for each HRU is computed using an equation of the same form as eq. (9.2) and the variables TMN and TCRN(MO).

Precipitation

Total daily precipitation depth (PPT) received on an HRU is computed by

$$PPT = PDV \bullet PCOR \tag{9.3}$$

where

PDV is daily precipitation depth measured in the precipitation gage associated with the HRU, in inches (in.), and
PCOR is the correction factor for the HRU.

A maximum of five gages can be used to define the spatial and temporal variation of precipitation on a watershed. A PCOR value is defined for each HRU and accounts for the effects of elevation, spatial variation, topography, gage location, deficiencies in gage catch due to the effects of wind, and other factors. Different values of PCOR can be specified for summer and winter periods to account for different precipitation forms and for changes in storm characteristics. PCOR is set to parameters DRCOR for daily rainfall, DSCOR for daily snowfall, and UPCOR for storm rainfall.

In some geographic and climatic regions, the influence of elevation on precipitation volume varies seasonally with general storm patterns (Barry, 1972). Frontal origin storms may show a strong elevation effect, while convective storms show little elevation effect. If DRCOR and DSCOR include strong elevational influence, they can be overridden for a user-defined period where convective storms are dominant to account for these seasonal changes.

Precipitation form (rain, snow, or mixture of both) on each HRU is estimated from the HRU maximum (TM) and minimum (TN) daily air temperatures and their relation to a base temperature (BST) (Willen et al., 1971). Precipitation is all snow if TM is less than or equal to BST, and all rain if TN is greater than BST. If TM is greater than BST and TN is less than BST, then the part of the total precipitation occurring as rain (PRMX) is computed by

$$PRMX = \left(\frac{TM - BST}{TM - TN} \right) \bullet AJMX(MO) \tag{9.4}$$

where AJMX is a monthly mixture adjustment coefficient for month MO. For mixed events, rain is assumed to occur first.

The precipitation form estimation procedure can be overridden in two ways. One is the use of a parameter PAT. PAT is an air temperature value which when exceeded by TM, forces all precipitation to be rain, regardless of the TN value. This capability permits accounting for convective rainfall during spring, when TN may be below BST. PAT is user-specified for each month. The second override is for days when the form is known. Precipitation form can be specified in the model input. Also, snowpack volumes on each HRU can be adjusted using measured snowcourse data.

Shortwave Radiation

Daily shortwave radiation, expressed in langleys per day (ly/d), is used in snowmelt computations and can be used in evapotranspiration computations. Shortwave radiation is an optional input for areas with no snowmelt. Measured shortwave radiation (ORAD), from a horizontal surface, is adjusted to estimate daily shortwave radiation received on the slope-aspect combination of each HRU (SWRD) using a method described by Swift (1976). SWRD is computed by

$$SWRD = ORAD \bullet \frac{DRAD}{HORAD} \bullet \frac{1.}{COSSLP} \tag{9.5}$$

where

DRAD is daily potential solar radiation for the slope and aspect of an HRU (ly),
HORAD is daily potential solar radiation for a horizontal surface (ly), and
COSSLP is cosine of the slope inclination of an HRU.

DRAD and HORAD are linearly interpolated from a table of 13 potential solar radiation values that are calculated for each HRU slope-aspect combination and the horizontal surface. Because of the symmetry of the solar year, these 13 values represent the potential solar radiation for 24 dates throughout the year. These 13 values are computed using a combination of methods described in Frank and Lee (1966) and Swift (1976).

For periods of missing solar radiation data or where these data are unavailable, ORAD can be estimated using air temperature, precipitation, and computed potential solar-radiation data. One of two procedures can be selected. The first is a modification of the degree-day method described by Leaf and Brink (1973). This method was developed for a section of the Rocky Mountain region of the United States. It is most applicable to this region where predominantly clear skies prevail on days without precipitation.

The second procedure uses a relation between sky cover and daily range in air temperature demonstrated by Tångborn (1979) and a relation between solar radiation and sky cover developed by Thompson (1976). This procedure is applicable to more humid regions where extensive periods of cloud cover occur with and without precipitation.

9.4.2. Land Phase Components

Interception

Interception is computed as a function of vegetation cover density and the storage available on the predominant vegetation of an HRU. Net precipitation (PTN) on an HRU is computed by

$$PTN = [PPT \bullet (1. - COVDN)] + [PTF \bullet COVDN] \tag{9.6}$$

where

PPT is total precipitation received on an HRU (in.),
COVDN is seasonal cover density, and
PTF is precipitation falling through the canopy (in.).

COVDN is defined for summer (COVDNS) and winter (COVDNW). PTF is computed by

$$\begin{aligned} PTF &= PPT - (STOR - XIN) \quad & PPT > (STOR - XIN) \\ PTF &= 0. & PPT \leq (STOR - XIN) \end{aligned} \tag{9.7}$$

where

STOR is maximum interception storage depth on vegetation (in.), and
XIN is the current depth of interception storage (in.).

STOR is defined by season and precipitation form; winter snow (SNST), winter rain (RNSTW), and summer rain (RNSTS). When rain falls on intercepted snow, STOR changes from SNST to the seasonal rain-storage value. Any snow in storage that exceeds the new rain-storage value is added to throughfall PTF. When snow falls on intercepted rain, STOR changes to SNST and currently available rain interception is used to satisfy SNST. Intercepted rain in excess of SNST is added to PTF.

Intercepted rain is assumed to evaporate at a free-water-surface rate (EVCAN). If pan evaporation data are used, EVCAN equals the pan loss

rate. If potential evapotranspiration (PET) is computed from meteorologic variables, EVCAN is computed by

$$EVCAN = \frac{PET}{EVC(MO)} \tag{9.8}$$

where EVC is the evaporation-pan coefficient for month MO. Sublimation of intercepted snow is assumed to occur at a rate that is expressed as a user-defined percentage of PET. In addition, intercepted snow can be removed from the canopy by melting. An energy budget is computed for a 12-hour period (assumed 0600 to 1800). If the energy balance is negative or zero, no melt occurs, When the energy balance is positive, all melt is delivered to the snowpack or soil surface as net precipitation.

