The Implementation and Validation of Improved Land-Surface Hydrology in an Atmospheric General Circulation Model

KEVIN D. JOHNSON, DARA ENTEKHABI, AND PETER S. EAGLESON

Ralph M. Parsons Laboratory, Department of Civil and Environmental Engineering, Massachusetts Institute of Technology, Cambridge, Massachusetts

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ABSTRACT

New land-surface hydrologic parameterizations are implemented into the NASA Goddard Institute for Space Studies (GISS) General Circulation Model (GCM). These parameterizations are: 1) runoff and evapotranspiration functions that include the effects of subgrid-scale spatial variability and use physically based equations of hydrologic flux at the soil surface and 2) a realistic soil moisture diffusion scheme for the movement of water and root sink in the soil column.

A one-dimensional climate model with a complete hydrologic cycle is used to screen the basic sensitivities of the hydrological parameterizations before implementation into the full three-dimensional GCM. Results of the final simulation with the GISS GCM and the new land-surface hydrology indicate that the runoff rate, especially in the tropics, is significantly improved. As a result, the remaining components of the heat and moisture balance show similar improvements when compared to observations.

The validation of model results is carried from the large global (ocean and land-surface) scale to the zonal, continental, and finally the regional river basin scales.

1. Introduction

General circulation models (GCMs) numerically integrate the state and balance equations governing large-scale atmospheric motions using either finite-difference or spectral methods. Due to computational constraints, the grid resolution or wavenumber cutoff in the respective methods results in the consideration of large areas (on the order of $10^2$ to $10^6$ km$^2$) as homogeneous with regard to the state variables.

There are numerous other physical processes whose dynamics occur on considerably smaller scales. Atmospheric radiation, turbulence, cloud-processes, convection, condensation, and surface hydrology are examples of such processes whose dynamics depend on heterogeneities at smaller scales than that resolved in GCMs. These physical phenomena are therefore only parameterized in climate models. Since they govern important mechanical and thermodynamic effects at the boundaries or within the global atmosphere, the model climate is sensitive to the parameterizations. Many of these parameterizations in operational GCMs are based on poorly scaled and calibrated empirical relations. The relations established between variables in a particular climatic region and at a definite scale are often applied to entirely different scales and sets of circumstances. Calibration and fine-tuning of purely empirical parameters also contribute to the reduced reliability of GCMs as numerical laboratories.

When the physical process under consideration has a nonlinear response, then the function representing the process evaluated at the mean of the subgrid conditions is not the same as the mean of the function evaluated at all subgrid locations. Processes with thresholds are typical of such cases. In this paper, the land-surface hydrology is the physical parameterization that is considered.

In hydrology, the soil infiltration and evaporation rates at points on the land surface define thresholds; the presence of the threshold essentially insures the nonlinear response of large land areas to atmospheric forcing. The determination of the effective or large-area response given the thresholds and the subgrid-scale spatial variability is the key problem. The use of nested grids and the subdivision of the large GCM grid into smaller units is an expensive proposition in terms of the computational costs. Furthermore, the scale of variability for hydrologic processes is on the order of a few meters, considerably more detailed than that which may be achieved by grid subdivision. The detrimental averaging process is, in that case, simply transferred to another scale. Subdivision of the grid into smaller areas has, nonetheless, the distinct advantage that different landscape types distinguished by terrain, vegetation, and soil types may be represented explicitly if the reliable global fine-resolution datasets may
be assembled. For similar reasons of computational efficiency and data availability, the more detailed models of biosphere processes have to be simplified before they can become part of operational GCMs.

An alternative approach makes use of statistical distributions for the key forcing parameters, and in combination with physically based soil flux equations, the large-area or grid-effective responses may be established (Warrilow et al. 1986; Entekhabi and Eagleson 1989a). In this paper, the statistical–dynamical surface runoff and evapotranspiration GCM parameterization of Entekhabi and Eagleson (1989a) is incorporated into the National Aeronautics and Space Administration (NASA) Goddard Institute for Space Studies (GISS-II; Version II) GCM (Hansen et al. 1983). The new parameterization replaces the alternative empirical formulas that have been part of the GISS-II GCM. A description of the original GISS-II formulation and the parameterization of Entekhabi and Eagleson (1989a) are briefly introduced in section 2. The inclusion of subgrid-scale spatial variability and physically based treatment of soil moisture flux and diffusion are demonstrated to considerably improve the model performance when comparisons are made against observations in section 3.

