Topographic Influence on the Seasonal and Interannual Variation of Water and Energy Balance of Basins in North America

JI CHEN AND PRAVEEN KUMAR

Environmental Hydrology and Hydraulic Engineering, Department of Civil and Environmental Engineering, University of Illinois at Urbana–Champaign, Urbana, Illinois

(Manuscript received 24 January 2000, in final form 21 September 2000)

ABSTRACT

A large area basin-scale (LABs) hydrologic model is developed for regional, continental, and global hydrologic studies. The heterogeneity in the soil-moisture distribution within a basin is parameterized through the statistical moments of the probability distribution function of the topographic (wetness) index. The statistical moments are derived using GTOPO30 (30 arc sec; 1-km resolution) digital elevation model data for North America. River basins and drainage network extracted using this dataset are overlaid on computed topographic indices for the continent and statistics are extracted for each basin. A total of 5020 basins with an average size of 3255 square kilometers, obtained from the United States Geological Survey HYDRO1K data, is used over the continent.

The model predicts runoff generation due to both saturation and infiltration excess mechanisms along with the baseflow and snowmelt. Simulation studies are performed for 1987 and 1988 using the International Satellite Land Surface Climatology Project data. Improvement in the terrestrial water balance and streamflow is observed due to improvements in the surface runoff and baseflow components achieved by incorporating the topographic influences. It is found that subsurface redistribution of soil moisture, and anisotropy in hydraulic conductivities in the vertical and horizontal directions play an important role in determining the streamflow and its seasonal variability. These enhancements also impact the surface energy balance. It is shown that the dynamics of several hydrologic parameters such as basin mean water table depth and saturated fraction play an important role in determining the total streamflow response and show realistic seasonal and interannual variations. Observed streamflow of the Mississippi River and its subbasins (Ohio, Arkansas, Missouri, and Upper Mississippi) are used for validation. It is observed that model baseflow has a significant contribution to the streamflow and is important in realistically capturing the seasonal and annual cycles.

1. Introduction

During the last two decades significant advances have been made in the representation of terrestrial hydrologic processes for studying the feedback between the land surface and atmosphere. These studies were largely facilitated by the improved understanding of the role of vegetation in determining the transfer of energy and carbon dioxide fluxes at the land–atmosphere interface that lead to the subsequent development of macroscale hydrologic models generally termed as Soil–Vegetation–Atmosphere-Transfer (SVAT) schemes (e.g., Sellers et al. 1986; Dickinson et al. 1986; Abramopoulos et al. 1988; Famiglietti and Wood 1991; Koster and Suarez 1992a,b; Bonan 1996; Pitman et al. 1991; Yang et al. 1995). Studies using these models have demonstrated the important role of the terrestrial moisture and heat stores in modulating the climate system. With the increasing sophistication, resolution and performance of the general circulation models (GCMs) it is now possible to study the impact of climatic variability on terrestrial systems.

Hydrologic models, which provide the link between the physical climate and several terrestrial systems, are now being adapted for such studies. Important advances have been made in the prediction of streamflow (see Abdulla et al. 1996; Nijssen et al. 1997; Wood et al. 1997). However, important limitations exist primarily due to two reasons. First, the emphasis of the SVAT model development has been on the exchange of moisture and energy fluxes at the land–atmosphere interface and not on the subsurface and ground water flow dynamics that are largely responsible for the seasonal to interannual streamflow variability. Second, the existing SVAT models are run on rectangular grids that coincide with that of the GCMs, typically of the order of 200 to 400 km on a side. However, the natural unit for the representation of hydrologic processes is a basin. In a basin, in the absence of geologic controls, both the surface and subsurface flows are controlled by the topog-
raphy with the water leaving the basin through its mouth. Soil moisture is higher in regions of flow convergence such as at the bottom of hill slopes, whereas uphill areas are associated with flow divergence and higher soil-moisture deficit in the vertical profile. A significant portion of the total runoff is generated when rain falls on the saturated regions in the low lying areas near stream channels. This saturated area expands or contracts during and between storms. This mechanism can result in surface runoff production in humid and vegetated areas with shallow water tables, even where infiltration capacities of soil surface are high relative to normal rainfall intensities. The main controls on the saturated contributing areas are the topographic and hydrologic characteristics of the hill slopes. The heterogeneity in the distribution of soil moisture influences the evapotranspiration rate and consequently the energy balance. A basin-based representation not only provides a logical way of modeling the vegetation and moisture heterogeneity but also opens the door for the study of climate influences on surface hydrology to address issues related to water resource management, and ecological and environmental issues over seasonal to interannual timescales.

Topmodel (Beven and Kirkby 1979), which has undergone significant enhancements over the years and incorporates both infiltration and saturation excess runoff generation mechanisms, is often used for modeling the basin runoff. The basic approach of this model has been adapted into the SVAT schemes (e.g., Famiglietti and Wood 1991) for the representation of soil-moisture heterogeneity using the probability distribution of the topographic index. However, the streamflow prediction is not emphasized.

In another study, Stieglitz et al. (1997) developed a model that coupled an SVAT scheme (Pitman et al. 1991) with a Topmodel-based basin runoff model (Beven and Kirkby 1979). Their essential idea was to treat a basin, rather than a rectangular grid, as a column and couple the Topmodel equations with that of the single column model. From a hydrologic perspective, a column model has its strength in predicting (i) the partition of the incident radiation into various heat components, and (ii) the vertical movement of soil moisture. However, it is weak in predicting the runoff. In contrast, Topmodel can predict the runoff well but does not model the energy fluxes and the vertical transport of soil moisture. Stieglitz et al. (1997) tested the coupled model in the 8.4 km² W-3 subwatershed of the Sleepers River basin located in the glaciated highlands of Vermont and obtained very promising results. This development enables us to account for and validate water balance along with the energy balance. In addition, it offers the opportunity for studying the seasonal, interannual, and decadal fluctuations of land surface hydrologic parameters such as streamflow, water table depth, subsurface storage, etc., in a meaningful way.