Soil-Moisture Accounting

The watershed system is conceptualized as a series of reservoirs, the responses of which combine to produce the total watershed response. The soil-zone reservoir represents that part of the soil mantle that can lose water through the processes of evaporation and transpiration. Average rooting depth of the predominant vegetation covering the soil surface defines the depth of this zone. The maximum available water-holding capacity of the soil-zone reservoir (SMAX) is the difference between field capacity and wilting point of the profile.

The soil-zone reservoir is treated as a two-layered system. The upper zone is termed the recharge zone and the remaining profile is the lower zone. The recharge zone has a user definable maximum available water-holding capacity (REMX) and losses from this zone occur from evaporation and transpiration; losses from the lower zone are assumed to occur only through transpiration. The maximum available water-holding capacity of the lower zone is the difference between SMAX and REMX.

Water storage in the soil-zone reservoir is increased by infiltration of rainfall and snowmelt and is decreased by evapotranspiration. Evapotranspiration losses occur at a rate that is a function of available soil-moisture storage and soil type. The attempt to satisfy potential evapotranspiration is made first from the recharge zone. Soil-moisture accretions must fill the recharge zone before water will move to the lower zone. When the soil-zone reservoir reaches SMAX, additional infiltration is routed to the subsurface and ground-water reservoirs.

The apportioning of water in excess of SMAX to the subsurface and ground-water reservoirs is done using a ground-water recharge parameter SEP. Daily ground-water reservoir inflow is equal to or less than SEP depending on the magnitude of soil-water excess. When soil-water excess is larger than SEP, the difference becomes subsurface reservoir inflow.

Evapotranspiration

Three procedures are available to compute potential evapotranspiration (PET). One procedure uses daily pan-evaporation data and a monthly pan-adjustment coefficient. PET (in./day) is computed by

$$PET = EPAN \bullet EVC(MO) \tag{9.9}$$

where

EPAN is daily pan evaporation (in.), and
EVC is a monthly pan-adjustment coefficient for month MO.

A second procedure computes PET as a function of daily mean air temperature and possible hours of sunshine (Hamon, 1961). PET is computed by

$$PET = CTS(MO) \bullet (DYL)^2 \bullet VDSAT \tag{9.10}$$

where

CTS is a coefficient for month MO,
DYL is possible hours of sunshine, in units of 12 hours, and
VDSAT is the saturated water vapor density (absolute humidity) at the daily mean air temperature, in g/m^3.

VDSAT is computed using empirical relations between VDSAT and saturated vapor pressure (Federer and Lash, 1978) and between saturated vapor pressure and mean daily air temperature (Murray, 1967). DYL varies with time of year and HRU slope and aspect. A table of 13 DYL values is computed for each HRU for the same 13 dates used in the solar radiation estimation procedure. Daily DYL values for each HRU are linearly interpolated from this table of 13 values.

The third procedure is a modified Jensen-Haise technique (Jensen et al., 1969) that computes PET using air temperature, solar radiation, and two coefficients that can be estimated using regional air-temperature, elevation, vapor-pressure, and vegetation data. PET is computed by

$$PET = CTS(MO) \bullet (TAVF - CTX) \bullet RIN \tag{9.11}$$

where

CTS is a coefficient for the month MO,
TAVF is the daily mean air temperature (° F),
CTX is a coefficient, and

RIN is daily solar radiation expressed in inches of evaporation potential.

Actual evapotranspiration (AET) is computed as a function of soil type, water currently available in the soil-zone reservoir, and the storage capacity of the soil-zone reservoir using an approach developed by Zahner (1967). When available water is nonlimiting, AET equals PET. PET is first satisfied from interception storage, retention storage on impervious surfaces, and evaporation/sublimation from snow surfaces. Remaining PET demand then is applied to the soil-zone storage. AET is computed separately for the recharge zone and the lower zone using the unsatisfied PET demand and the ratio of currently available water in the soil zone to its maximum available water-holding capacity. AET computed for the recharge zone is used first to satisfy PET; any remaining demand is attempted to be met by AET from the lower zone. Relations between AET and available soil moisture are defined for three soil types; sand, loam, and clay.

The active transpiration period is user defined by a beginning and ending month. The specific date of the start of transpiration is computed for each HRU using an HRU temperature parameter TST. The sum of the maximum daily air temperatures is accumulated for each HRU, starting on the first day of the month transpiration begins. When this sum exceeds TST, transpiration is assumed to begin. Transpiration is assumed to end on the first day of the month specified as the month transpiration ends.

Surface Runoff

Net precipitation reaches the snowpack or soil surface where it accumulates in the snowpack or is available for surface runoff and infiltration. Daily surface runoff from rainfall on pervious snow-free HRU's is computed using a contributing-area concept (Dickinson and Whiteley, 1970; Hewlett and Nutter, 1970). Net precipitation not becoming surface runoff infiltrates the soil surface. The percent of an HRU contributing to surface runoff (CAP) can be computed as either a linear or a nonlinear function of antecedent soil moisture and rainfall amount. In the linear case

$$CAP = SCN + \left[(SCX - SCN) \bullet \left\{ \frac{RECHR}{REMX} \right\} \right] \tag{9.12}$$

where

SCN is the minimum possible contributing area,
SCX is the maximum possible contributing area,
RECHR is the current available water in the soil recharge zone (in.), and
REMX is maximum storage capacity of the recharge zone (in.).

Surface runoff (SRO) then is computed by

$$SRO = CAP \bullet PTN \tag{9.13}$$

where PTN is the daily net precipitation.

The nonlinear scheme uses a moisture index (SMIDX) similar to that developed by Dickinson and Whiteley (1970) for estimating CAP. CAP is computed by

$$CAP = SCN \bullet 10.^{(SC1 \bullet SMIDX)} \tag{9.14}$$

where

SCN and SC1 are coefficients, and
SMIDX is the sum of the current available water in the soil zone (SMAV) plus one-half of the daily net precipitation (PTN) (in.).

Surface runoff from snowmelt is computed only on a daily basis. Snowmelt runoff from pervious areas is assumed to occur only when the soil zone of an HRU reaches field capacity. At field capacity, a daily maximum infiltration rate is assumed and user defined. Any daily snowmelt in excess of this maximum infiltration rate is assumed to become surface runoff. For impervious areas, snowmelt first satisfies available retention storage, and the remaining snowmelt becomes surface runoff.

