# Evaluating GCM Land Surface Hydrology Parameterizations by Computing River Discharges Using a Runoff Routing Model: Application to the Mississippi Basin

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ABSTRACT

To relate general circulation model (GCM) hydrologic output to readily available river hydrographic data, a runoff routing scheme that routes gridded runoffs through regional- or continental-scale river drainage basins is developed. By following the basin overland flow paths, the routing model generates river discharge hydrographs that can be compared to observed river discharges, thus allowing an analysis of the GCM representation of monthly, seasonal, and annual water balances over large regions. The runoff routing model consists of two linear reservoirs, a surface reservoir and a groundwater reservoir, which store and transport water. The water transport mechanisms operating within these two reservoirs are differentiated by their time scales; the groundwater reservoir transports water much more slowly than the surface reservoir. The groundwater reservoir feeds the corresponding surface store, and the surface stores are connected via the river network.

The routing model is implemented over the GEWEX (Global Energy and Water Cycle Experiment) Continental-Scale International Project Mississippi River basin on a rectangular grid of 2° × 2.5°. Two land surface hydrology parameterizations provide the gridded runoff data required to run the runoff routing scheme: the variable infiltration capacity model, and the soil moisture component of the simple biosphere model. These parameterizations are driven with 4° × 5° gridded climatological potential evapotranspiration and 1979 First GARP (Global Atmospheric Research Program) Global Experiment precipitation. These investigations have quantified the importance of physically realistic soil moisture holding capacities, evaporation parameters, and runoff mechanisms in land surface hydrology formulations.

## 1. Introduction

Atmospheric general circulation models (GCMs) have become a common tool for studying large-scale hydrological and hydroclimatological problems. Typically, GCMs include a land-surface hydrology parameterization that governs the land surface–atmosphere interaction of physical processes such as precipitation, evapotranspiration, infiltration, and runoff. These hydrology parameterizations range in complexity from the simple soil slab model of Budyko (see Manabe 1969) to the biospheric models such as the biosphere–atmosphere transfer scheme (BATS) by Dickinson et al. (1986), the simple biosphere model (SiB) of Sellers et al. (1986), and the bare essentials of surface transfer (BEST) model of Pitman (1988).

A primary goal of these parameterizations is to produce realistic surface water balances when driven by realistic near-surface atmospheric forcing: precipitation, temperature, winds, humidity, and radiation. Unfortunately, the data and schemes available to validate GCM hydrologic descriptions of precipitation, evapotranspiration, soil moisture storage, and runoff are lacking. In contrast, river runoff data are readily available. This river runoff is an important integrator of the hydrologic cycle and is measured more accurately than the aforementioned hydrologic components. To relate GCM hydrologic output to river hydrographs, we have developed a runoff routing model that routes GCM-computed runoff through regional- or continental-scale river drainage networks. The modeled river runoff can then be compared with observed river discharges, thus providing a measure of monthly, seasonal, and annual water balances over large regions. GCM-computed runoff, after it leaves a grid box, is of no consequence for a typical GCM and basically does not

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interact with other components of the GCM. The production of realistic river hydrographs allows an important diagnostic verification of the grid box balances of precipitation, evaporation, soil moisture, and runoff time series. Thus, a runoff routing model can be a valuable tool for identifying weaknesses and suggesting modifications to GCM land surface hydrology parameterizations. Since GCMs typically use grid box areas of over 100 000 km$^2$ (4° latitude × 5° longitude), the transport model must be of regional or continental scale, and the modeled river basins will typically be large. Examples include the Mississippi and Amazon basins.

The transport or routing model is based on a rectangular grid of 2° latitude × 2.5° longitude, of finer resolution than a representative 4° × 5° GCM grid. It is through this finer grid that the GCM-gridded runoff output is routed. The model is composed of a coupled system of ordinary differential equations, where each equation represents a fine-grid box that considers such water balance features as inflow from adjacent upstream box(es), runoff input from the coincident GCM grid box, and downstream discharge. In addition, each of these equations requires information related to the water residence time within a grid box, such as soil type, slope, and stream length. In this paper, two land surface hydrology parameterizations are used to provide the gridded runoff data required to run the runoff routing model. These parameterizations are driven by gridded climatological values of potential evapotranspiration and climatologically scaled precipitation.

## 2. Runoff routing model

The routing model assumes that the GCM hydrology parameterization outputs two runoff values, a direct, surface storm runoff and a groundwater, subsurface discharge. Accordingly, it is assumed that the routing model consists of a surface reservoir corresponding to the storm runoff and a groundwater reservoir corresponding to groundwater base-flow discharge. Due to the different water transport mechanisms operating within these two reservoirs, they are differentiated by their time scales; the groundwater reservoir transports water much more slowly than the surface reservoir. In addition, we assume that the groundwater reservoir feeds the corresponding surface store, and the surface stores are connected via the river network.

Applying conservation of mass principles to a routing model grid box yields the following continuity equations:

$$\frac{dS_s}{dt} = Q_{si} + Q_{sr} + Q_g - Q_s \qquad (1a)$$

$$\frac{dS_g}{dt} = Q_{gr} - Q_g, \qquad (1b)$$

where $S_s$ is the surface storage; $S_g$ is the groundwater storage; $Q_s$ is the surface store flow; $Q_g$ is the groundwater store flow; $Q_{sr}$ is the grid-box surface runoff (from the GCM gridded surface runoff); $Q_{gr}$ is the grid-box groundwater runoff (from the GCM gridded groundwater runoff); $Q_{si}$ is the surface store inflow from adjacent grid cells; and $t$ is time.

To solve these equations the relationship between the storage and outflow must be defined. While nonlinear relationships between storage and flow have been developed (Singh 1988), their use is not justified for the coarse spatial and temporal resolutions considered in this model. Thus, assume that these two storage reservoirs are linear—that is, storage is proportional to outflow (Overton and Meadows 1976):

$$S(t) = kQ(t), \qquad (2)$$

where $k$ is a constant with dimension of time, equal to the typical residence or transit time of a fluid element passing through the reservoir. The parameter $k$ is a function of such things as travel distance (which is a function of grid size), streambed slope, streambed roughness, and stream length, width, and depth. After substitution, the original set of equations becomes

$$k_s \frac{dQ_s}{dt} = Q_{si} + Q_{sr} + Q_g - Q_s \quad \text{and} \qquad (3a)$$

$$k_g \frac{dQ_g}{dt} = Q_{gr} - Q_g. \qquad (3b)$$

This set of equations, when applied to each grid box of the runoff routing model, is connected via the river network through the presence of the $Q_{si}$ term. To illustrate the two-dimensional character of the contributing flow network, expand $Q_{si}$ to yield

$$Q_{si} = Q_{si\mathrm{N}} + Q_{si\mathrm{NE}} + Q_{si\mathrm{E}} + Q_{si\mathrm{SE}} + Q_{si\mathrm{S}} + Q_{si\mathrm{SW}} + Q_{si\mathrm{W}} + Q_{si\mathrm{NW}}, \qquad (4)$$

where the subscripts N, NE, E, SE, S, SW, W, and NW indicate the compass direction of the adjacent connecting grid box. One of the right-hand-side terms will be zero (the one corresponding to the outflow boundary), and possibly all eight will be zero (for the case of a grid box located at the head of a watershed), depending on the gridded representation of the river network. Figure 1 provides a schematic illustrating the relationship between the runoff routing model components of overland flow paths, grid-box groundwater and surface flows, inflow from upstream grid boxes, and grid-box discharge.