Recent availability of digital elevation model (DEM) data, and basin and stream coverages from the United States Geological Survey (USGS) for the various continents (Verdin and Verdin 1999) has made it possible to implement basin-scale models for regional, continental, and global studies. The emphasis of this paper is twofold. First, we describe a basin-scale model, based on the scheme proposed by Stieglitz et al. (1997), for studying the hydrologic response over the entire North American continent. Several enhancements are made to the basic scheme. The National Center for Atmospheric Research (NCAR) Land Surface Model (LSM; Bonan 1996) is chosen as the underlying column model for this work. It is a well-tested model and already coupled to the Community Climate Model, Version 3 (CCM3), at NCAR. For application over the North American continent, estimates of topographic indices and basin delineation are obtained using USGS GTOPO30 digital elevation and HYDRO1K data (Verdin and Verdin 1999). This latter dataset employs an efficient representation scheme for the nested basin topology (Pfafstetter 1989) that facilitates studies of streamflow response at several aggregation levels. Model simulations are performed using the International Satellite Land Surface Climatology Project (ISLSCP) dataset for 1987 and 1988 (Sellers et al. 1995; Meeson et al. 1995). The model is validated using observed streamflow from several rivers such as the Mississippi, Arkansas, Missouri, Upper Mississippi, and Ohio (EarthInfo 1995).

Second, we explore the role of topographic control in the hydrologic response at the regional and continental scales. The study demonstrates the improvement in the terrestrial water balance and streamflow through improvements in the surface runoff and baseflow components achieved by incorporating the topographic influences. These enhancements also have an impact on the surface energy balance. It is shown that several hydrologic parameters such as water table depth and saturated fraction play an important role in determining the total streamflow response and show realistic seasonal and interannual variations.

This paper is organized as follows. In section 2 we described the model used for the study. The data and the results from the simulation are described in section 3. Summary and conclusions are given in section 4. The appendixes summarize the technical details.

2. Model description

a. Coupling scheme

Our hydrologic model, called the large area basin-scale (LABs) model (Fig. 1), is based in part on the scheme developed by Stieglitz et al. (1997). Each basin is represented as a single column in order to capture the dynamics of moisture and energy transport. For the vertical transfer of heat and moisture fluxes we use the NCAR LSM model (Bonan 1996), which is a one-dimensional column model that simulates energy, mo-
momentum, water, and CO₂ exchanges between the atmosphere and the land surface. It incorporates subgrid variability by subdividing a grid box into mosaics with different landuse categories. All computations are performed independently for each subgrid and the area-weighted fluxes are then computed for the entire grid. The vertical column is divided into six layers. The near-surface layer is 10 cm deep and thickness doubles for subsequent layers, with the bottom of the last layer at 6.3-m depth.

The NCAR LSM model simulates ecosystem dynamics, and biophysical, hydrologic, and biogeochemical processes. Ecosystem dynamics includes vegetation phenology (Dorman and Sellers 1989) that simulates leaf and stem area indices that are updated daily. The biogeochemical processes that are simulated by the model are photosynthesis, plant and microbial respiration, and net primary production. Biophysical processes consist of computation of albedo, radiation fluxes through the canopy, and heat and momentum fluxes at the land–atmosphere interface. The momentum, sensible, and latent heat fluxes are derived by using Monin–Obukhov similarity theory for the surface layer. Canopy and ground temperatures are computed by considering their energy balance equations. Soil and lake temperatures are obtained by using the heat diffusion equation.

The hydrologic processes that are simulated in NCAR LSM include interception, throughfall, and stemflow; snow accumulation and melt; infiltration and runoff; and vertical transfer of soil moisture. The soil properties such as soil hydraulic conductivity and soil matrix potential are calculated as a function of the soil texture (Clapp and Hornberger 1978). Additional details of this model can be found in Bonan (1996). Several enhancements are made to the model in our implementation and they are described in section 2b.

In NCAR LSM the soil-moisture heterogeneity is parameterized through an assumed exponential distribution. In LABs, the spatial heterogeneity of soil moisture within the basin is parameterized through the probability distribution of the topographic (wetness) index (Beven and Kirkby 1979). The topographic index at any location in the basin is defined as

\[
\text{Time}
\]

\[
\text{Exponential decay rate } \zeta
\]

\[
\text{Vertical } K_{sz} = \alpha K_{sz}
\]

\[
\text{Lateral } K_{sx} = \alpha K_{sz}
\]

\[
\text{Runoff} = \text{Baseflow} \left[ f(\lambda, Z_{0T}, \zeta, \alpha, K_{sz}) \right] + \text{Surface flow}
\]

\[
\text{PDF of Topographic Index}
\]

\[
\text{Saturated zone}
\]

\[
\text{Supplied Limited ET}
\]

\[
\text{Demand Limited ET}
\]

\[
\text{Saturated Hydraulic Conductivity}
\]

\[
\text{Total Evapotranspiration}
\]

\[
\text{No moisture and heat flux from the bottom}
\]
\[ x = \ln \frac{a}{\tan \beta} \]  

where \( a \) is the upstream contributing area, from the watershed divide, per unit contour length, and \( \tan \beta \) is the local slope. Regions of flow convergence and low vertical soil-moisture deficit are identified through large values for the topographic index and the smaller values correspond to uphill areas of flow divergence and/or high vertical soil-moisture deficit. The topographic index in a basin can be easily computed using DEM data (Quinn et al. 1991; Jenson and Domingue 1988). It can be shown (Beven et al. 1995) that the relationship between the basin average water table depth \( z_T \) and the local water table depth \( z \) (positive downward from the surface) at any location within the basin is given as

\[ z = z_T - \frac{1}{\xi} (x - \lambda) \]  

where \( \xi \) is a hydrological constant and is a measure of the decline of the saturated hydraulic conductivity with increasing depth [see Eq. (5)]. The parameter \( \lambda \) is the mean value of the topographic indices over the basin.

Equation (2) implies that all locations in the catchment with the same value of topographic index will have the same relationship between local water table depth and mean water table depth. Therefore, a hydrologic similarity assumption is invoked, which states that all locations in a basin with the same topographic index will have the same hydrologic response. Instead of discretizing the basin, which is used in traditional applications of Topmodel, the analytic form of the Topmodel equations (see Stieglitz et al. 1997) is applied in LABs. Consequently, the spatial heterogeneity is modeled through the probability distribution of the topographic index, \( f(x) \), often parameterized as a three-parameter gamma distribution (Sivapalan et al. 1987; Kumar et al. 2000). For any particular value of \( z_T \), knowledge of the pattern of topographic index allows prediction of those areas where \( z \leq 0 \), that is, the saturated contributing area, \( A_s \), where all rainfall contributes to runoff without any infiltration. The contributing area for direct runoff, \( A_{dr} \), in a basin of area \( A \), is given as

\[ A_{dr} = \int_{x=z_T+\lambda}^{\infty} f(x) \, dx. \]  

During rainfall, the water table rises resulting in larger fractions of the area contributing to direct runoff.