Subsurface Flow

The subsurface reservoir simulates the relatively rapid component of flow that may occur in the saturated-unsaturated and ground-water zones during periods of rainfall and snowmelt. The subsurface reservoir can be defined as being linear or nonlinear. Outflow form the subsurface reservoir (RAS) is computed by

$$RAS = (RCF \bullet RES) + \left(RCP \bullet (RES)^2\right) \tag{9.15}$$

where

RCF and RCP are routing coefficients, and
RES is the storage volume in the subsurface reservoir (in.).

This is a nonlinear relationship. However, setting RCP equal to 0 makes the relationship linear. A subsurface reservoir can receive inflow from one or several HRU's. The number of subsurface reservoirs can range from one to the number of HRU's delineated.

Ground Water

The ground-water reservoir simulates the slower component of flow from the ground-water zone. It is conceptualized as a linear reservoir and is assumed to be the source of all baseflow (BAS). Inflow to the ground-water reservoir can be from both soil-water excess and one or more subsurface reservoirs. The vertical movement of water from a subsurface reservoir to a ground-water reservoir is computed as a function of the current volume of storage in subsurface reservoir and a linear routing coefficient. The shape of the baseflow recession of the simulated hydrograph will be effected by the relative proportion of ground-water recharge from the two source terms. Recharge from the soil zone occurs only on days when SMAX is exceeded while recharge from the subsurface reservoir occurs every day water is available in the subsurface reservoir. BAS is computed by

$$BAS = RCB \bullet GW \tag{9.16}$$

where

RCB is a routing coefficient, and
GW is the ground-water reservoir storage (acre-in.)

The movement of water through the ground-water reservoir to points outside the surface drainage boundary is treated using a ground-water sink which is computed as a function of GW and a linear routing coefficient.

9.4.3. Channel Reservoir Components

Channel reservoir components simulate the storage and routing response of channel reservoirs. Reservoir inflows are computed as the sum of the streamflow contributions from all HRU's and the parts of subsurface and ground-water reservoirs above the channel reservoir. Reservoir inflow also can include the outflow of up to 3 upstream channel reservoirs. Two types of routing procedures are available for simulating reservoir outflow. Both are based on the equation of continuity. One is a linear-storage routing procedure in which outflow (O) is computed as a linear function of storage (STO) by

$$O = RCS \bullet STO \tag{9.17}$$

where RCS is a routing parameter.

The second is a modified-Puls routing procedure (U.S. Soil Conservation Service, 1971). In this procedure the outflow (O) and inflow (I) terms are considered averages of O and I and the beginning and end of the

computational time interval. The continuity equation for this approach is expressed as

$$\frac{2 \bullet STO_2}{\Delta t} + O_2 = (I_2 + I_1) + \frac{2 \bullet STO_1}{\Delta t} - O_1 \qquad (9.18)$$

where the subscripts 1 and 2 specify start and end values for the computational time interval Δt. All terms on the right-hand side of equation (9.18) are known at the beginning of a routing period. An interpolation procedure described by Dawdy et al. (1978) is used to solve for O_2.

9.4.4. Snow Components

The snow components simulate the initiation, accumulation, and depletion of a snowpack on each HRU. A snowpack is maintained and modified on both a water-equivalent basis and as a dynamic-heat reservoir. A snowpack water balance is computed daily and an energy balance is computed twice each day for 12-hour periods (designated day and night).

The energy-balance computations are a combination of equations and functional relationships taken or derived from several sources. The conceptual model for the snowpack system and its energy relations is one described by Obled and Rosse (1977). The snowpack is assumed to be a two-layered system. The surface layer consists of the upper 3-5 centimeters of the snowpack, and the bottom layer is the remaining snowpack. Heat transfer between the surface layer and the snowpack occurs by conduction when the temperature of the surface layer (Ts) is less than 0 °C. When Ts equals 0 °C, heat transfer occurs as conduction when the net energy balance at the air-snow interface is negative; but heat transfer occurs as mass transfer by surface melting when the net energy balance is positive. Heat transfer from precipitation occurs as a mass-transfer process. Conduction of heat from the soil surface to the snowpack is assumed to be negligible compared to the energy exchange at the air-snow interface. The conceptual snowpack system and the components of the snowpack energy balance are shown in Fig. 9.3.

Ts for the daylight period is computed as the mean of the maximum and mean-daily air temperature for an HRU. For the night period, it is computed as the mean of the minimum and mean-daily air temperature for an HRU. Ts equals 0 °C when these means exceed 0 °C. When Ts is less than 0 °C, the mean air temperature for the 12-hour period is assumed to integrate the effects of the radiation, latent heat, sensible heat, and diffusion processes expressed in the complete equation for the surface temperature. See Anderson (1968) for a complete discussion of the equation.

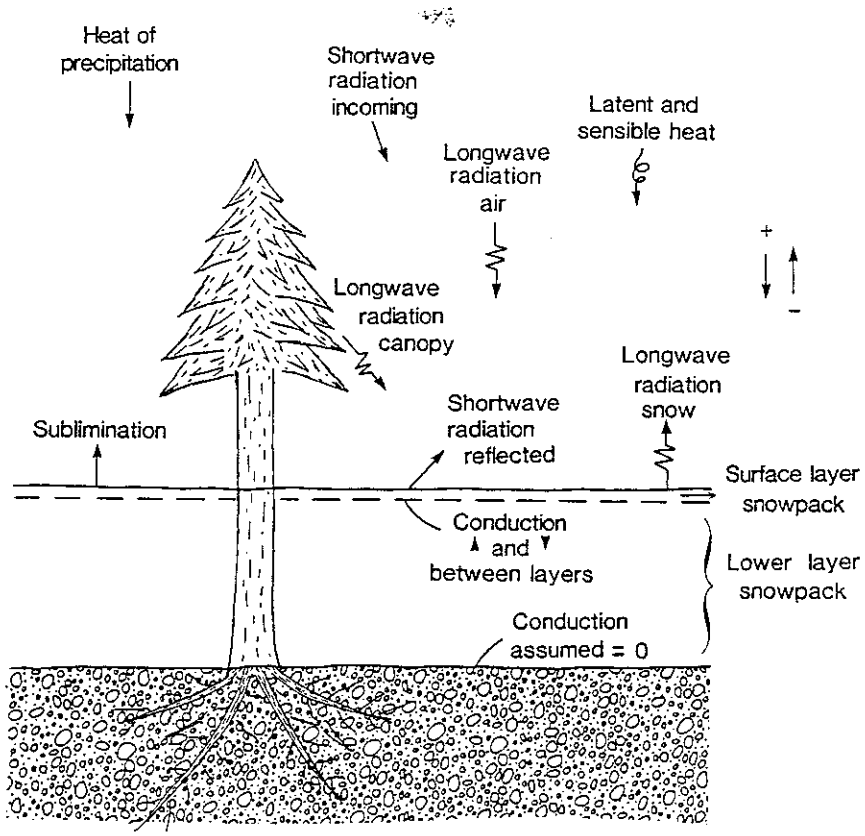


Fig. 9.3 Components of the snowpack energy balance.