A simple climate model based on the larger three-dimensional GCM has been used by Entekhabi and Eagleson (1989b) and Johnson et al. (1991) in order to screen various sensitivity experiments. The justification for the use of the simpler model is based on the need to isolate some of the basic sensitivities in a considerably reduced and tractable model before experimenting with the full GCM with its inherent interconnectedness of perturbations and states. In appendix A of this paper, a very basic sensitivity common to all GCM land-surface parameterizations is identified using the simple climate model. Results from that finding guide the efforts in the proper implementation of new schemes in the full GCM.

Finally two important general considerations are introduced that apply to many studies in this field. First, the land-surface hydrology module in GCMs essentially receives as input, from the output of other modules, the precipitation and net radiation flux from the overlying atmosphere. The responsibility of the land-surface hydrology parameterization is to partition the incoming heat and moisture flux between output flux and the change in soil and canopy storage. The surface storage dynamics are critical since they determine the seasonality of heat and moisture forcing at the lower boundary of the atmosphere and they also act as low-pass filters. Dimensionless partitioning parameters have occupied much of the attention in improving the soil hydrology in GCMs. It is important to remember that the error in the surface flux is proportional to the errors in the forcings, that is, surface precipitation and net radiation. Thus, the amount of resources and effort spent in modeling the partitioning should be in proportion with the errors in the forcing. For this reason, it is necessary that comparable efforts be placed on improving the surface-layer parameterization (which gives estimates of the potential evaporation in part determined by the net radiation) and the parameterization that generates precipitation (large-scale condensation and moist convection).

Once modules in the GCM are replaced, the model climate needs to be validated against observed values of the global and regional heat and moisture balance. This is the second consideration. Contoured observed-versus-simulated maps and difference maps are difficult to interpret; furthermore, they imply the comparison of quantities at scales too detailed to be valid in the context of climate models. For this reason, in this study the comparisons are made beginning from the largest spatial aggregations down to a succession of lesser aggregations. Besides the issue of scales, it must be remembered that what are referred to as observed quantities (e.g., evaporation) are often determined through residual analysis or model assimilation. There are important data-quality issues to consider; estimates from different publications are often not independent sources of verification due to their derivation from common measurements.

In this study, the observations used to validate the model results are derived from several sources. The global (with land and ocean distinction) and continental precipitation, evaporation, and runoff estimates are derived from Henning (1989). The estimates from this recently published source are believed to be more accurate than previous ones (also cited in the source) because the revised estimates are based on methodologically improved data-assimilation procedures that yield consistent water and heat balance (Henning 1989). The precipitation figures are based on insufficient spatial sampling, especially over the oceans and dryland regions and across strong orographic gradients. The net radiation estimates that form the basis for the latent heat flux values also contain significant errors that are due to the lack of data on ground heat flux and atmospheric thermal radiative flux to the surface, among other factors. The runoff estimates are based on residual analysis and assimilation using simple water-balance models (Henning 1989). For global and continental area averages, the Henning (1989) estimates of the hydrologic fluxes are only approximate figures with wide confidence bands. The hydrologic fluxes over major river basin subdivisions are based on the work of Russell and Miller (1990), in which they estimate the components of the water balance from several recent sources and isolate the river basin yield by topographic delineation. The zonally averaged hydrologic fluxes and temperature observations are derived from Henning (1989), Zubenok (1970), and Legates and Willmott (1990). The latter dataset is considered to be up-to-date and accurate since it is based on direct measurements over an extensive network.
2. Parameterization of land-surface hydrology