The basic computational algorithm for the coupled scheme is as follows (see appendix A for details). Using the infiltration as the upper boundary condition and no flow boundary condition at the bottom, the mean soil moisture in each layer is obtained by solving the Richard’s equation (Bonan 1996). The mean water table depth is then computed using the knowledge of the soil-moisture deficit in the entire column. The probability distribution function for the topographic index, \( f(x) \), pre-computed from the analysis of DEM data is then used to generate the saturated fraction of the catchment [see Eq. (3)]. This determines the fraction of rainfall that becomes direct runoff and infiltration, at the subsequent time step. The infiltration contributes to the vertical transport of soil moisture, and the cycle is repeated. The no-flow boundary condition at the bottom of the last layer implies that any infiltrated water is lost from the basin only as baseflow and evapotranspiration. The streamflow consists of the surface runoff and baseflow. The baseflow, \( Q_b \), is computed as a function of the mean water table depth, given as (Sivapalan et al. 1987):

\[ Q_b = AT_0 k_s e^{-(a + cz_T)}, \]  

where \( T_0 = k_s(0)/\xi \) is the basin lateral transmissivity when the soil is just saturated, and \( k_s(0) \) is the saturated lateral hydraulic conductivity of the soil at the surface and the methodology to determine this value is described in section 2b(1).

In order to incorporate the heterogeneity due to vegetation, subregions of four landuse categories along with a category for water body are implemented. Each of the four landuse categories can be either in the saturated region or unsaturated region. In each region they are assumed to be homogeneously distributed. We, therefore, effectively have nine possible subregions in each basin consisting of four landuse categories in the saturated and unsaturated regions each, and a water body (see Fig. 2). All surface fluxes are obtained as area-weighted averages of each subregion. The details of the moisture flux computations of the above-described scheme are given in appendix A.

b. Model enhancements

1) ANISOTROPY IN SOIL-HYDRAULIC CONDUCTIVITY

An anisotropic factor, \( \alpha \), is introduced to model the differences in the saturated hydraulic conductivities in the vertical and lateral directions. The lateral value \( k_v(z) \) is used in the computation of the baseflow due to surface runoff and ground-water flow, and the vertical value \( k_s(z) \) is used in the computation of the vertical moisture transport through the different layers. Sensitivity analysis showed that \( k_v(z) = \alpha k_s(z) \), where \( \alpha \approx 2 \times 10^3 \) was required to simulate the desired streamflow response. It is well known that lateral hydraulic conductivities are significantly larger than the vertical ones (Freeze and Cherry 1979, p. 34). In the model, the anisotropy is required to appropriately capture the baseflow recession [see Eq. (4)]. The value of \( \alpha \) given above is used for all basins studied since no observational data is available for this parameter. A better estimate for \( \alpha \) as a function of vegetation, soils, and topographic attributes such as slopes, needs to be developed for the improved prediction of subsurface redistribution of flow.
Further it is assumed that both \( k_s(z) \) and \( k_{sz}(z) \) have an exponential decay with depth, that is,

\[
  k_{sz}(z) = k_0 z e^{-\alpha z}. \tag{5}
\]

It is assumed that at a 1-m depth the vertical saturated hydraulic conductivities reach the compacted value \( k_{sz} \) given in Clapp and Hornberger (1978) under the assumption that most roots lie above this level. The lateral conductivity \( k_{sx} \) is obtained by scaling \( k_{sz} \) with the factor \( \alpha \). For each soil layer, the hydraulic conductivity at the center of the thickness is used. In the Topmodel framework \( \zeta \) is a calibration parameter (Quinn and Beven 1993; Beven et al. 1995). In our model calibration is not used. Instead a value of \( \zeta = 1.8 \, \text{m}^{-1} \) obtained through sensitivity analysis showed acceptable performance and this value is used for all basins in North America. In theory \( \zeta \) can be obtained by the analysis of the streamflow recession curves (Quinn and Beven 1993; see Eq. (4)) or through the observed decline of \( k_{sz}(z) \) with depth if such data are available. The former...
has the advantage of providing a basin-averaged estimate that is more appropriate for the proposed modeling framework. These alternatives will be explored in the future.

2) COMPUTATION OF WATER TABLE DEPTH

To define the water table depth, that is, the interface between the saturated and the unsaturated soil, Stieglitz et al. (1997) selected a soil-moisture level greater than or equal to 70% of the field capacity as the defining threshold, that is,

$$
\begin{align*}
z_b(i) & \text{ for } \theta(i) \leq 0.7 \theta_{33}, \\
\left[z_b(i) - \left(\frac{\theta(i) - 0.7 \theta_{33}}{\theta_{sat} - 0.7 \theta_{33}}\right)\right] & \text{ for } \theta(i) > 0.7 \theta_{33}.
\end{align*}
$$

(6)

where $z_b(i)$ and $\overline{\theta}(i)$ are the depth of the bottom boundary and mean soil water content of soil layer $i$, respectively, and $\theta_{sat}$ and $\theta_{33}$ are the saturated soil water content and field capacity, respectively. However, this methodology results in an abrupt change in the water table depth as it crosses layer boundaries. This in turn impacts the runoff and soil-moisture computations.

In order to resolve this problem, we use a quasi-steady-state condition (e.g., Abramopoulos et al. 1988). After the vertical transport of soil moisture is computed in the soil column using the Richards equation, the soil-moisture deficit, $D_u = \sum_i (\theta_{sat} - \overline{\theta}(i)) \Delta z(i)$, for the entire column is computed. We then assume the vertical moisture profile is such that there is no vertical flux. If $z_V$ represents the water table depth from the surface, then this condition gives

$$
\psi(z) + (z_V - z) = C,
$$

(7)

that is,

$$
\psi\left[\frac{\theta(z)}{\theta_{sat}}\right]^{-b} + (z_V - z) = C,
$$

(8)

where $\psi$ is the matrix potential, $b$ is the Brooks–Corey parameter (Brooks and Corey 1964), and $C$ is a constant. Using the conditions at the capillary fringe it is easily
Table 1. Average area of North American Pfafstetter units (after Verdin and Verdin 1999).

<table>
<thead>
<tr>
<th>Level</th>
<th>Average area (km$^2$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Level 1</td>
<td>2,209,207</td>
</tr>
<tr>
<td>Level 2</td>
<td>232,684</td>
</tr>
<tr>
<td>Level 3</td>
<td>28,713</td>
</tr>
<tr>
<td>Level 4</td>
<td>6240</td>
</tr>
<tr>
<td>Level 5</td>
<td>3255</td>
</tr>
</tbody>
</table>

Fig. 5. Level 5 Pfafstetter subdivision of a region of the eastern United States overlaid on the GTOPO30 DEM. Black and white lines represent the streams and the basin boundaries, respectively. The bar displays the elevation of the region and in units of meters.

established that the constant $C$ is $\psi_{\text{sat}}$ and the resulting moisture profile is

$$
\bar{\theta}(z) = \psi_{\text{sat}} \left[ \frac{\psi_{\text{sat}} - (z_{\tau} - z)}{\psi_{\text{sat}}} \right]^{-1/b}.
$$

(9)

The water table depth, $z_{\tau}$, is then computed by solving the equality

$$
D_{\theta} = \int_{0}^{z_{\tau}} \left[ \psi_{\text{sat}} - \bar{\theta}(z) \right] dz
$$

(10)

for $z_{\tau}$.