The conduction of heat between the snow surface and the lower layer of the snowpack (QCOND) is computed by

$$QCOND = 2 \bullet DEN \bullet CS \bullet \sqrt{\frac{KEFF}{(DEN \times CS)} \bullet \frac{\Delta t}{\Pi}} \bullet (TS - PACT) \quad (9.19)$$

where

DEN is snowpack density (g/cm^3),
CS is the specific heat of ice in calories per gram per °C (cal/g/°C),
KEFF is the effective thermal conductivity of the snowpack (cal/s/cm/°C),
Δt is the time interval (s),
Π is pi, and
PACT is the temperature of the lower layer of the snowpack (°C).

KEFF is assumed equal to 0.0077 x DEN^2 (Anderson, 1968). DEN is computed daily using a procedure developed by Riley et al. (1973). PACT is recomputed each 12-hr period as a function of the snowpack water equivalent and the number of calories of heat needed to bring the lower layer to an isothermal state at 0 °C.

When Ts equals 0 °C, an energy balance at the air-snow interface is computed. For each 12-hr period the energy balance CALN in calories is computed by

$$CALN = SWN + LWN + CEN \tag{9.20}$$

where

SWN is net shortwave radiation (cal),
LWN is net longwave radiation (cal), and
CEN is an approximation of the convection-condensation energy terms (cal).

Daily precipitation modifies both the water and energy balances of the snowpack. The energy input from precipitation is computed as a function of the mean of the maximum and minimum daily air temperatures. Water and energy balance changes resulting from precipitation are made on a 24-hr basis and prior to the application of eq. (9.20).

SWN is computed by

$$SWN = SWRD \bullet (1.0 - ALB) \bullet TRNCF \tag{9.21}$$

where

SWRD is the computed shortwave radiation received on an HRU (cal),
ALB is the albedo of the snowpack surface, and
TRNCF is the transmission coefficient for the vegetation canopy over the snowpack.

ALB is computed as a function of the number of days since the last snowfall and whether the snowpack is in an accumulation or a melt phase (U.S. Army, 1956). A snowpack begins in an accumulation phase, but shifts to a melt phase when the temperature of the lower layer is isothermal at 0°C for 5 consecutive days or the Julian date exceeds a user defined forcing date. TRNCF is computed as a function of the winter cover density of the vegetation canopy over the snowpack. The relations between cover density and TRNCF were developed from relations presented by Miller (1959) and Vézina and Péch (1964).

LWN has two components, the longwave exchange between the air and the snowpack and the exchange between the vegetation canopy and the snowpack. LWN is computed by

$$LWN = (1.0 - COVDNW) \bullet [(EMIS \bullet AIRN) - SNON] + COVDNW \bullet (AIRN - SNON) \tag{9.22}$$

where

COVDNW is the winter cover density of the predominant vegetation above the snowpack,
EMIS is the emissivity of the air,
AIRN is the longwave energy emitted from a perfect black body at the average air temperature for the 12-hr period (cal), and
SNON is the longwave energy emitted from the snowpack surface at the temperature Ts for the 12-hr period (cal).

AIRN and SNON are computed using the Stefan-Boltzmann law. EMIS is a function of the moisture content of the air and ranges between 0.757 and 1.0 (U.S. Army, 1956). In the absence of humidity data, simplifying assumptions are made in the estimation of EMIS. During periods when frontal storms are dominant, EMIS is assumed to be 1.0 on days with precipitation and a user-defined parameter value for days without precipitation. During periods when convective storms are dominant, EMIS is computed separately for each 12-hr period as a function of precipitation occurrence and the ratio of observed to potential solar radiation.

The full equation for computing latent and sensible heat flux includes terms for temperature, vapor pressure, wind speed, and diffusivities of heat and vapor (U.S. Army, 1956). However, wind and vapor pressure or humidity data are generally not available. Therefore, the full equation is simplified to a temperature-index form to compute latent and sensible heat terms (CEN) by

$$CEN = CECN(MO) \bullet TAVC \tag{9.23}$$

where CECN is a parameter for month MO and TAVC is the mean air temperature for the 12-hr period (° C). CEN is computed only on days of rainfall or when the ratio of observed to potential solar radiation is less than or equal to 0.33. Areas of forest cover are assumed to receive only one-half of CEN computed by eq. (9.23).

When the energy balance CALN (eq. 9.20) is positive, there is a net gain of energy to the snowpack. This energy melts snow in the surface layer and the melt then transports heat into the snowpack by mass transfer. If the snowpack temperature is less than 0° C, the melt water is refrozen and

decreases the cold content of the snowpack. When the snowpack becomes isothermal at 0° C, snowmelt is first used to satisfy the freewater holding capacity of the snowpack. Any remaining melt leaves the bottom of the snowpack to become infiltration or surface runoff. When melt reduces the snowpack water equivalent below a user-defined threshold, the snowcovered area of an HRU is decreased using the areal-depletion-curve approach developed by Anderson (1973). Up to 10 different delpetion curves may be user-defined.

Evaporation and sublimation from the snow surface are assumed to occur only when there is no transpiration from vegetation above the snowpack. Loss from the snow surface is computed as a percentage of the daily PET value. The daily percentage is a user-defined parameter.

9.5. STORM-MODE COMPONENTS

A watershed is configured into flow-plane and channel segments for storm-mode computations. An HRU can be considered a single flow plane or it can be divided into 2 or more flow planes to account for variations in slope and surface roughness. Cascading flow planes are not supported, so all flow planes discharge into channel segments. The watershed drainage network is characterized as a system of channel, reservoir, and junction segments that jointly describe the drainage pattern. Each channel segment can receive upstream inflow from as many as three other segments. In addition, each channel segment can receive lateral inflow from as many as two flow planes (left bank and right bank).