The land-surface hydrology parameterizations of the GISS GCM are improved in the areas of 1) the surface fluxes of runoff and evapotranspiration and 2) the diffusion of moisture in the soil column. The following sections review the current Hansen et al. (1983) GISS GCM parameterization and the improved parameterizations for the runoff and evapotranspiration flux at the surface.

a. Spatial variability parameterization of surface hydrology

1) Surface Runoff

Surface runoff $Q$ in the GISS GCM is taken to be a fraction $R$ (the runoff ratio) of the model-generated precipitation intensity $P$ as

$$ Q = RP \quad 0 < R < 1. \quad (1) $$

The value of $R$ in the GISS GCM is taken to be one-half of the value of the relative soil saturation $s$ of the surface layer as

$$ R = \begin{cases} \frac{1}{2} s & 0 \leq s < 1 \\ 1 & s = 1, \end{cases} \quad (2) $$

where $s$ is defined as

$$ s = \frac{w}{w_{sat}} \quad (3) $$

and $w$ is the soil moisture mass per unit area in the top (surface) soil layer. Any infiltrating water that would exceed soil field capacity $w_{sat}$ is also drained as runoff.

Entekhabi and Eagleson (1989a) developed an advanced runoff parameterization by 1) introducing subgrid-scale spatial variability over the large GCM grid areas and 2) using a physically based equation of infiltration. The following is a review of their model of $R$; further detail is available in the reference.

Surface runoff generation is essentially governed by the interaction of two key processes with significant spatial variability at small scales: 1) the precipitation intensity and 2) the soil infiltration capacity. The infiltration rate depends on the saturation state of the near-surface soil. Entekhabi and Eagleson (1989a) model the large-area runoff response as the interaction of the precipitation intensity and surface relative soil saturation random fields. The mean of these fields corresponds to the grid precipitation and grid soil-water prognostics in the GCM. Following observational evidence (e.g., Eagleson et al. 1987), Entekhabi and Eagleson (1989a) consider the precipitation intensity over a fraction $\kappa$ of the grid as exponentially distributed. The near-surface soil saturation is taken to be distributed according to the gamma probability function. This distribution is characterized by its mean, which is updated for the grid according to water balance, and a dimensionless coefficient of variation, $\text{cv}$. Based on observations, Wetzel and Chang (1988, their Fig. 3), among others, show that this coefficient of variation is dependent on the area that is considered. For areas approaching the grid resolution of GCMs, $\text{cv}$ approaches unity. This value is used in the simulation experiments reported in this paper. The infiltration rate of the soil, $f^*$, is defined by the Darcy equation of vertical steady flow into porous media.

Runoff results from both infiltration-excess and partial area (rainfall over impermeable and saturated surfaces) mechanisms. The grid-mean runoff rate is determined by integrating the contribution, weighted by probability mass, of all points over the grid. In the GCM, the grid-mean value of soil moisture is updated at every time step based on the grid-mean precipitation forcing and the runoff ratio based on subgrid-scale spatial variability.

At any location, runoff is generated by

Point surface runoff $Q$

$$ = \text{infiltration excess} (P - f^* \text{ for } P > f^* \text{ and } s < 1) $$

$$ + \text{partial area} (P \text{ for } s \geq 1). \quad (4) $$

Using the subgrid spatial descriptions of rainfall and soil moisture, Entekhabi and Eagleson (1989a) derive the closed-form expression for GCM grid-mean runoff ratio as

$$ R = 1 - \frac{\gamma(\alpha, \kappa \theta/\theta) - \gamma(\alpha, \kappa \theta E(s)/s)}{\Gamma(\alpha)} + \frac{\kappa \theta E(s)}{\alpha + 1} \frac{\gamma(\alpha, \kappa \theta E(s)/s)}{\Gamma(\alpha)}, \quad (5) $$

where $\gamma[\cdot, \cdot]$ and $\Gamma[\cdot]$ are incomplete and complete gamma functions, $\nu$ is a soil hydraulic parameter resulting from a physically based infiltration function, and $E[\cdot]$ is a spatial expectation operator. The parameter $\alpha$ is the inverse-square of the dimensionless spatial coefficient of variation, $cv$.