Once the mean water table depth is known, the base-
flow is computed using Eq. (4). This baseflow is extracted from each soil layer in proportion to the product of the soil-moisture volume and hydraulic conductivity in each layer, that is, the baseflow $q_i^b(i)$ from the $i$th layer is given as

$$q_i^b(i) = \frac{k_{i}(i)\Delta z(i)\left(\bar{\theta}(i) - \theta_{a_{i}}\right)}{\sum_{i=1}^{n} k_{i}(i)\Delta z(i)\left(\bar{\theta}(i) - \theta_{a_{i}}\right)} Q^b,$$

where $\theta_{a_{i}} = \frac{\theta_{a_{i}}(i)}{20 \deg}$ (see Bonan 1996, p. 98). This allows us to consider the lateral flow contribution to the baseflow from the unsaturated zone along with that from the saturated zone. Complete details of the water balance computations for LABs is given in appendix A.

It is remarked here that, since we impose a no-flow boundary condition from the bottom of the last soil layer, sufficient care needs to be taken in specifying the depth of the column in the model. Through sensitivity studies it was found that the soil column depth should be deep enough so that it provides sufficient storage in the column to realistically capture the water storage fluctuations of a basin. If this storage is not sufficiently provided then the water table could fluctuate more rapidly than desired, resulting in undesirable effects in runoff and surface energy balance (for details see J. Chen 2000, unpublished manuscript). In our study it is found that a soil depth that is no less than that found in nature is desired. As stated earlier, we use a depth of 6.3 m for all basins studied. ISLSCP data [see section 3a(3)] used in this study provides soil depth information for all basins studied. ISLSCP data [see section 3a(3)] used in this study provides soil depth information for the entire North America, and a depth of 6.3 m is sufficient to capture the soil depth variation across the North America continent. It should be noted that the above-described representation of subsurface hydrology is a fairly simplified representation of reality. It ignores any complicated geologic formations such as confined aquifers and their recharge zones. Consequently the average soil depth in the model may be different from the true soil depth depending upon the climate, soil, vegetation, and geology of the region.

3. Application over North America

a. Data description

1) TOPOGRAPHIC DATA AND BASIN LAYOUT

U.S. Geological Survey’s Earth Resources Observation Systems Data Center in Sioux Falls, South Dakota, has developed a global digital elevation model called GTOPO30. The resolution of this dataset is 30 arc sec (8.3 $\times$ 10$^{-3}$ deg). The vertical resolution is 1 m, and the elevation values for the globe range from -407 to 8752 m above mean sea level. Additional details about the dataset are available in Gesch et al. (1999).

A subset of this dataset corresponding to the North American continent is used for this study. The dataset was first projected from geographic coordinates to Lambert azimuthal equal-area coordinate system at a resolution of 1 km. This renders each cell, regardless of the latitude, to represent the same ground dimensions (length and area) as every other cell. Consequently, derivative estimates such as drainage areas, slopes, etc., are easier, consistent, and reliable. The extraction of these hydrographic features from the 30-arc-sec dataset are based on the drainage analysis algorithm of Jenson and Domingue (1988). This algorithm computes the flow accumulation for each cell using the direction of steepest descent. A 1000-km$^2$ threshold is then applied to the flow accumulation values to obtain a drainage network in raster format and then vectorized (Verdin and Jenson 1996). The drainage network is then used to identify basins and subbasins. In order to represent the basins hierarchically, a system developed by Pfafstetter (1989) is used that utilizes an efficient coding scheme. For the sake of completeness, this coding scheme is briefly summarized next (see Verdin and Verdin 1999 for details).

After identifying the arcs for the stream network over a continent, they are sorted into three classes: those that drain directly to the sea, those that drain directly into closed basins, and those that are tributary to arcs in

<table>
<thead>
<tr>
<th>Table 2. Downscaling equations for obtaining L moments at 90 m equivalent (y) using estimates from 1-km DEM data (x) (adapted from Kumar et al. 2000).</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mean</td>
</tr>
<tr>
<td>L scale</td>
</tr>
<tr>
<td>L skewness</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Table 3. Streamflow stations used for validation of model simulations. Basin area is obtained from the HYDR01K data using the Pfafstetter scheme. The drainage area is that documented by USGS for the station. See Fig. 4 for the location of the gauging stations.</th>
</tr>
</thead>
<tbody>
<tr>
<td>Basin name</td>
</tr>
<tr>
<td>---------------------------------------------------------------</td>
</tr>
<tr>
<td>Upper Mississippi</td>
</tr>
<tr>
<td>Missouri</td>
</tr>
<tr>
<td>Ohio</td>
</tr>
<tr>
<td>Arkansas</td>
</tr>
<tr>
<td>Mississippi</td>
</tr>
</tbody>
</table>

* The reason for basin area from HYDR01K data being less than drainage area at the station is being explored.
these first two cases. From all arcs draining directly into the sea, the four with greatest area drained are identified and assigned codes 2, 4, 6, and 8 going clockwise around the continent. Basins corresponding to each of these arcs are identified and given the same codes 2, 4, 6, and 8. The largest closed basin is given the code 0. The interbasin between basins 2 and 4 is given the code 3, that between 4 and 6 is given the code 5, and that between 6 and 8 is given the code 7. The area between basins 2 and 8 is divided between interbasins 1 and 9, which is partitioned by choosing a divide that connects basin 0 with the coast. The 10 basins resulting from this scheme, which define the level 1 subdivision, for the North America continent are shown in Fig. 3. The four
The topographic index is computed using the single-flow algorithm (Wolock 1993) that utilizes the flow accumulation and slope at every grid point. The topographic index obtained using the GTOPO30 DEM data captures the general spatial distribution over the North American continent. Using topographic index obtained from 90-m DEM data for several $1^\circ \times 1^\circ$ latitude–longitude grid boxes it is found that a simple relationship between the statistics obtained at the 1-km and 90-m resolutions can be developed (Kumar et al. 2000). The mean, standard deviation, skewness, L scale, and L skewness all show approximate linear relationships between the two resolutions making it possible to use the moment estimates from the GTOPO30 data for this study by applying a simple linear downscaling scheme. The analysis in Kumar et al. (2000) also suggests that the use of L moments for downscaling provides a better estimate than the usual moments. The L-moment ratio diagram suggests that three-parameter gamma distribution provides a reasonable approximation to the probability distribution function of the topographic index. Linearly downscaled L moments of topographic indices obtained from the GTOPO30 data for each basin at the level 5 representation are used to obtain the parameters for the three-parameter gamma distribution (see Table 2 for the downscaling equations). This parameterization is used for the computation of the water table depth for each basin in the North American continent.