9.5.1. Surface Runoff

Storm precipitation is reduced by interception using eqs. (9.6) and (9.7) and resulting net precipitation is available for infiltration. On pervious areas, infiltration is computed using a variation of the Green and Ampt equation (Green and Ampt, 1911). Point infiltration capacity (FR) is computed by

$$FR = KSAT \bullet \left(1.0 + \frac{PS}{SMS}\right) \tag{9.24}$$

where

KSAT is hydraulic conductivity of the transmission zone (in./hr),
PS is effective value of the product of capillary drive and moisture deficit (in.), and
SMS is the current value of accumulated infiltration (in.).

PS is varied linearly as a function of the ratio of the soil moisture quantity RECHR to REMX and user-defined parameters that represent the value of PS at field capacity (PSP) and the ratio of PSP at field capacity to PSP at wilting point. Point infiltration is converted to net infiltration over a flow plane using a procedure first presented by Crawford and Linsley (1966) which assumes that net infiltration varies linearly as a function of FR and rainfall rate.

Rainfall excess (net precipitation less net infiltration) is then routed as surface runoff over the flow planes into the channel segments using the kinematic-wave approximation to overland flow developed by Leclerc and Schaake (1973). The equation solved for each flow plane is

$$\frac{\partial h}{\partial t} + \frac{\partial q}{\partial x} = re \tag{9.25}$$

where

h is the depth of flow (ft),

q is the rate of flow per unit width ($ft^3/s/ft$),

re is the rate of rainfall excess inflow (ft/s),

t is time (s), and

x is the distance down plane (ft).

The relation between h and q is given as

$$q = \alpha h^m \tag{9.26}$$

where α and m are computed as functions of overland flow-plane characteristics using equations described by Dawdy et al. (1978).

9.5.2. Channel Flow

Channel flow is routed through the watershed channel system using the kinematic-wave approximation for channel flow described by Dawdy et al. (1978). The equation solved for each channel reach is

$$\frac{\partial A}{\partial t} + \frac{\partial Q}{\partial x} = q \tag{9.27}$$

where

A is the area of flow (ft^2),

Q is the flow rate (ft^3/s), and

q is the lateral inflow per unit length of channel ($ft^3/s/ft$).

The kinematic wave approximation relating Q to A is

$$Q = \alpha A^m \tag{9.28}$$

where α and m are computed from selected channel segment characteristics using equations described by Dawdy et al. (1978).

Routing through channel reservoirs can be computed using the same linear routing scheme or modified-Puls routing procedure described for the daily mode. Equations (9.17) and (9.18) are used but at shorter time steps.

9.5.3. Sediment

Sediment detachment and transport from flow planes is computed using a rill-interrill concept approach presented by Hjelmfelt et al. (1975) and is expressed as

$$\frac{\partial(ch)}{\partial t} + \frac{\partial(cq)}{\partial x} = ER + EF \tag{9.29}$$

where

c is sediment concentration in kilograms per ft^3 (Kg/ft^3),

h is flow depth (ft),

q is flow rate per unit width ($ft^3/s/ft$),

ER is the rainfall detachment rate of sediment ($Kg/ft^2/s$),

EF is the overland flow detachment rate of sediment ($Kg/ft^2/s$), and

x is distance down the flow plane (ft).

ER is assumed to occur in the interrill areas and is computed as a nonlinear function of rainfall rate and the mean depth of flow on the plane using a relationship described by Smith (1976). EF is assumed to occur in the rills and is computed as a linear function of the difference between

sediment transport capacity at the current flow depth and the current sediment transport rate using a relationship described by Hjelmfelt et al. (1975).

Sediment delivered from a flow plane is currently transported as a conservative substance in the channel system; detachment and deposition are not included. The channel sediment-routing component will be upgraded in the future with the addition of detachment and deposition capabilities. The continuity of mass equation used to describe sediment transport is expressed as

$$\frac{\partial(cA)}{\partial t} + \frac{\partial(cQ)}{\partial x} = TLAT \tag{9.30}$$

where the equation variables are as defined above and TLAT is the lateral inflow rate of sediment per unit length of channel expressed in Kg/s/ft

9.6. OPTIMIZATION AND SENSITIVITY ANALYSIS COMPONENTS

Optimization components control the automatic adjustment of model parameters to obtain better agreement between observed and predicted runoff. A model parameter is broadly defined as a value that is used to represent a physical or hydrologic characteristic of a watershed, and is held constant during a simulation run. This definition produces a large number of parameters for optimization and sensitivity analysis; however, the availability of a large number of parameters is not meant to encourage optimization of all of them. The parameters are available primarily for development work in the theory of errors for the modeling system and to permit model users to evaluate the sensitivity of each parameter, and its interactions with others.

Two optimization techniques are available: one is the Rosenbrock technique (Rosenbrock, 1960); the second is a Gauss-Newton technique similar to the linearization method described by Draper and Smith (1966), but with the incorporation of the Box-Kanemasu interpolation modification (Beck and Arnold, 1977). One of four objective functions can be selected for use with either technique.

Spatially distributed parameters have an initial value assigned to each HRU, subsurface reservoir, or ground-water reservoir. Temporally distributed parameters have an initial value assigned for each time increment. One or any combination of parameters can be selected for optimization. For each iteration of a distributed parameter, all values of the parameter are moved in the same direction at the same time. The amount that each value is moved can be selected as the same magnitude or as the same percentage of the initial value. A major assumption in this fitting

procedure is that the initial estimates of the values of a given distributed parameter are correct with regard to their relative differences in space or time.

An option in the fitting procedure allows the user to adjust different subsets of a distributed parameter independently. A special case of this option is that a distributed parameter may be adjusted independently on all HRU's.

Sensitivity-analysis components allow the user to determine the extent to which uncertainty in the parameters results in uncertainty in the predicted runoff. When sensitivity analysis is coupled with optimization, the user also can assess the magnitude of parameter standard errors and parameter intercorrelations. Discussions of sensitivity analysis and its interpretation are presented by Mein and Brown (1978) and Beck and Arnold (1977).

9.7. EXTENDED STREAMFLOW PREDICTION (ESP)

A modified version of the ESP program has been coupled to PRMS to provide forecasting capabilities. The ESP procedure uses historic or synthesized meteorologic data to forecast future streamflow, given the simulated hydrologic conditions for a watershed at a specified point in time. When historic data are used, the procedure assumes that past meteorological events are representative of future meteorological events. Alternative assumptions about future meteorological conditions can be made with the use of synthesized meteorological data.