The dimensionless parameter $I$ in (5) is

$$ I = \frac{K_{sat}}{E(P)}, \quad (6) $$

where $K_{sat}$ is the soil hydraulic conductivity at saturation and $E(P)$ is the grid-mean precipitation intensity generated by GCM atmospheric computations.

The behavior of $R$ as a function of $E(s)$ and $E(P)$ is plotted in Fig. 1 for the case of a light textured soil [$K_{sat} = 7 \times 10^{-3}$ in (m h$^{-1}$), with $cv = 1$ and $\kappa = .6$ and .1 for a typical range of $E(P)$]. As a comparison, the values of $R$ based on the GISS parameterization (2) are shown in these plots as dotted lines.

The runoff ratio in the GISS formulation is independent of precipitation intensity (Fig. 1). The Ente-
two scales of spatial variation to consider. First, there is the size of a typical storm area relative to GCM grid area. Second, there is the fraction within the storm itself, which is actually wetted by rainfall (e.g., there are generally rain-free gaps in mesoscale storm bands). Observational analysis of airmass thunderstorms over Arizona and the Sudan (Eaglestone et al. 1987) suggests this value may be roughly 0.60 in the mean for moist-convective events. The fraction, $\kappa$, then, ought to be determined based on factors related to these interactions. In more advanced applications, $\kappa$ would be determined by the fraction of the model air column mass that is experiencing convection. In Kuo-type moist convection schemes, this fraction is variable depending on the environmental conditions. In ongoing studies, $\kappa$ is also estimated from hourly observed precipitation for the climates of the continental United States over grid areas comparable to GCMs (Johnson et al. 1991).

As a simple approximation, one value of $\kappa$ is prescribed for supersaturation and one value for moist-convective rainfall. Due to the large spatial scales associated with supersaturation events, a value of $\kappa = 1.0$ is used for this type of precipitation in all GCM simulations. For moist-convective rainfall, in the first few simulations Entekhabi and Eaglestone (1989a) are followed by using $\kappa = 0.60$ (value corresponding to area wetted inside mesoscale bands). In light of the previous discussion, however, it is possible that $\kappa$ ought to be set at a much lower value. To test this sensitivity, $\kappa$ for moist-convective rainfall is set to 0.15 in a final simulation run with the GCM (value corresponding to area wetted by mesoscale bands times 0.25, indicating the relative scale of convective storms to typical GISS GCM grid size).

2) EVAPOTRANSPIRATION

Evaporation $e$ from the soil is taken as a fraction $\beta$ of the model updated potential evaporation $e_p$ (evaporation rate under conditions of unlimited moisture supply or the energy-limited evaporation rate) as

$$e = \beta e_p \quad 0 < \beta < 1.$$  \hfill (7)

In the GISS GCM, the value of $\beta$ is parameterized as

$$\beta = s \quad 0 < s < 1.$$  \hfill (8)

Entekhabi and Eaglestone (1989a) formulated an advanced parameterization of evaporation based on subgrid-scale spatial variability and a physically based exfiltration function (Philip 1957). The parameter $\beta$ in (7) is separated into two components: $\beta_s$ for bare-soil evaporation and $\beta_v$ for transpiration over vegetated areas.

Defining a mean desorptivity $f_e$ based on soil hydraulics, $\beta_s$ is obtained following a derived distribution (Entekhabi and Eaglestone 1989a). Whenever $f_e$ is less than the potential evaporation rate $e_p$, the value of $f_e$ is the evaporation rate. This is the “soil-controlled” evaporation regime. For $f_e$ greater than $e_p$, the value of $e_p$ determines the evaporation rate; this is the “cli-
mate-controlled" evaporation regime. A transitional value of relative surface soil saturation $s^*$ is defined for which the value of $f_e$ and $e_p$ are equal (i.e., $f_e = e_p$).