### 2) Streamflow Observations

The observed daily streamflow discharges for 1987 and 1988 are obtained from the USGS (EarthInfo 1995). Stations near the basin outlet are chosen. Five stations are selected within the Mississippi River region for validation. These stations correspond to the Upper Mississippi River at Grafton, Illinois; the Missouri River at Hermann, Missouri; the Ohio/Tennessee River at Metropolis, Illinois; the Arkansas River at Murray Dam, Arkansas; and the Mississippi River at Vicksburg, Mississippi. These locations are marked on Fig. 4 and the details are summarized in Table 3.

All model simulations are performed at the level 5 description of the basin and the runoff is then aggregated to the appropriated level (level 1 for the Mississippi, and level 2 for its tributaries) using the basin coding scheme. This aggregated value is used for validation with the observations.

### 3) ISLSCP Data

Model simulation for the entire North American basins is performed using the ISLSCP Initiative I dataset (Sellers et al. 1995; Meeson et al. 1995). This dataset contains variables related to vegetation; hydrology and soils; snow, ice, and oceans; radiation and clouds; and near-surface meteorology, and covers the period 1987–88. The dataset has been mapped to a common $1^\circ \times 1^\circ$ latitude–longitude grid. The temporal frequency for most of parameters in the dataset is monthly. However, a few of the near-surface meteorological parameters are

<table>
<thead>
<tr>
<th></th>
<th>OBS</th>
<th>LABs</th>
<th>LSM</th>
</tr>
</thead>
<tbody>
<tr>
<td>Upper Mississippi</td>
<td>79.9</td>
<td>99.0</td>
<td>140.9</td>
</tr>
<tr>
<td>Missouri</td>
<td>86.1</td>
<td>85.5</td>
<td>151.4</td>
</tr>
<tr>
<td>Ohio</td>
<td>180.0</td>
<td>189.1</td>
<td>292.7</td>
</tr>
<tr>
<td>Arkansas</td>
<td>63.7</td>
<td>87.4</td>
<td>149.9</td>
</tr>
<tr>
<td>Mississippi</td>
<td>492.2</td>
<td>556.9</td>
<td>926.8</td>
</tr>
</tbody>
</table>
available both as monthly means and 6-h values. This latter part of the dataset provides all the parameters/variables to drive the model.

The LABs model is run at a time step of 0.5 h. This time step is chosen to facilitate the incorporation of the LABs model in general circulation and regional climate models for coupled model simulation studies. Therefore, it was necessary to develop temporal interpolation schemes for the shortwave and longwave radiation, as well as a methodology to model the temporal variability of the precipitation. These schemes are described in appendixes B and C, respectively. Other meteorological
data are linearly interpolated to the model time step. The forcing for each basin is obtained using an area-weighted average of the parameters from all the 1° × 1° latitude–longitude grid boxes that it straddles.

Soil texture, soil color, and soil type of each basin in North America are obtained by interpolating from NCAR LSM’s data file that uses the surface types of Olson et al. (1983), the soil colors of Biosphere–Atmosphere Transfer Scheme (BATS) T42 data (Dickinson et al. 1993), and the soil textures of Webb et al. (1993).

b. Results and discussion

Two sets of simulations are performed. The first uses the LABs model for each of the 5020 level 5 basins. The second treats each basin as a column but only vertical moisture and heat transfer using only the NCAR LSM model components is activated, that is, without the modifications due to the topographic influence as described in section 2. The second set of the simulation will be hereafter referred to as the NCAR LSM simulation. Com-
Comparison of the two allows us to assess the differences in response due to the incorporation of the topographic influence and other enhancements. Both models are run at a 0.5-h time step. The forcing data from ISLSCP is used. Initialization is done using a spinup strategy. Starting with guess initial conditions the model is driven using the 1987 dataset two times with the conditions at the end of the first run providing the initial conditions for the second run. The conditions at the end of the second run are used to obtain the initial conditions for the 2-yr simulation starting from 1 January 1987.

Since there was a severe drought in the Midwest (U.S.) during the summer of 1988, the 2-yr simulation provides an opportunity to assess the model performance for a dry year (1988) as well as an average year (1987). The simulation results below are organized around the following themes:

1) the topographic and anisotropic hydraulic conductivity influence,
2) the seasonal variations,
3) the interannual variations between 1987 and 1988,
4) water balance studies, and
5) model-simulated streamflow statistics.

1) Topographic and Anisotropic Influence

Figure 6 shows the biweekly averaged time series of model-simulated specific streamflow (streamflow per
Fig. 10. Comparison of daily specific streamflow and baseflow of the LABs simulation with observations for various subbasins in the Mississippi River region.

unit drainage area), both from the LABs and the NCAR LSM simulations, along with the observations. The figure shows the plots for the Mississippi River and four of its major tributaries. It is seen that the predictions from the LABs model match those of observation much more closely than from the NCAR LSM model. Given the one-dimensional structure of the NCAR LSM, the streamflow consists of two components: surface runoff and baseflow (including flow leaving from the bottom of the lowest layer). The baseflow constitutes 90% of
the runoff generated by the NCAR LSM model and is responsible for the excessively high total runoff volumes observed. When the model physics is improved by parameterizing the spatial variability through the Top-model formulation (see section 2a) the generated runoff volumes are significantly reduced and closer to those seen in observations.

We also notice that there is a large range in the specific discharge between the different basins which the LABs model is able to capture. The Missouri River basin shows the lowest specific discharge while the Ohio/Tennessee River basin shows the maximum. It is also noted that the Ohio/Tennessee River basin has a dominant runoff contribution to the Mississippi River and this behavior is captured by the LABs model. The simulations are consistent with the observations in capturing the early spring high flow conditions. The annual flow volumes for each of the rivers studied is given in Table 4. The LABs streamflow volumes are in general a little larger than observations. This seems to result from consistently higher prediction of low flows (see Fig. 6). To an extent this could be because aquifer recharge, regulation storage, consumptive use, and ground-water flow across the basin boundaries (see Zektser and Loaiciga 1993) are not modeled. However, the exact cause of this discrepancy can be assessed only using streamflow observations from unregulated streams. This is under investigation. From both Fig. 6 and Table 4 it is
Fig. 12. Seasonal variability of specific runoff for 1987 for a region in the eastern United States. The topographic variation of this region is shown in Fig. 5.

evident that incorporation of the topographic control in describing the basin flow dynamics improves the model performance significantly.