With the assumptions and approximations identified above, the model reduces to a coupled system of ordinary differential equations whose solution produces a runoff hydrograph for each routing model grid box. The governing equations can be solved by noting that (3) takes the general form

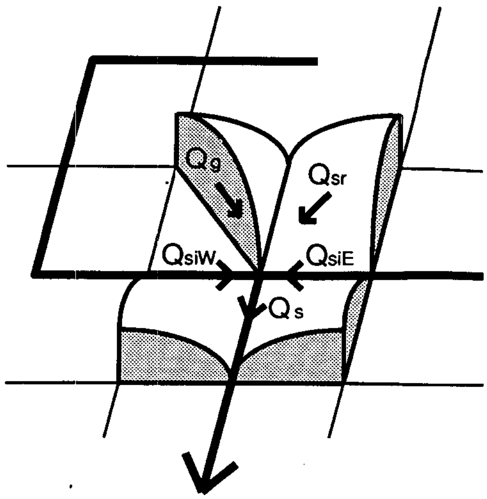


FIG. 1. Schematic showing runoff routing model grid, overland flow paths, grid-box groundwater and surface flows, inflow from upstream grid boxes, and grid-box surface outflow or discharge.

$$\frac{dQ}{dt} + \frac{1}{k} Q(t) = \frac{1}{k} I(t), \tag{5}$$

where $I(t)$ represents the sum of all inflow terms. This equation has the integrating factor

$$\exp\left(\int \frac{1}{k} dt\right) = \exp\left(\frac{t}{k}\right). \tag{6}$$

Multiplying both sides of (5) by the right-hand side of (6) yields

$$\exp\left(\frac{t}{k}\right)\frac{dQ}{dt} + \frac{1}{k}\exp\left(\frac{t}{k}\right)Q(t) = \frac{1}{k}\exp\left(\frac{t}{k}\right)I(t), \tag{7}$$

and since

$$\frac{1}{k}\exp\left(\frac{t}{k}\right)Q(t) = Q(t)\frac{d}{dt}\left[\exp\left(\frac{t}{k}\right)\right], \tag{8}$$

(7) can be written as

$$\frac{d}{dt}\left[Q\exp\left(\frac{t}{k}\right)\right] = \frac{1}{k} I(t)\exp\left(\frac{t}{k}\right). \tag{9}$$

Thus,

$$Q(t)\exp\left(\frac{t}{k}\right) = \frac{1}{k}\int I(t)\exp\left(\frac{t}{k}\right)dt + C, \tag{10}$$

where $C$ is an integration constant determined from the initial conditions. After dividing through by the exponent term, the final solution is

$$Q(t) = \exp\left(-\frac{t}{k}\right)\frac{1}{k}\int I(t)\exp\left(\frac{t}{k}\right)dt + C\exp\left(-\frac{t}{k}\right). \tag{11}$$

For the relatively complex time variation of GCM-computed runoff, the solution of (11) requires a numerical computation of the integral and, for this application, a tracking of the river network from its headwaters to the mouth. An alternative approach is to solve the system of equations numerically in their ordinary differential equation form [(3a) and (3b)]. These model equations typically involve steady-state terms that do not grow significantly with time, together with, depending on the magnitude of $k$, rapidly decaying transient terms. The steady-state terms typically result from groundwater flow, while the transient terms are due to the surface flow. The presence of significantly different time constants in the equations leads to a class of problems called "stiff systems" of differential equations. In such problems it is critical that the numerical solution be able to resolve the steady-state portion without becoming dominated by errors encountered in resolving the transient part. While this problem can be overcome by a reduction of the time step, frequently the time step must be made so small that round-off errors may dominate the solution, and the computational expense becomes exorbitant. The scheme presented by Gear (1971) is implemented to handle the "stiffness" problem. Further discussion of the characteristics of stiff differential equations is provided by Sampine and Gear (1979).

## 3. Computational procedures

To implement the water routing model for an actual drainage basin, the watershed river network must be specified, the time constants $k$ must be determined, initial conditions provided, and runoffs computed using a GCM land surface hydrology parameterization. These topics are discussed in the following paragraphs.

### a. Routing considerations

The watershed of interest is defined and described by a grid and flow network. The routing model was implemented over the GEWEX (Global Energy and Water Cycle Experiment) Continental-Scale International Project (GCIP) Mississippi River basin on a rectangular grid of 2° × 2.5°. Figure 2 illustrates the model approximation of the basin river network, along with the representation of the Mississippi basin boundary and the locations of routing model grid boxes, which correspond to the river discharge data collection stations located at the mouths of the Missouri, Ohio, and Arkansas subbasins of the Mississippi. Since the GCM grid box that contains the mouth of the Mississippi River is a part of the Gulf of Mexico in the Goddard Laboratory for Atmospheres (GLA) GCM, it is not included in the watershed. The area comprising the routing model representation of the entire Mississippi basin is 99.3% of the United States Geological Survey (USGS) value (van der Leeden 1975). The modeled

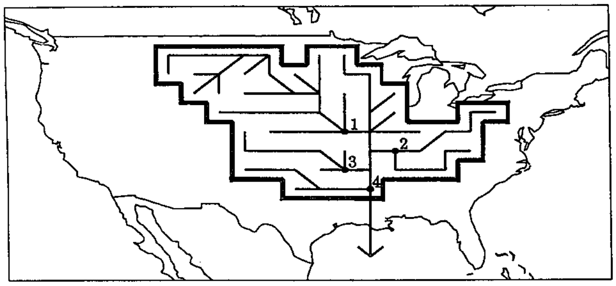


FIG. 2. A 2° × 2.5° runoff routing model depiction of the Mississippi basin river network and locations of discharge stations at the mouths of the 1) Missouri, 2) Ohio, 3) Arkansas, and 4) Mississippi Rivers.

area depictions of individual watersheds within the Mississippi basin are roughly 98.7%, 95.2%, 102.1%, and 103.5%, of the USGS values for the Missouri, Arkansas, Ohio, and Mississippi above the Missouri, respectively. These basins cover the equivalent area of 7.5, 2.0, 2.75, and 2.5 GCM grid boxes, respectively. The total modeled Mississippi basin covers the area of 17.0 GCM grid boxes; the remaining 2.25 boxes are contained along the Mississippi River below the Missouri and within the Red River basin.