To implement the ESP procedure, the user defines in PRMS a forecast period, the meteorological data set to be used, and the forecast variables of interest. The forecast period can vary from a few days to an entire water year. PRMS simulates to the start of the forecast period then stores all model variables that describe the state of the watershed system. A streamflow hydrograph for the forecast period then is simulated for each year in the meteorological data base. PRMS is reinitialized at the start of each iteration of the forecast period using the previously stored variables. The simulation results obtained from this iterative procedure are termed the conditional simulation.

Streamflow variables of interest are extracted from each forecast hydrograph and stored in an ESP file. The variables that can be extracted are the maximum and minimum daily flow values, flow volume, and the date that the flow decreases to less than a selected threshold value. One variable, or any combination of variables, can be selected for a given model run.

The ESP program reads the ESP file, performs a frequency analysis on the variables, and produces a probabilistic forecast for each of the variables in the file. The Log-Pearson Type III probability distribution is currently supported, but other distributions can be added. Frequency analysis results

for the conditional simulation are presented in tabular form and in a printer plot.

A frequency analysis also can be performed with ESP using the historic streamflow record or the simulation results obtained from running PRMS using the historic or synthesized meteorologic data without resetting the initial conditions each year. The results of this continuous simulation are termed the historic simulation when historic data are used. A comparison of frequency-analysis results between the measured data and the historic simulation provides a measure of the adequacy of the PRMS calibration. Differences between the analyses reflect the effects of model and data errors on the historic simulation, and they also provide a measure of the bias in the conditional-simulation results.

A continuous simulation using synthesized meteorologic data is termed the synthetic simulation. Differences between the analyses for the historic simulation and the synthetic simulation provide a measure of the effect of the assumed meteorological changes in the synthesized record on streamflow.

The ESP component provides a valuable tool for a variety of hydrologic forecasting and analysis applications. These applications include short-term or seasonal forecasting for flooding and water supply, evaluation of the effects of land-use changes on streamflow characteristics under a variety of meteorologic and watershed conditions, and the evaluation of the potential effects of climatic changes on streamflow under existing or future watershed conditions.

9.8. SYSTEM APPLICATIONS

The modeling system as discussed above has been applied in a variety of climatic and physiographic regions. Reports on snowmelt-runoff applications of PRMS include Leavesley and Striffler (1979), Leavesley et al. (1981), Brendecke and Sweeten (1985), and WMO (1986). A comparison of PRMS with the Streamflow Synthesis and Reservoir Regulation (SSARR) model was presented by Brendecke et al. (1985). A frozen soil algorithm was developed for PRMS by Emerson (1991) to simulate this cold–region process

PRMS was used in a number of studies to investigate the effects of surface coal mining on basin hydrologic response. A summary of these applications on approximately 50 basins in selected regions of the United States was presented by Stannard and Kuhn (1989). More recent applications of PRMS have been to address the potential effects of climate change on basin hydrologic response. Climate change related investigations in Colorado using PRMS coupled with selected atmospheric models were reported by Leavesley et al. (1992) and Hay et al. (1993). Application of selected storm-mode components, including sediment detachment and

transport, to tephra deposits from the eruption of Mount St. Helens, Washington, were reported by Leavesley et al. (1989). Other reports focusing on the sediment components of PRMS were presented by Carey and Simon (1984), Reed (1986), and Rankl (1987).

Model parameter estimation and evaluation techniques have been examined by several authors. Parameter estimation and calibration studies were reported for Colorado (Norris and Parker, 1985) and for Montana and Wyoming (Cary, 1991). The transferability of parameters to noncalibrated bains was examined by Kuhn and Parker (1992). A review and application of the parameter optimization and sensitivity analysis components was presented by Troutman (1985). Error analysis methodology was also examined by Rivera–Santos (1990).

A number of approaches have been developed to use GIS tools in the delineation of HRUs for PRMS applications. Leavesley and Stannard (1990a,b) presented a polygon HRU approach while Battaglin et al. (1993) presented a gridded HRU approach. Leavesley and Stannard (1990a) also coupled remotely sensed snowcovered–area data with the GIS tools to provide the ability to compare measured and simulated snowcovered area on selected dates. Flügel and Lüllwitz (1993) developed GIS procedures to compare modeling of micro– and mesoscale catchments in the United States and Germany.

9.9. THE NEXT GENERATION OF PRMS

One of the objectives in the development of PRMS was to provide a modular modeling system that could be used to address a variety of interdisciplinary environmental and water-resource problems. The modular concept enables the testing and development of a variety of modeling approaches that can incorporate knowledge from a broad range of scientific disciplines. Assessment of the initial PRMS modular structure identified a number of deficiencies for meeting these goals, so a new, more flexible, modular system was developed to provide a framework in which PRMS, and any other model or model component, can be incorporated.

The new system is called the Modular Modeling System (MMS) (Leavesley et al., 1992). MMS is an integrated system of Unix-based computer software that has been developed to provide the research and operational framework needed to support development, testing, and evaluation of physical-process algorithms and to facilitate integration of user-selected sets of algorithms into operational physical-process models. MMS uses a master library that contains compatible modules for simulating a variety of water, energy, and biogeochemical processes. A model is created by selectively coupling the most appropriate process algorithms from the library to create an "optimal" model for the desired application.

Where existing algorithms are not appropriate, new algorithms can be developed and incorporated in the library.

A geographic information system (GIS) interface has been developed for MMS to facilitate model development, application, and analysis. This interface permits application of a variety of GIS tools to characterize the topographic, hydrologic, and biologic features of a physical system for use in a variety of lumped- and distributed-parameter modeling approaches. MMS display capabilities permit visualization of the spatial distribution of model parameters and of the spatial and temporal variation of simulated state variables during a model run.

MMS provides a common framework in which to focus multidisciplinary research and operational efforts to provide improved understanding of complex water, energy, and biogeochemical processes. All the components of PRMS and ESP described above have been incorporated in MMS and the MMS modular library. Additional modules from a variety of other hydrologic and ecosystem models are currently being developed for incorporation into the modular library.