In order to investigate the impact of the anisotropy in the hydraulic conductivity, the LABs model is run with $\alpha = 1$, with all other conditions including the initial conditions and the spinup strategy, remaining identical. Figure 7 shows the comparison of the two LABs simulations along with the observations. It is evident that disregarding the anisotropy degrades the model perfor-
It is interesting to note that when the anisotropy is ignored, the runoff fluctuates more rapidly. This happens because the lower lateral hydraulic conductivities reduce the baseflow, resulting in more storage in the soil column. This in turn raises the water table leading to increased surface runoff. Therefore, during rainfall events, there is more surface runoff generated by the model, and between the rainfall events the baseflow is lower. Comparison of \( \alpha = 1 \) case with the NCAR LSM model in Fig. 6 also shows that when topographic effects are neglected the model performance is worse than when only anisotropic effects are neglected. This indicates that the topographic effects are more important than the anisotropic effects.

Figure 8 shows the comparison of the average latent and sensible heat fluxes for the entire Mississippi River basin for 1987 and 1988. We see that LABs predictions for the latent heat flux are in general larger than that from the NCAR LSM model. During summer this difference can be about 20% of the prediction. This should be expected since the lower streamflow of the LABs results in higher soil-moisture storage. The larger latent heat flux seems to be compensated by the sensible heat flux that is lower than that of the NCAR LSM model. These results demonstrate that the improvements in the prediction of runoff have a significant impact on the prediction of the energy balance. Figure 9 shows the spatial distribution of the mean annual latent and sensible heat fluxes for 1987. The difference of the mean annual fluxes between the LABs and NCAR LSM model are also plotted. We see that the LABs model prediction for latent heat is generally larger in the lower Mississippi/Gulf of Mexico region. This is a region of convergence of subsurface moisture and also receives high precipitation. The difference plot for the sensible heat flux (see Fig. 9) shows that increased latent heat flux is compensated by a reduction in the sensible heat flux in the same region. This characteristic also holds in the East Coast and northwest region of the United States.
2) Seasonal variations

Figure 10 shows that the comparison of daily runoff and model baseflow with the observed streamflow data. We see that the baseflow captures the seasonal variability. The surface runoff overriding the baseflow does not make a significant contribution directly to the fluctuations at the seasonal scale. The model-simulated baseflow accounts for about 60% of the annual streamflow in the Mississippi River basin.

As shown in Fig. 1 and Eq. (4), the baseflow is computed as a function of the water table depth and is contributed by all the soil layers, in proportion to the moisture content and the hydraulic conductivity. It should therefore be expected that in order to capture the seasonal variations in the streamflow, the water table depth should show appropriate dynamical response. Figure 11 shows the seasonal fluctuations of the mean water table depth for all the five basins during 1987 and 1988. This mean is obtained using a simple average of the mean water table depth of all the level 5 subbasins in the domain. The expected decrease in water table depth during the early spring and an increase during the summer is captured well. The saturated fraction, which is a function of the mean water table depth, also shows the expected seasonal variation.

Figure 12 shows the seasonal variability in the spa-
Fig. 15. Direct surface runoff (represented as layer 0) and baseflow contributions from various soil layers (layers 1–6) to the total streamflow in the various basins for 1987 and 1988. The mean annual water table depth is indicated as WT.

3) INTERANNUAL VARIATIONS

In Fig. 7 we see that the model predicted streamflow during summer of 1988, for all the basins studied, is generally lower than that in 1987. Figure 10 shows that the baseflow is lower for an extended period of time.
4) Water balance studies

Figure 14 shows the monthly averages of various components of the water balance. The evapotranspiration is very significant during the summer months and is substantially larger than the runoff. The figure also shows, during the drier 1988 summer, the evapotranspiration was largely supported by drawing on the water storage. The water storage plays a very active role in regulating the dynamics of the water balance and in general is substantially larger in magnitude than the runoff. This underscores the need for improved representation of the subterrestrial hydrology.

Figure 15 illustrates the contribution of direct surface runoff and baseflow from various soil layers to the total streamflow. The percentage contributions are summarized in Table 5. We see that the surface runoff constitutes only 30%–40% of the total runoff. The shallow soil zone (unsaturated) about 1.5 m deep from layer 1 to 4 contributes about 10%–15% with the rest of flow coming from the deep soil zone. The contribution from the shallow soil zone underscores the importance of subsurface redistribution of moisture in maintaining the water balance. It also directly impacts the surface energy balance by controlling the moisture that is available to the plant roots.

5) Model-simulated streamflow statistics

The LABs model is targeted for understanding the impact of climate variability on the terrestrial systems such as stream ecology, water quality, water resources management, etc. For this it is important that the streamflow generated from the model reproduces not only the observed patterns of seasonality and interannual variations but also the statistics of high and low flows, and the correlation structure. A comparison of model statistics with those from observations is difficult because of the regulations due to dams and reservoirs. In addition, we have not yet implemented a streamflow-routing scheme and losses due to aquifer recharge and consumptive use. Given these caveats, we studied the quantile-quantile plots of the model and observed streamflows, at temporal aggregation scales of 1 day, 1 week, 2 weeks, and 1 month. We observed that even at the 1-week timescale the model predictions gave good correspondence with the observations. Figure 16 shows the quantile-quantile plot of the observed and modeled weekly average specific runoff. If the distribution were identical, then the points would fall on the 1:1 line. We see that although the midrange quantiles are captured well, the model tends to deviate at the lower and higher range. The deviations in the lower range generally result in the over prediction by the LABs model, for reasons discussed in section 3b(1). No consistent pattern exists for the extreme events. For the Arkansas River the model tends to over predict whereas for the Ohio River we get under prediction.

Figure 17 shows the comparison of autocorrelation function (ACF) of the simulated and observed specific streamflow discharges for the various basins. The ACF of the LABs model simulation shows good correspondence with the observation.

4. Summary and conclusions

A model for the prediction of streamflow and various surface energy components is used to study the impact of the topographic influence on the seasonal and interannual variations of energy and water balances. Simulation study is performed over the North American continent for the 1987–88 period using the ISLSCP dataset.
(Sellers et al. 1995; Meeson et al. 1995). The continent is segmented into basins using the USGS HYDRO1K data (Verdin and Verdin 1999). This dataset delineates the entire continent into basins using a hierarchical scheme. Level 5 representation with an average basin size of 3255 km² is used for the simulation study. At this level of representation, the continent is segmented into 5020 basins.