To implement the model, the river network must be indexed to identify the relationship between the GCM grid, the routing model grid, and the connectivity of the grid boxes via the river network. There are 112 routing model grid boxes contained within the 4 × 7 rectangular cluster of GCM grid boxes that cover the Mississippi basin. Sixty-eight of these routing model boxes are contained within the drainage basin.

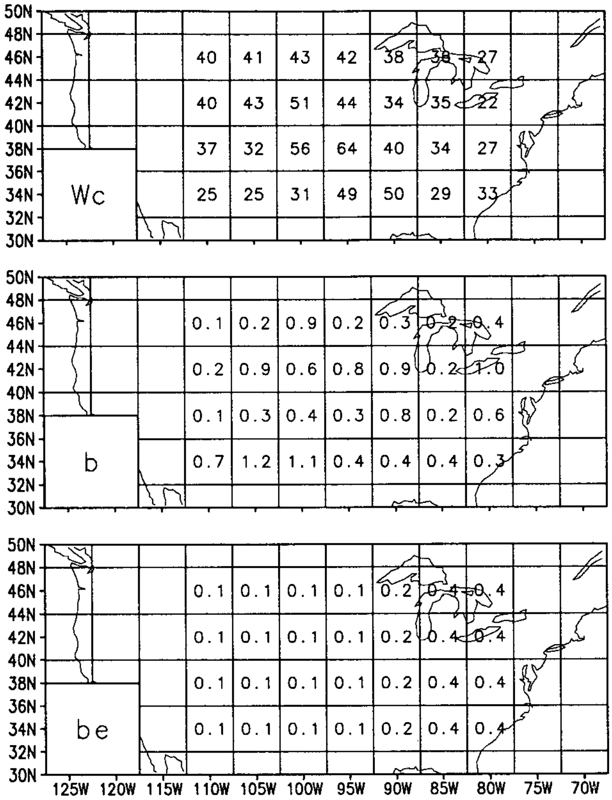


FIG. 3. Spatial distribution of 4° × 5° gridded soil moisture storage capacity $W_c$ (top), the infiltration parameter $b$ (middle), and the evaporation parameter $b_e$ (bottom), used in the simulation using modified VIC parameters.

### *b. Time-scale considerations*

Grid-specific values for the time constants $k$ must be defined. Askew (1970) presented empirical formulas for computing lag times based on observations of storm-produced runoff from five catchment systems. Askew concluded that the lag time was correlated with stream length $L$, overland slope $S_o$, and mean discharge $Q_m$, leading to a general lag–discharge relationship of the form

$$k_s = \alpha L^{\beta} S_o^{-\lambda} Q_m^{-\eta}, \tag{12}$$

where $\alpha$, $\beta$, $\lambda$, and $\eta$ were found to be 0.88, 0.80, 0.33, and 0.23 respectively; $k_s$, $L$, and $Q_m$ have dimensions of hours, kilometers, and cubic meters per second, respectively, and $S_o$ is dimensionless. This formulation supports the premise that a dominant factor affecting residence time is the fluid travel distance. The small negative exponents used for the terms $S_o$ and $Q_m$ are also consistent with many other theoretical and empirical studies considering watershed response time characteristics (Singh 1988). Overland slopes represented by $S_o$, are computed from 10-min gridded topographic data and averaged over the routing model grid boxes. Discharges $Q_m$ are obtained from van der Leeden (1975) and extrapolated upstream as required. Gridded stream lengths $L$ are computed based on latitude-dependent model grid dimensions and the configuration of the overland flow network defined in Fig. 2. These lengths are scaled by a factor of 1.25 to account for deviations from a straight path (Leopold et al.

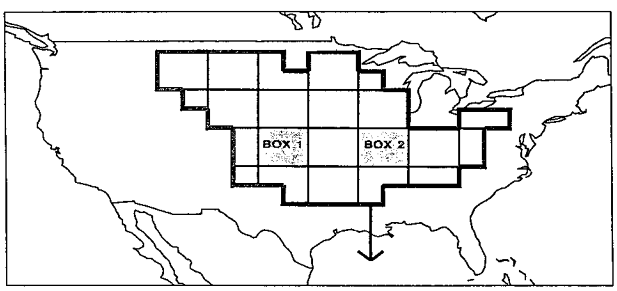


FIG. 4. Locations of the 4° × 5° land surface hydrology parameterization grid boxes labeled as "box 1" and "box 2" in Figs. 5, 6, 7, and 8.

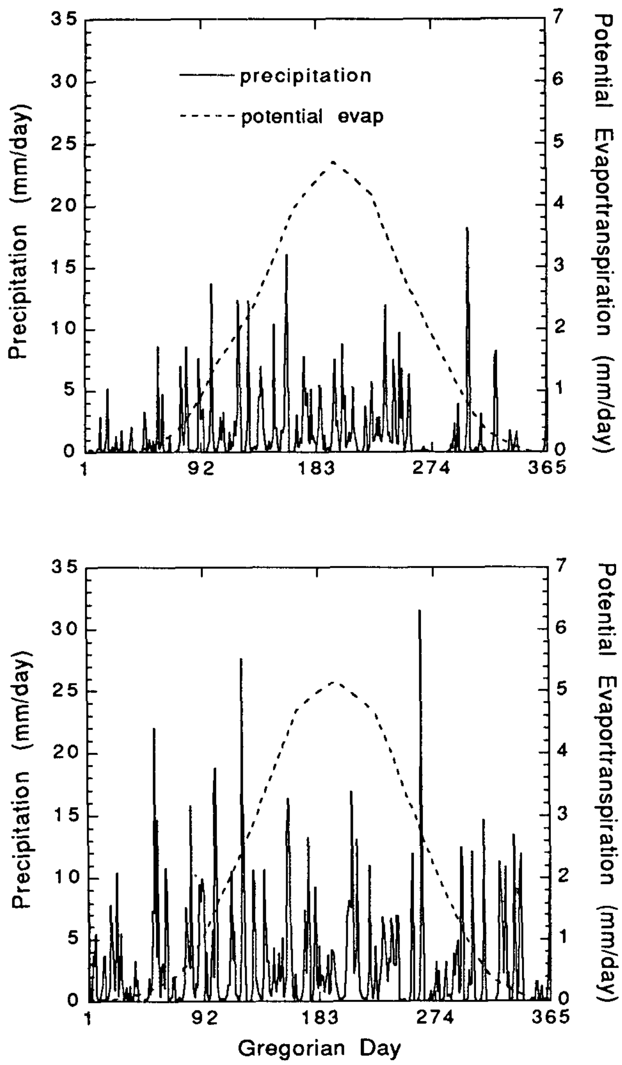


FIG. 5. Atmospheric precipitation and potential evapotranspiration forcing for box 1 (top) and box 2 (bottom) identified in Fig. 4.