REFERENCES

Anderson, E.A., 1968, Development and testing of snow pack energy balance equations: Water Resour. Res., v. 4, no. 1, p. 19–38.

Anderson, E.A., 1973, National Weather Service river forecast system—Snow accumulation and ablation model: NOAA Tech. Memorandum NWS Hydro–17, U.S. Dept. of Commerce, Silver Spring, Maryland.

Barry, R.G., 1972, Climatic environment on the east slope of the Colorado front range: Boulder, University of Colorado, Institute of Arctic and Alpine Research Occasional Paper 3, 206 p.

Battaglin, W.A., Hay, L.E., Parker, R.S., and Leavesley, G.H., 1993, Applications of GIS for modeling the sensitivity of water resources to alterations in climate in the Gunnison River basin, Colorado: Water Res. Bull. v. 25, no. 6, p. 1021–1028.

Beck, J.V. and Arnold, K.J., 1977, Parameter estimation in engineering and science: New York, John Wiley and Sons, Inc., 340-349.

Brendecke, C.M., Laiho, D.R., and Holden, D.C., 1985, Comparison of two daily streamflow simulation models of an alpine watershed: J. Hydrol., v. 77, p. 171–186.

Brendecke, C.M., and Sweeten, J.G., 1985, A simulation model of Boulder's alpine water supply, in Proceedings of 53rd Annual Meeting of the Western Snow Conference: Boulder, Colorado, pp. 63-71.

Carey, W.P., and Simon, A., 1984, Physical basis and potential estimation techniques for soil erosion parameters in the Precipitation–Runoff Modeling System (PRMS): U.S. Geol. Surv. Water–Resour. Invest. Rep. 84–4218, 32 p.

Cary, L.E., 1991, Techniques for estimating selected parameters of the U.S. Geological Survey's Precipitation–Runoff Modeling System in eastern Montana and northeastern Wyoming: U.S. Geol. Surv. Water–Resour. Invest. Rep. 91–4068, 39 p.

Crawford, N.H., and Linsley, R.K., 1966, Digital simulation in hydrology: Stanford, Calif., Stanford Watershed Model IV, Tech. Rept. 39, Civil Eng. Dept., Stanford Univ., 210 p.

Dawdy, D.R., Schaake, J.C., Jr., and Alley, W.M.,1978, Distributed routing rainfall-runoff model: U.S. Geol. Surv. Water Resour. Invest. Rep. 78-90.

Day, G.N., 1985, Extended streamflow forecasting using NWSRFS: J. Water Resour. Plan. and Manag. Am. Soc. Civ. Eng., 111(2), 157-170.

Dickinson, W.T., and Whiteley, H.Q., 1970, Watershed areas contributing to runoff: IAHS Publ. no. 96, 1.12-1.28.

Draper, N.R., and Smith, H., 1966, Applied regression analysis: New York, John Wiley and Sons, Inc.

Emerson, D.G., 1991, Documentation of a heat and water transfer model for seasonally frozen soils with application to a precipitation–runoff model: U.S. Geol. Surv. Open–File Report 91–462, 97 p.

Federer, Anthony C., and Lash, Douglas, 1978, Brook—A hydrologic simulation model for eastern forests: Durham, New Hampshire, University of New Hampshire, Water Resources Research Center, Research Report No. 19, 84 p.

Flügel, W.A., and Lüllwitz, Th., 1993, Using a distributed hydrologic model with the aid of GIS for comparative hydrological modelling of micro– and mesoscale catchments in the USA and Germany, in Wilkinson, W.B., ed., Macroscale modelling of the hydrosphere: IAHS Pub. no. 214, p. 59–66.

Frank, E.C., and Lee, Richard, 1966, Potential solar beam irradiation on slopes: U.S. Department of Agriculture, Forest Service Research Paper RM–18, 116 p.

Green, W.H., and Ampt, G.A., 1911, Studies on soil physics, I-- Flow of air and water through soils: J. Agric. Res., 4, 1-24.

Hamon, W.R., 1961, Estimating potential evapotranspiration: J. Hydraul. Div. Am. Soc. Civ. Eng., 87(HY3), 107-120.

Hay, L.E., Battaglin, W.A., Parker, R.S., and Leavesley, G.H., 1993, Modeling the effects of climate change on water resources in the Gunnison River basin, Colorado, in Goodchild, M.F., Parks, B.O., and Steyaert, L.T., eds., Environmental modeling with GIS: Oxford Univ. Press, p. 173–181.

Hewlett, J.D., and Nutter, W.L., 1970, The varying source area of streamflow from upland basins, in Proceedings of Symposium on Interdisciplinary Aspects of Watershed Management: Montana State University, Bozeman, Montana, pp. 65-83.

Hjelmfelt, A.T., Piest, R.P., and Saxton, K.E., 1975, Mathematical modeling of erosion on upland areas in Proceedings of Congress of the 16th Int. Assoc. for Hydraul. Res.: Sao Paulo, Brazil, 2, pp. 40-47.

Jensen, M.E., Robb, D.C.N., and Franzoy, C.E., 1969, Scheduling irrigation using climate-crop-soil data, in Proceedings of the American Society of Engineers National Conference on Water Resoures Engineering: New Orleans, Louisiana, 20 pp.

Kuhn, G., and Parker, R.S., 1992, Transfer of watershed–model parameter values to noncalibrated basins in the Gunnison River basin, Colorado: AWRA 28th Annual Conference and Symposium, Managing water resources during global change, Reno, Nevada, November 1–5, 1992, p. 741–750.

Leaf, C.F., and Brink, G.E., 1973, Hydrologic simulation model of Colorado subalpine forest: U.S. Department of Agriculture, Forest Service Research Paper RM–107, 23 p.

Leavesley, G.H., Branson, M.D., and Hay, L.E., 1992, Using coupled atmospheric and hydrologic models to investigate the effects of climate change in mountainous regions: AWRA 28th Annual Conference and Symposium, Managing water resources during global change, Reno, Nevada, November 1–5, 1992, p. 691–700.

Leavesley, G.H., Lusby, G.C., and Lichty, R.W., 1989, Infiltration and erosion characteristics of selected tephra deposits from the 1980 eruption of Mount St. Helens, Washington: Hydrol. Sci. J., 34(3).