The model exploits the fact that both surface and subsurface flows, from rain and snowmelt, are determined by topographic controls in the absence of complex geologic formations. Recharge to confined aquifers and consumptive use are not modeled. Topmodel formulation is used to model the runoff and an SVAT scheme based on NCAR LSM (Bonan 1996) is used for the vertical transport of the moisture and energy fluxes. Topographic index is determined using the USGS GTOPO30 1-km DEM data. The first three L moments of the topographic index are computed for each basin and linearly scaled down to provide equivalent estimates at
Fig. 17. Autocorrelation function for the biweekly averaged specific streamflows from the LABs model compared with observations.

the 90-m resolution following the work of Kumar et al. (2000). A three-parameter gamma probability distribution is used to model the spatial variability of the topographic index, parameters of which are determined using the scaled down L moments. Anisotropy between the lateral and horizontal hydraulic conductivities is also incorporated. The vertical hydraulic conductivity determines the vertical moisture transport through the six model layers. The lateral hydraulic conductivity, which is specified as three orders of magnitude larger than that in the vertical, is used in determining the baseflow from the various soil layers. Sensitivity studies showed that the depth of the basin soil column in the model needs to be consistent with that in nature to realistically capture the fluctuations of the water storage in a basin.

Validation studies are performed using streamflow observations from several tributaries of the Mississippi River that represent a range of topographic and climatic conditions. The study shows that the incorporation of the topographic control, and lateral subsurface and ground-water flows, improves the streamflow prediction significantly. The anisotropy between the vertical and lateral soil properties is also important in the streamflow prediction, but it is less influential than the topographic influence. The improved prediction of streamflow impacts the surface energy balance as well. In our study,
this resulted in an increase in the latent heat flux during the summer and a compensatory decrease in the sensible heat flux.

The model captures the seasonal and the interannual variability quite realistically. The seasonal patterns of streamflow in all the tributaries of the Mississippi River basin are consistent with the observations. The spatial patterns capture the topographic influence realistically. The model shows the impact of the 1988 drought through a decrease in streamflow and increased water table depths. The spatial distribution of the increased soil-moisture deficit, as represented by the water table depth, shows the impact in the Midwestern United States. The study also demonstrates that the baseflow predictions are extremely important for capturing the seasonal and interannual variations of the streamflow.

On an average it accounts for 60% of the annual streamflow in the different basins with 10%–15% contributed by the flow through the shallow soil zone. Comparison of streamflow statistics through quantile–quantile plots showed that in general the model slightly overpredicts the low flows. The extreme events are not captured. It is unclear at this time if this is due to model deficiency or because the streamflow observations have the effects of reservoir regulations, consumptive use, and aquifer recharge. This will be investigated further.

One of the motivations for the development of the LABs model is for providing a mechanism for assessing the impact of the climatic fluctuations on terrestrial hydrology. This will enable impact assessment on related systems such as ecological and environmental. As shown in the study, the model is able to predict hydrologic variables, such as water table depth, that are spatially and temporally consistent with what we would expect in a drier climatic condition. The consistency between the observed and the model-simulated streamflow statistics, for a range of quantiles, is also very encouraging. It leads to the possibility that the model can be used for quantitative impact assessment and decision making.

All model simulations are performed without calibration for the model parameters. Except for two parameters all model parameters are obtained using published datasets. The two parameters that are specified and fixed for all basins are the hydraulic conductivity decay constant $\zeta$ and the anisotropic factor $\alpha$. Their values were determined through extensive sensitivity analysis using the data from the Little Washita basin in Oklahoma (Allen and Nancy 1991) obtained through field experiments such as Washita'92 (Jackson and Schiebe 1993) and Southern Great Plains 1997 (SGP97).

This is a limitation of the current study but no datasets are available for their estimates directly. The simulation results presented in this paper provide at least an a posteriori justification of their importance and for the choice of their specific values used in the simulation. In theory it is possible to obtain $\zeta$ either directly from observations of hydraulic conductivity decay with depth or an analysis of baseflow recession curves (Beven et al. 1995). The anisotropic parameter $\alpha$ is even harder to determine. Additional research is required to establish the model sensitivity to these parameters depending on the specific objective of the study and ways to determine them efficiently. This issue is currently under investigation.

**Acknowledgments.** Support for this project has been provided by NASA Grant NAG5-3361 and NSF Grant EAR 97-06121. Computational support was also provided by NCSA Grant ATM990003N. Thanks are also due to Gordon B. Bonan for making the NCAR LSM model available in the public domain.

**APPENDIX A**

**Soil-Moisture Dynamics in LABs**

Each basin is divided into eight subregions ($j = 1, \ldots, 8$; excluding the water body; see Fig. 2) for computing infiltration and evaporation. Each basin column is divided into six layers ($i = 1, \ldots, 6$) for simulating soil-moisture dynamics. The amount of water available for infiltration and runoff in the $j$th subregion denoted as $P(j)$ is dependent on the interception by the vegetation in the subregions and snowmelt characteristics (see Bonan 1996 for details). The infiltration capacity is assumed to be zero for the saturated subregions, and in the unsaturated subregion it is given as (Entekhabi and Eagleson 1989):

$$f_u^* = k_u(0)\nu s + k_u(0)(1 - \nu),$$  \hspace{1cm} (A1)

where $k_u(0)$ is the saturated hydraulic conductivity at ground surface, $\nu = [-(d\psi/ds)]/[0.5\Delta z(1)]$ evaluated for $s = 1$. $\psi$ is the soil matrix potential, and $\Delta z(1)$ is the thickness of the first layer. The relative saturation in the first layer is given as $s = [\theta(1)/[\theta_u]]$, where $\theta(1)$ is the average first layer soil moisture in the unsaturated subregion of the basin [see also Eq. (A10)]. Note that the infiltration capacity is the same for all unsaturated subregions, that is, it is dependent on soil texture, which is specified as a uniform value across the entire basin, and not vegetation type. The infiltration in the $j$th subregion is computed as

$$q_{inf}(j) = \begin{cases} f_u^* & \text{if } P(j) > f_u^* \\ P(j) & \text{if } P(j) \leq f_u^* \end{cases}.$$  \hspace{1cm} (A2)

The total infiltration for the entire basin is computed as

$$q_{Tinf} = \sum_{j=1}^{4} w_{area}(j)q_{inf}(j),$$  \hspace{1cm} (A3)

where $w_{area}(j)$ is the fractional area of the $j$th subregion (note $j = 1, \ldots, 4$ since infiltration occurs in the four unsaturated subregions only, see Fig. 2).