1964). Computed values for $k_s$ within the Mississippi basin range from 3 to 7 days. These are consistent with continental-scale watershed transfer coefficients computed by Vorosmarty et al. (1989), who arrived at $k_s$ residence times ranging from 0.6 to 2.4 days for a $1/2°$ latitude × $1/2°$ longitude grid, from consideration of mean annual velocities, channel and grid dimensions, and river sinuosity. For the groundwater reservoir, $k_g$ is defined to be 30 days.

### c. Initial discharge determination

Initial discharge conditions must be supplied for the groundwater and surface stores. In this study the land surface parameterizations are driven with climatological atmospheric forcing, and the model outputs are compared with climatological discharges. Thus, to establish satisfactory initial conditions, the model is run iteratively over a few annual cycles while substituting the end-of-year discharges, or zero for the first iteration, for the initial conditions. This procedure is continued until the solution converges, typically after three years.

### d. Gridded surface and groundwater runoff

The model requires input of both storm runoff and subsurface drainage (groundwater) base flow, such as that available from a GCM land surface hydrology parameterization. Gridded runoff inputs have been computed using two different GCM land surface hydrology

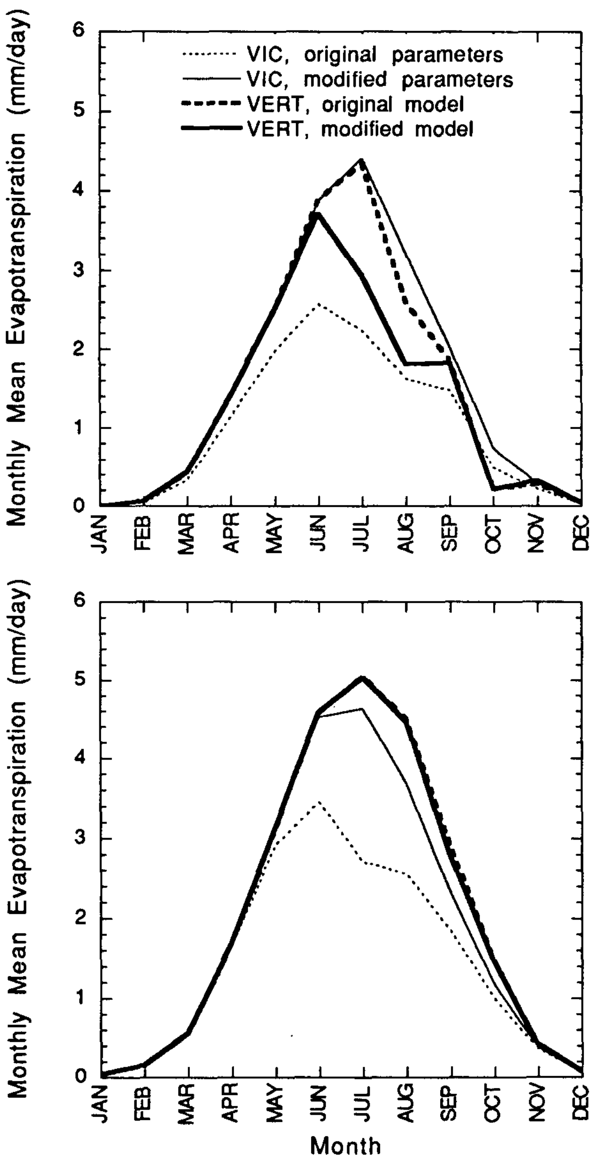


FIG. 6. Monthly mean evapotranspiration in box 1 (top) and box 2 (bottom) from the original and modified VIC parameters, and the VERT models.

parameterizations, which are driven by climatological values of potential evapotranspiration and precipitation. These models fall within the middle to high range of complexity currently used in GCMs. The first parameterization contains a detailed vertical accounting of moisture transfer between multiple soil layers and vegetation-controlled evapotranspiration. This parameterization assumes that there is no spatial variability across individual GCM grid boxes. The second model does not contain the complex vertical structure, but does include subgrid variability of infiltration capacity within each GCM grid box.

## 4. Parameterizations analyzed

### *a. Simple biosphere model*

The first parameterization follows the three-layer soil module used in the SiB model of Sellers et al. (1986). The land area contained within each GCM grid box is considered horizontally uniform, but the soil–vegetation types can vary from grid box to grid box. Since the key characteristic of this parameterization is the relatively detailed vertical accounting of soil moisture transport, it will be referred to as VERT. The general flux-gradient structure of this multilayer soil model is similar to several other state-of-the-art land surface hydrology parameterizations used in GCMs, including BATS (Dickinson et al. 1986), the Goddard Institute for Space Studies model (Abramopoulos et al. 1988), and the Canadian land surface scheme (CLASS) (Verseghy 1991). The moisture exchange between soil layers is determined by the steady-state, unsaturated, one-dimensional solution of Darcy's law (Freeze and Cherry 1979),

$$Q_f = -K\left(\frac{d\psi}{dz} + 1\right), \qquad (13)$$

where $Q_f$ is the downward flux of water, $K$ is the hydraulic conductivity of the soil, and $\psi$ is the soil moisture potential. Computation of the hydraulic diffusion and gravitational drainage of water within the three soil layers is computed based on Clapp and Hornberger (1978) and Milly and Eagleson (1982).

Runoff occurs via the moisture draining out of the lowest layer and is assumed to be driven by gravity only, with no diffusive transport occurring. Thus,

$$Q_3 = -K_3 \sin(\theta), \qquad (14)$$

where $\theta$ is the mean slope angle, the subscript 3 indicates the third and lowest soil layer, and $K_3$ is the hydraulic conductivity of the third soil layer. The conductivity is given by (Campbell 1974)

$$K_3 = K_s W_3^{(2B+3)}, \qquad (15)$$

where $W$ is the ratio of soil moisture held to total soil moisture capacity, and the saturated hydraulic conductivity $K_s$ and the coefficient $B$ are functions of soil type (Clapp and Hornberger 1978).