Leavesley, G.H., Lichty, R.W., Troutman, B.M., and Saindon, L.G., 1981, A precipitation-runoff modeling system for evaluating the hydrologic impacts of energy-resource development in Proceedings of 49th Annual Meeting of the Western Snow Conference: St. George, Utah, pp. 65-76.

Leavesley, G.H., Lichty, R.W., Troutman, B.M., and Saindon, L.G., 1983, Precipitation-runoff modeling system--User's manual: U.S. Geol. Surv. Water Resour. Invest. Rep. 83-4238.

Leavesley, G.H., Restrepo, P., Stannard, L.G., and Dixon, M., 1992, The modular hydrologic modeling system – MHMS: AWRA 18th Annual Conference and Symposium, Managing water resources during global change, Reno, Nevada, November 1–5, 1992, p. 263–264.

Leavesley, G.H., and Stannard, L.G., 1990a, Application of remotely sensed data in a distributed– parameter watershed model, in Kite, G.W., and Wankiewicz, eds., Proceedings of Workshop on Applications of Remote Sensing in Hydrology: National Hydrologic Research Centre, Environment Canada, p. 47–64.

Leavesley, G.H., and Stannard, L.G., 1990b, A modular watershed–modeling system for use in mountainous regions: Schweizer Ingenieur und Architekt, n. 18, p. 380–383.

Leavesley, G.H., and Striffler, W.D., 1979, A mountain watershed simualtion model, in Colbeck, S.C., and Ray, M., eds., Modeling of snow cover runoff: Hanover, New Hampshire, U.S. Army CRREL, 379-386.

Leclerc, G., and Schaake, J.C., Jr., Methodology for assessing the potential impact of urban development on urban runoff and the relative efficiency of runoff control alternatives, Mass. Instit. of Tech. Ralph M. Parsons Laboratory Report 167, 1973.

Lumb, A.M., Kittle, J.L., Jr., and Flynn, K.M., 1990, User's manual for ANNIE, a computer program for interactive hydrologic analyses and data management: U.S. Geol. Surv. Water Resour. Invest. Rep. 89–4080, 236 p.

Mein, R.G. and Brown, B.M., 1978, Sensitivity of optimized parameters in watershed models: Water Resour. Res. 14(2), 299-303.

Miller, D.H., 1959, Transmission of insolation through pine forest canopy as it effects the melting of snow: Mitteilungen der Schweizerischen Anstalt für das forstliche Versuchswesen, Versuchsw. Mitt., v. 35, p. 35–79.

Murray, F.W., 1967, On the computation of saturation vapor pressure: J. Applied Meteorology, v. 6, p. 203–204.

Obled, Ch. and Rosse, B., 1977, Mathematical models of a melting snowpack at an index plot: J. Hydrol. 32, 139-163.

Norris, J.M., and Parker, R.S., 1985, Calibration procedure for a daily flow model of small watersheds with snowmelt runoff in the Green River coal region of Colorado: U.S. Geol. Surv. Water–Resour. Invest. Rep. 83–4263, 32 p.

Rankl, J.G., 1987, Analysis of sediment production from two small semi arid basins in Wyoming: U.S. Geol. Surv. Water–Resour. Invest. Rep. 85–4314, 27 p.

Reed, L.A., 1986, Verification of the PRMS sediment–discharge model, in Proceedings of the Fourth Federal Interagency Sedimentation Conference: March 24–26, 1986, Las Vegas, v. II, p. 6–44 to 6–54.

Riley, J.P., Israelsen, E.K., and Eggleston, K.O., 1973, Some approaches to snowmelt prediction, in The role of snow and ice in hydrology: IAHS Pub. 107, p. 956–971.

Rivera–Santos, J., 1990, Parameter estimation in conceptual precipitation–runoff models with emphasis on error analysis, in Tropical Hydrology and Caribbean Water Resources: Proceedings of the International Symposium on Tropical Hydrology and Fourth Caribbean Islands Water Resources Congress, San Juan, Puerto Rico, American Water Resources Association, p. 91–100.

Rosenbrock, H.H., 1960, An automatic method of finding the greatest or least value of a function: Computer Journal, 3, 175-184.

Smith, R.E., 1976, Field test of a distributed watershed erosion/sedimentation model, in Soil erosion—Prediction and control: Ankeny, Iowa, Soil Conservation Society of America, p. 201–209.

Stannard, L.G., and Kuhn, G., in press, Watershed modeling, in Britton, L.J., Anderson, C.L., Goolsby, D.A., and Van Haveren, B. P., eds., Summary of the U.S. Geological Survey and U.S. Bureau of Land Management National Coal-Hydrology Program, 1974-84: U.S. Geol. Surv. Prof. Pap. 1464.

Swift, L.W., 1976, Algorithm for solar radiation on mountain slopes: Water Resour. Res. 12 (1), 108-112.

Tangborn, W.V., 1979, A model to predict short–term snowmelt runoff using synoptic observations of streamflow, temperature, and precipitation in Colbeck, S.C., and Ray, M., eds., Modeling of snow cover runoff: Hanover, New Hampshire, U.S. Army, CRREL, p. 414–426.

Thompson, E.S., 1976, Computation of solar radiation from sky cover: Water Resour. Res., v. 12, no. 5, p. 859–865.

Troutman, B.M., 1985, Errors and parameter estimation in precipitation-runoff modeling: Water Resour. Res., 21 (8), 1214-1222.

U.S. Army, 1956, Snow hydrology, U.S. Army Corps. of Engineers, Portland, Oregon, USA.

U.S. Soil Conservation Service, 1971, SCS National Engineering Handbook, Section 4--Hydrology, Washington, D.C., U.S. Government Printing Office.

Vézina, P.E., and Péch, G.Y., 1964, Solar radiation beneath conifer canopies in relation to crown closure: Forest Science v. 10, no. 4, p. 443–451.

Willen, D.W., Shumway, C.A., and Reid, J.E., 1971, Simulation of daily snow water equivalent and melt, in Proceedings of 39th Annual Meeting of the Western Snow Conference: Billings, Montana, pp. 1-8.

WMO, 1986, Intercomparison of models of snowmelt runoff: WMO Operational Hydrol. Rep. no. 23, WMO Publ. no. 646, WMO, Geneva.

Zahner, R., 1967, Refinement in empirical functions for realistic soil–moisture regimes under forest cover, in Sopper, W.E., and Lull, H.W., eds., International symposium of forest hyrology: New York, Pergamon Press, p. 261–274.