The latent heat flux for the entire basin is obtained through the sum of the following contributions:

1) Total surface evaporation computed by considering...
demand and supply limited rates from the saturated and unsaturated subregions, respectively, for each land-use category
\[ q_{\text{trans}} = \sum_{j=1}^{8} w_{\text{area}}(j) q_{\text{trans}}(j), \quad \text{(A4)} \]
where \( q_{\text{trans}}(j) \) is the evaporation corresponding to subregion \( j \).

2) The transpiration is computed based on the similarity theory for each subregion and is a function of the landuse category (see Bonan 1996, section 8 for details). This transpiration is assumed to be derived from each of the soil layer in proportion to the root fraction. The total transpiration from the \( i \)th soil layer is given as
\[ q_{\text{trans}}(i) = \sum_{j=1}^{8} w_{\text{area}}(j) q_{\text{trans}}(j) r_{\text{root}}(i, j), \quad \text{(A5)} \]
where \( q_{\text{trans}}(j) \) is the transpiration from the \( j \)th subregion and \( r_{\text{root}}(i, j) \) is the root fraction in the \( i \)th layer for that subregion (see Bonan 1996, section 1.2 for details).

The moisture transport in the soil column is described by the Richards equation:
\[ \frac{\partial \theta}{\partial t} = \frac{\partial}{\partial z} \left[ k(z) \left( \frac{\partial \theta(z)}{\partial z} \frac{\partial \psi(z)}{\partial \theta} - 1 \right) \right]. \quad \text{(A6)} \]
For each layer the equation is
\[ \frac{\Delta \theta(i) \Delta z(i)}{\Delta t} = -q_{\text{in}}(i) + q_{\text{out}}(i) - \bar{e}(i), \quad \text{(A7)} \]
where \( \bar{e}(i) \) is the \( i \)th layer soil water content for the entire basin, \( q_{\text{in}}(i) \) and \( q_{\text{out}}(i) \) are the averaged quantities of input and output water fluxes (positive in the upward direction) from layer \( i \) over the time period of \( \Delta t \), \( e(i) \) is the averaged evaportranspiration loss, \( \Delta z(i) \) is the thickness of soil layer \( i \), and \( \Delta t \) is the model time step. This equation, with the boundary conditions of \( q_{\text{out}} \) as the flux of water into the soil and the zero flux of water at the bottom of the soil column, and with \( \bar{e}(1) = q_{\text{trans}} + q_{\text{trans}}(1) \) for the first layer and \( \bar{e}(i) = q_{\text{trans}}(i) \) for the other layers, is numerically implemented for a six-layer soil column to calculate soil water (see Bonan 1996, 98–101 for details).

In considering the baseflow, \( q_e(i) \), from each soil layer [see Eq. (11)], the soil water obtained from Eq. (A6) is updated by using the following equation:
\[ \bar{\theta}(i) = \bar{\theta}(i) - \frac{q_e(i) \times \Delta t}{\Delta z(i)}. \quad \text{(A8)} \]
The total soil water deficit, \( D_s \), through the entire soil column is computed as
\[ D_s = \sum_{i=1}^{8} [\theta_{\text{sat}} - \bar{\theta}(i)] \Delta z(i). \quad \text{(A9)} \]

The mean water table depth of the basin, \( z_r \), is computed by using Eqs. (A9) and (10). This is then used to compute the saturated fraction, \( s_r \), of the basin using Eq. (3). The average soil moisture content over the unsaturated subregion for layer \( i \) is obtained as
\[ \theta(i) = \frac{\bar{\theta}(i) - s_r \theta_{\text{sat}}}{1 - s_r}. \quad \text{(A10)} \]

Although the LABs model structure is one-dimensional, the spatial variability is parameterized through the probability distribution function of the topographic index (see Fig. 1). The model predictions for the soil moisture in each layer are predictions for the mean state, and this mean state determines the water table depth. In the LABs model, the saturated and unsaturated fractions in a basin are determined by this mean basin water table depth as well as topography [see section 2a and Eq. (3) for details]. Each of the four landuse categories, which are the same as NCAR LSM’s surface types, can exist in both saturated and unsaturated subregions (see Fig. 2). This results in different runoff and infiltration rates corresponding to different subregions. This enables us to model the subregion variabilities in water and energy fluxes, since evaporation and transpiration are at potential rates over the saturated subregions and at soil-moisture-controlled rates over the unsaturated ones. These are then aggregated to obtain the basin level energy fluxes. Saturation overland flow occurs in the saturated region and infiltration excess overland flow occurs in the unsaturated region. The aggregated infiltration is used in the solution of moisture transport through the vertical column. Furthermore, the four landuse categories in both unsaturated and saturated regions have the various properties of vegetation phenology, providing variability in the interception of precipitation and in the radiation processes in the basin.

If necessary, the local water table depth (and consequently vertical moisture deficit) at any location can be recovered using the mean water table depth through Eq. (2). This is not pursued within the scope of this paper.

APPENDIX B

Radiation Simulation

The ISLSCP data gives four values of averaged short-wave down radiation, \( I_s \) (\( J \text{ m}^{-2} \text{ s}^{-1} \)), that correspond to each 6-h period in one day. In order to drive the model at subhourly timescale the following algorithm from Bras (1990) is used to simulate the radiation values using the 6-h average.

Clear-sky shortwave radiation, \( I_c \) (\( J \text{ m}^{-2} \text{ s}^{-1} \)), is computed as
\[ I_c(t) = I_s(t) \exp[-na(t)mt(t)], \quad \text{(B1)} \]
where \( m(t) \) is the optical air mass at time \( t \) given by the following equation:
The rainfall percentage \( f(g) \), \( g = 1, \ldots, 6 \) for each of the 30-min interval in the 3-h rainfall duration is obtained using the normalized CDF values given in Huff (1967) from first-quartile 50% probability. The precipitation rate, \( I(g) \), for each 0.5-h interval is obtained as

\[
I(g) = \frac{p_1 \times f(g)}{30 \times 60} \quad g = 1, \ldots, 6, \quad (C2)
\]

where \( I(g) \) is in millimeters per second.

Once the precipitation rate for each time interval is estimated, the spatial heterogeneity is specified as per the NCAR LSM scheme (see Bonan 1996, section 8 for details).

REFERENCES


parameterization for atmospheric general circulation models including subgrid scale spatial variability. *J. Climate*, 2, 816–831.