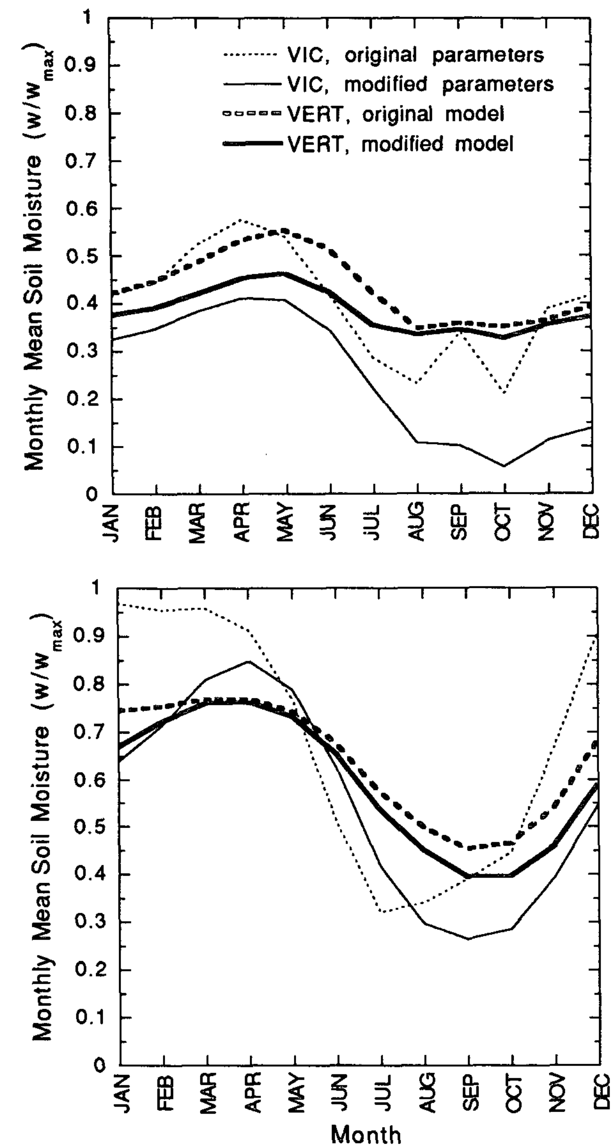


FIG. 7. Monthly mean soil moisture fraction in box 1 (top) and box 2 (bottom) from the original and modified VIC parameters, and the VERT models.

In the original SiB model (Sellers et al. 1986), runoff was also produced when the precipitation rate reaching the soil surface exceeded the infiltration rate of precipitation into the upper soil layer. In our off-line studies, this procedure was found to produce unrealistically large runoff events during periods of low hydraulic conductivity, such as late summer, because the precipitation was unable to infiltrate the soil. In the VERT implementation, all precipitation reaching the ground enters the soil matrix. An alternative approach is to save the moisture that did not infiltrate in a "pond"

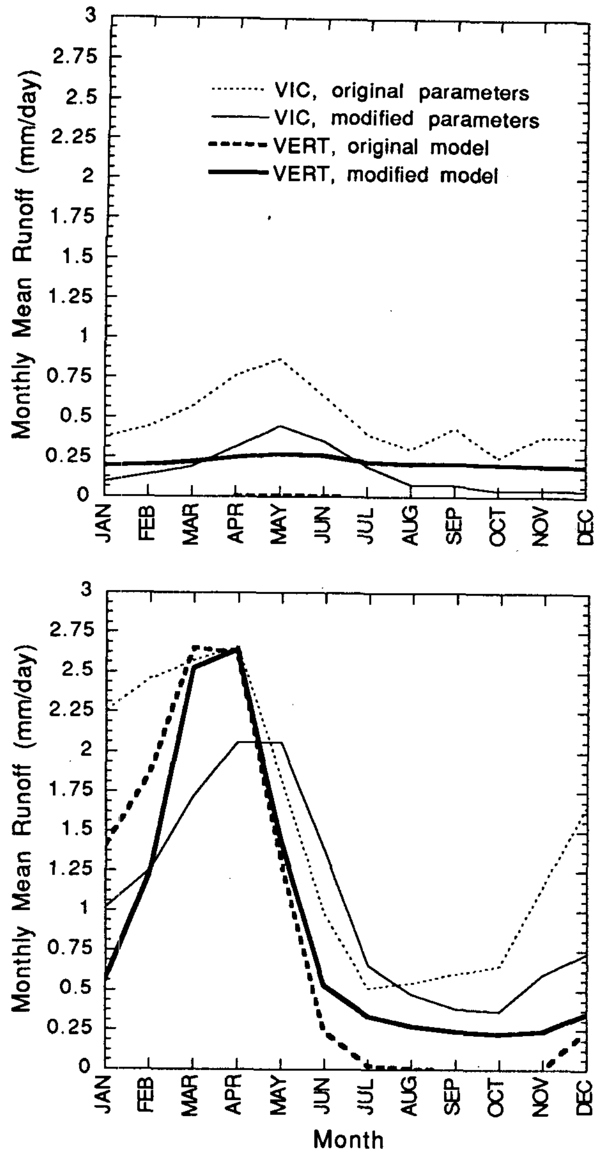


FIG. 8. Monthly mean runoff produced in box 1 (top) and box 2 (bottom) from the original and modified VIC parameters, and the VERT models.

that is saved between time steps (Verseghy 1991). In practice, these later two procedures were found to produce similar results. Other details of the soil moisture model formulation can be found in Sellers et al. (1986). In our study of the Mississippi basin, we have used the model parameters corresponding to a loam–grassland, soil–vegetation type 7 (Sud et al. 1990).

Soil moisture losses resulting from evapotranspiration are computed by scaling the potential evapotranspiration by the soil water stress factor

$$f(\psi) = 1 - \exp\{-c_2[c_1 - \ln(-\psi)]\}, \quad (16)$$

where $c_1$ and $c_2$ are empirical constants representing the progression of closing stomata (Xue et al. 1991), and

$$\psi = \psi_s W_2^{-B}, \quad (17)$$

where $\psi_s$ is the soil moisture potential when the middle layer $W_2$ is saturated (Clapp and Hornberger 1978).

### b. *Variable infiltration capacity model*

The second GCM land surface hydrology parameterization used to produce runoff input for the runoff routing model is the variable infiltration capacity (VIC) model of Wood et al. (1992). This parameterization attempts to account for the spatial variability found within a GCM grid. The model considers two runoff mechanisms, active runoff due to precipitation occurring during the current time period and base flow produced by drainage from the soil column and modeled as a linear reservoir. In the formulation, active runoff is a strong nonlinear function of precipitation and soil wetness. VIC can be viewed as depicting the GCM grid as a collection of regions with differing infiltration capacities; as the fraction of the grid box containing saturated regions increases, the active runoff produced by a precipitation event also increases.

In the model, the infiltration capacity $i$ is given by

$$i = i_m[1 - (1 - A)^{1/b}], \quad (18)$$

where $b$ is a parameter that determines the shape of the infiltration capacity curve, $i_m$ is the maximum infiltration capacity, and $A$ is the areal fraction of the grid cell where the infiltration capacity is less than $i$ ($0 \leqslant A \leqslant 1$). Integrating yields

$$i_m = W_c(1 + b), \quad (19)$$

where $W_c$ is the maximum soil moisture storage capacity. The soil infiltration capacity $i_o$ is

$$i_o = i_m\left[1 - \left(1 - \frac{W_o}{W_c}\right)^{1/(1+b)}\right], \quad (20)$$

where $W_o$ is the soil moisture storage. Runoff produced by precipitation falling on the saturated grid fraction is given by

$$Q_d = \begin{cases} P - W_c + W_o, & i_o + P \geqslant i_m \\ P - W_c + W_o + W_c\left[1 - \left(\dfrac{i_o + P}{i_m}\right)\right]^{(1+b)}, & i_o + P \leqslant i_m, \end{cases} \quad (21)$$

where $P$ is precipitation. Base-flow runoff is modeled as linear reservoir drainage out of the soil moisture store:

$$Q_b = k_b \frac{W_o}{W_c}, \quad (22)$$

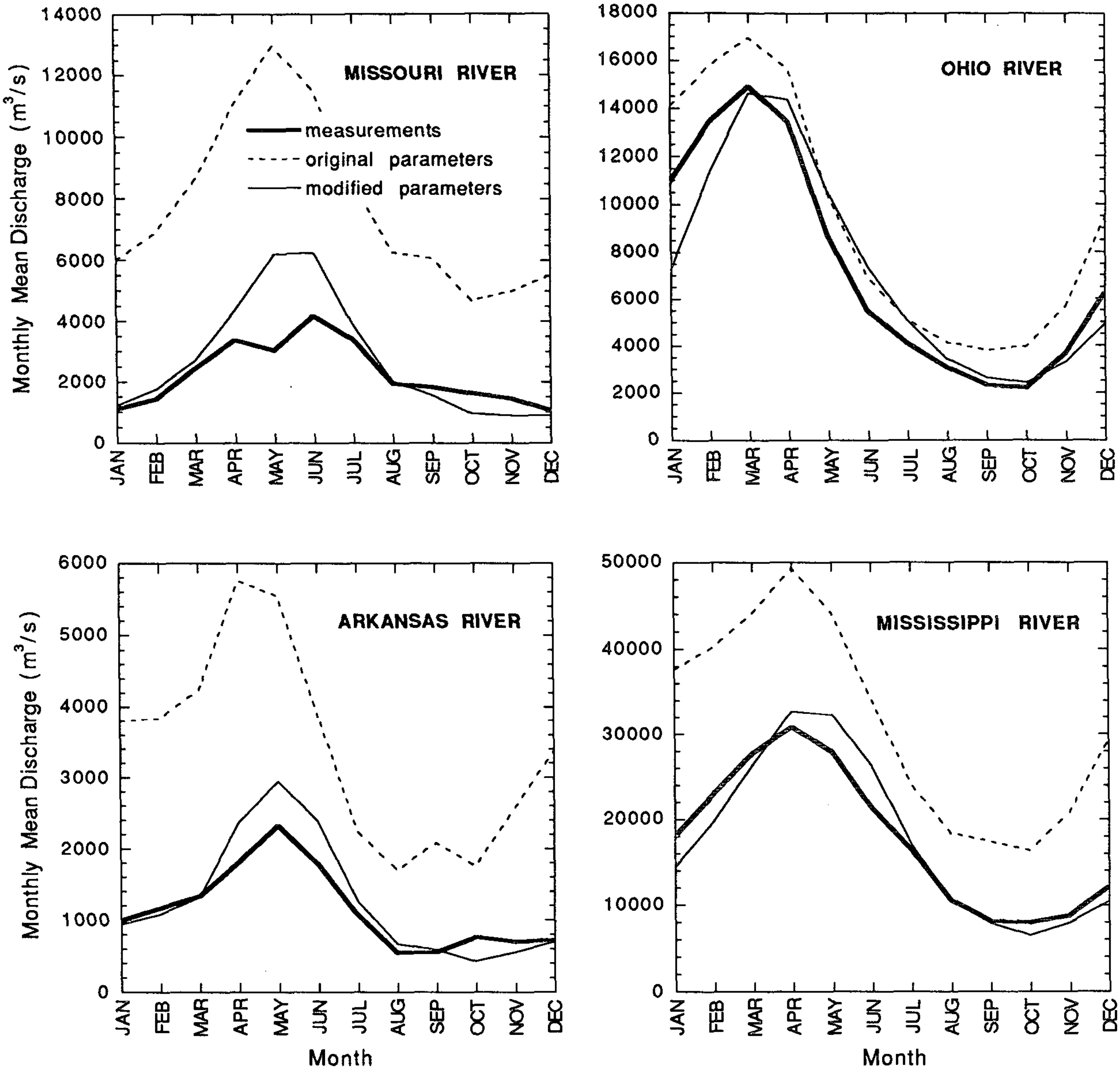


FIG. 9. Mississippi basin and subbasin hydrographs produced by running VIC gridded runoffs, with and without modifications, through the runoff routing model. Also shown are river discharge measurements (van der Leeden 1975).

where $k_b$ describes the drainage when the soil is at field capacity. The total runoff $Q_t$ is

$$Q_t = Q_d + Q_b. \tag{23}$$

Evapotranspiration $e$ is computed by scaling the potential evapotranspiration by a factor that is a function of the spatial variation of soil moisture,

$$e = e_p\left[1 - \left(1 - \frac{W_o}{W_c}\right)^{1/b_e}\right], \tag{24}$$

where $b_e$ is an evaporation shape parameter and $e_p$ is the potential evapotranspiration. The water balance is performed by updating the soil wetness at the previous time step $W_o^-$, such that

$$W_o^+ = W_o^- + P - Q_t - e, \tag{25}$$

where $W_o^+$ is used for $W_o$ in (20) and (21), and $W_o^-$ is used for $W_o$ in (22) and (24).

## 5. Results and discussion

The results of four model integrations are presented: the first two consider each parameterization in its original form, while the second two integrations include modifications made to each model or model parame-

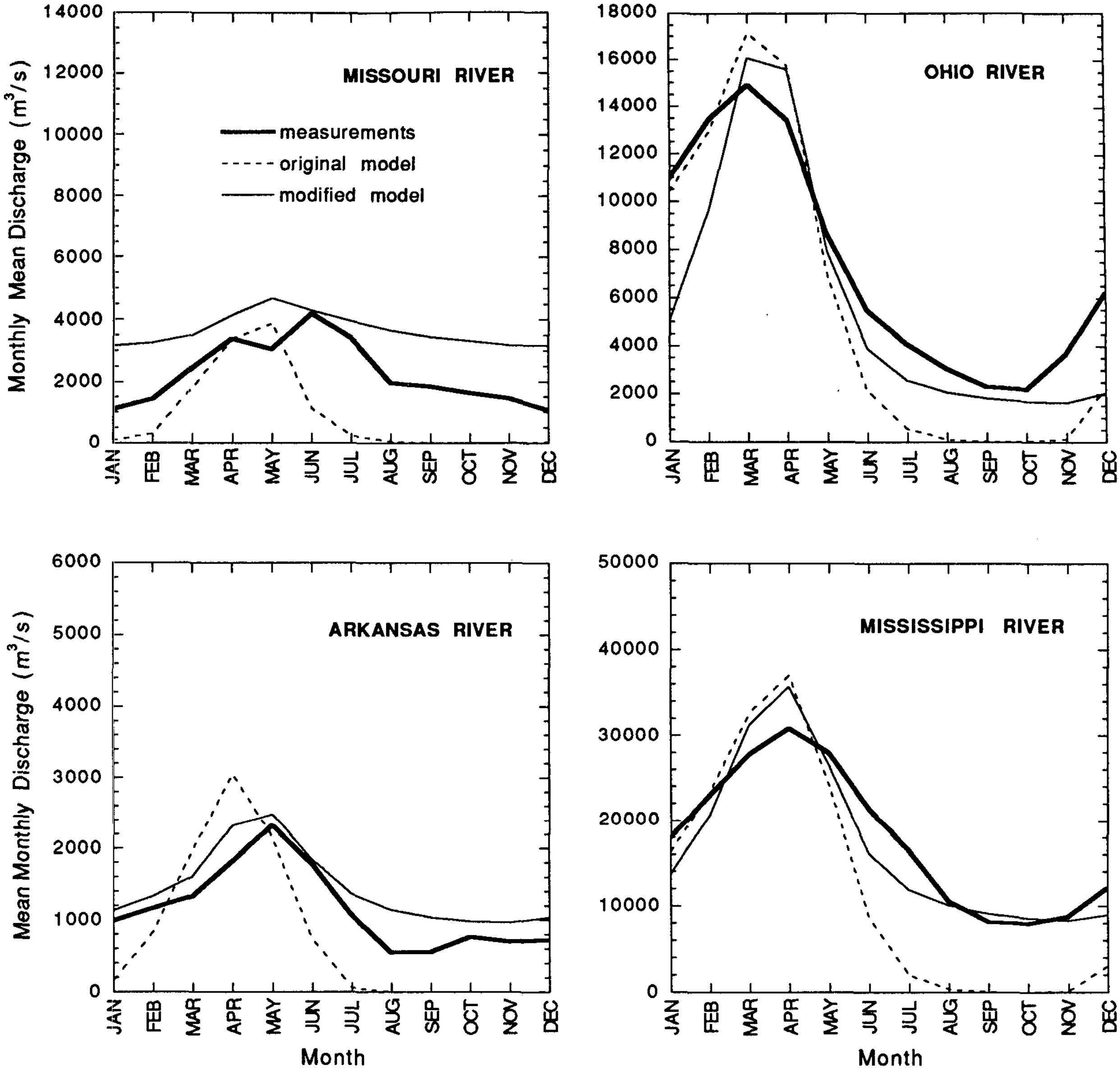


FIG. 10. Mississippi basin and subbasin hydrographs produced by running VERT gridded runoffs, with and without modifications, through the runoff routing model. Also shown are river discharge measurements (van der Leeden 1975).

ters. In the original VERT model, the grid box is considered horizontally homogeneous. In the modified version, to account for the grid-box spatial variability, we have assumed that the box contains a saturated areal fraction that is equal to the soil wetness of the third soil layer. This saturated region then drains as a linear reservoir, thus producing an additional term $Q_v$, proportional to $W_3$, in the governing equation for the third soil moisture store:

$$\frac{\partial W_3}{\partial t} = \frac{1}{\theta_s D_3}(Q_{2,3} - Q_3 - Q_v), \qquad (26)$$

with

$$Q_v = k_v W_3, \qquad (27)$$

where $\theta_s$ is the soil porosity, $D_3$ is the thickness of the third layer, $Q_{2,3}$ is the moisture flow between layers 2 and 3, and $k_v$ is the discharge from the third layer at saturation, equal to 0.6 mm day$^{-1}$ for the Mississippi basin based on optimizing the objective function $\sum (Q_o - Q_p)^2$, where $Q_o$ is the observed Mississippi basin discharge and $Q_p$ is the model prediction.

In this implementation of the original VIC model, the four model parameters, the moisture storage capacity $W_c$, the infiltration parameter $b$, the evaporation parameter $b_e$, and the base-flow parameter $k_b$ were all assumed to be spatially constant over the basin. The values used were taken from Wood et al. (1992) and are $W_c$ = 15 cm, $b$ = 0.3, $b_e$ = 0.5, and $k_b$ = 0.005. The values of these parameters were determined, in part, through an optimization procedure while applying the VIC model to the French Broad River. In the simulation using modified VIC parameters, these parameters are defined to vary spatially over the Mississippi basin according to the soil water holding capacities given by Patterson (1990). For each 4° × 5° grid cell covering the Mississippi basin, Patterson provides between 49 and 99 values of soil water holding capacity. The average capacity within each cell has been defined to be the moisture storage capacity $W_c$. These values range from 22 to 64 cm across the basin (Fig. 3). The infiltration parameter $b$ was defined according to (19) where the maximum infiltration capacity $i_m$ was defined to be the maximum soil water holding capacity of all the points within the cell. Values of the infiltration and evaporation parameters vary from 0.1 to 1.2 and from 0.1 to 0.4, respectively, and are presented in Fig. 3. The base-flow parameter $k_b$ was assumed to be spatially constant and equal to 0.01 over the basin.

The two land surface hydrology parameterizations were run using daily climatological values of potential evapotranspiration and climatologically scaled 1979 First GARP (Global Atmospheric Research Program) Global Experiment (FGGE) precipitation data. Air temperature measurements were obtained from the National Center for Atmospheric Research climatological dataset (Spangler and Jenne 1984). The climatological potential evapotranspiration data were computed following Thornthwaite (1948). Using measurements of the evapotranspiration and surface air temperature under potential or near-potential conditions, Thornthwaite obtained an empirical formulation for the potential evapotranspiration. Since the measured surface air temperature is frequently obtained under nonpotential conditions, measured temperatures often require modification before they can be used in the Thornthwaite equations. These temperature modifications were made following Mintz and Walker (1993), where the air temperature in the absence of soil moisture stress was determined as a function of the soil wetness. Magnitudes of the 1979 FGGE precipitation data were scaled by Jaeger (1976) precipitation climatology, while still preserving the frequency of the precipitation events. The data were reduced to a 4° × 5° GCM grid.

The implementation of VERT and VIC in GCMs generally includes the mechanisms to account for snow accumulation and melt in the land-surface hydrology simulations. In the off-line simulations performed in the current study, these snow-related processes have not been included, and all precipitation was treated as rainfall. For the Mississippi basin, we have assumed that this approximation will not significantly influence the monthly mean runoff simulations. The one notable exception to this is the relatively high-elevation Yellowstone River basin, which strongly feeds the Missouri River from its spring snowmelt. Clearly the neglect of snow-related processes would not be appropriate in other watersheds that contain a large fraction of runoff resulting from snowmelt, such as the Columbia and Colorado Rivers.

As an example of model precipitation and potential evapotranspiration inputs, consider the 4° × 5° grid boxes identified in Fig. 4 as "box 1" and "box 2." Figure 5 contains these inputs for boxes 1 and 2. The precipitation curves indicate that box 1 is in a much drier region than box 2. The evapotranspiration, soil moisture, and runoff produced by the two parameterizations over an annual cycle, with and without modifications, are displayed in Figs. 6, 7, and 8, respectively, for boxes 1 and 2.

The VIC and VERT depiction of evapotranspiration, soil moisture, and runoff display distinct differences between each other, and between boxes 1 and 2. Box 1 soil moisture and runoff values are typically lower than those found in box 2, and the seasonal variations in soil moisture produced by VIC are greater than those produced by VERT for both boxes. The 15-cm soil moisture capacity of the original VIC model produced a nearly saturated soil in box 2 during late winter, and consequently, winter precipitation rapidly found its way into runoff. Modification of the VIC evapotranspiration parameter has led to increased evapotranspiration in both boxes 1 and 2. The increased moisture capacities and increased evapotranspiration introduced in the modified model produced generally drier soil and decreased runoff. In box 2, the increased moisture capacity in the modified VIC model has allowed room for precipitation to accumulate over the winter before being released as runoff the following spring. The lower soil moisture obtained in the modified VERT simulation has led to reduced evapotranspiration, particularly in box 1, where soil moisture is low enough to significantly reduce the evapotranspiration from its potential value. In contrast, soil moisture in box 2 is high enough to produce evapotranspiration at nearly the potential rate.

Little or no runoff was produced by the original VERT model in box 1 during the year, while box 2 had no runoff during the late summer and early fall months. This results from the horizontally homogeneous character of the VERT model grid boxes. When the soil within a grid box reaches some minimal wetness (approximately 0.65 for the loam–grassland soil–vegetation type), the corresponding small value of soil hydraulic conductivity prevents any runoff from the lowest soil layer, over the grid box area. In the modified VERT model, the subgrid-scale influence of the linear reservoir produces a runoff component during all seasons. The increased runoff lowers the soil moisture, which, in box 2, decreases the winter runoff.

The 4° × 5° grid box runoffs produced by VIC and VERT were applied uniformly over each of the four coincident 2° × 2.5° runoff routing model grid boxes. The model-generated discharge hydrographs for the gauging stations identified in Fig. 2 are found in Figs. 9 and 10 for the VIC and VERT parameterizations, respectively. For comparison, climatological Mississippi basin discharge measurements (van der Leeden 1975) are also included (1897–1965 averages for the Missouri, 1927–65 averages for the Arkansas, and 1928–65 averages for the Ohio and Mississippi Rivers). The second peak in the Missouri River data is due to the snowmelt runoff from the Yellowstone River, a feature not addressed in the present study. A notable characteristic of the hydrographs resulting from the original VIC model is the overprediction of discharge from the western and central Mississippi basin. The modified model produces a much improved annual basin water balance. In the original VERT model the depletion of late summer and fall grid-box runoffs produced dry riverbeds during the fall. The addition of the linear reservoir drainage from VERT's lowest layer reduced the storm-produced runoff, while the additional groundwater runoff drained from the reservoir raises the autumn portion of the river hydrographs, producing more realistic hydrographs and seasonal water balances. VIC, while not considering the vertical soil structure modeled by VERT, has accounted for the horizontal variability of soil moisture holding capacity within a grid box, and thus partitions a fraction of the available precipitation into storm runoff even during the driest months. This, in addition to its linear reservoir drainage, enables VIC to support river discharge throughout the fall period.

## 6. Summary

We have presented a runoff routing model that produces discharge hydrographs from regional- or continental-scale gridded runoff data. As such, it provides an opportunity to compare gridded hydrologic model output with available river discharge measurements. It is a horizontally coupled model where upstream regions of the watershed influence regions downstream. In addition, the model includes the ability to account for spatial variation of geographic factors that affect the residence time of the fluid within specific regions of the watershed. Implementation of the model requires three basic groups of information, the connectivity of the overland flow network, the definition of residence time coefficients for differing geographic regions, and gridded runoff input from a land surface hydrology parameterization. Analysis of the two GCM parameterizations using the runoff routing model indicates a strong need to provide physically realistic soil moisture storage capacities, evapotranspiration coefficients, and runoff mechanisms, and account for the subgrid-scale variability of those parameters and components within land surface hydrology formulations. The modification considered in the VERT parameterization discussed above is being tested in the SiB version of the GLA GCM and is expected to improve the model simulation of land surface hydrologic balances.

The runoff routing model can be applied to other regions through definition of basin overland flow paths and computation of residence time coefficients for those locations. Expanded globally, the model can be used to transport continental GCM precipitation to the oceans for use in coupled atmosphere–ocean models. In regions where significant evapotranspiration occurs from runoff that had its origins far upstream, such as may occur due to river flooding in the tropics and elsewhere, or where water is collected for irrigation, the routing model can be expanded and coupled to the GCM, allowing for the horizontal transport of available moisture from one region of the GCM grid network to another.

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