Using these definitions, the expected value of area evaporation or the grid-evaporation rate is

$$E(e) = e_p \int_{s^*}^{\infty} f_e(s) ds + \int_0^{s^*} f_e f_s(s) ds,$$  

(9)

where $f_s(s)$ is the spatial probability density function of relative soil moisture. Entekhabi and Eagleson (1989a) define the dimensionless soil-climate parameter

$$\beta = \frac{E(s)}{s^*} = \frac{E(s)}{s^*} \left( \frac{K_{sat} \Omega}{e_p} \right)^{2m/14+4m},$$  

(10)

and write the grid-mean bare-soil evaporation efficiency $\beta_\delta$ as

$$\beta_\delta = \frac{E(e)}{e_p} = \frac{\Omega \gamma \left( \frac{1}{2m} + 2 + \alpha, \alpha \epsilon^{-1} \right) - \frac{1}{2} \gamma \left( \frac{2}{m} + 3 + \alpha, \alpha \epsilon^{-1} \right)}{\Omega \left( \alpha \epsilon^{-1} \right)^{1/2m} + 2 - \frac{1}{2} \left( \alpha \epsilon^{-1} \right)^{2m/14+4m} + 1 - \frac{\gamma(\alpha, \alpha \epsilon^{-1})}{\Gamma(\alpha)}},$$  

(11)

where

$$\Omega = \frac{8n \Psi_{sat}}{3K_{sat} T(1 + 3)(1 + 4m)} \left[ \frac{\alpha}{E(s)} \right]^{1/2m+1}.$$  

Here $n$ is the soil porosity, and $m$ is the pore-size distribution index for the given soil type as defined in the Brooks–Corey formulation of unsaturated soil hydraulic properties (Brooks and Corey 1966). Term $\Psi_{sat}$ is the saturated soil matrix suction parameter, and $T$ is the model time step.

The derivation of $\beta_\delta$ is similar to $\beta_\delta$ except that a soil moisture extraction function by roots $e_r$ takes the place of $E(e)$. As a simplification of a complex process, $e_r$ is taken as a linear function of $s$, where $e_r$ ranges from zero at the plant-wilting point $s_w$ to an upper limit of $e_{p}$ for values of $s$ greater than the transitional value $s^*$.

The derived distribution results in the grid-mean transpiration efficiency:

$$\beta_v = \frac{E(e_v)}{e_p} = 1 + \frac{\gamma(\alpha + 1, \alpha \epsilon^{-1}) - \alpha \epsilon^{-1} \gamma(\alpha, \alpha \epsilon^{-1}) - \gamma(\alpha + 1, \alpha \epsilon^{-1}) + \alpha \epsilon^{-1} \gamma(\alpha, \alpha \epsilon^{-1})}{\Gamma(\alpha)(\alpha \epsilon^{-1} - \alpha \epsilon^{-1})},$$  

(13)

mean of the spatial distribution. This approach to incorporating spatial variability and physics-based equations into GCM land-surface parameterizations is considerably more efficient (in terms of computation) and significantly more parsimonious (in terms of the number of parameters) than subdividing GCM grids into smaller units, each with an independent hydrologic balance that needs to be numerically aggregated.

The Entekhabi and Eagleson (1989a) parameterization for the runoff ratio, $R$, bare-soil evaporation efficiency, $\beta_v$, and the transpiration efficiency, $\beta_v$, require input parameters on 1) the soil type ($K_{sat}, \Psi_{sat}, n, m$), 2) vegetation type (willow point), 3) soil-saturation spatial coefficient of variation ($c_v$), and 4) fractional storm coverage ($\alpha$). The forcing parameters, 1) the mean-grid precipitation $E[P]$ and the mean-grid potential evaporation $e_{p}$, are provided at each model time-step by the moist-convection/large-scale condensation and boundary-layer modules in the GCM.

b. Soil moisture vertical diffusion parameterization

A finite-difference approximation to the standard Richards equation for soil moisture diffusion, including sink terms for water extraction by plant roots, is also considered for representing seasonal deep-moisture storage (Abramopoulos et al. 1988). The amplitude of the annual heat and moisture balance at the land sur-
choice of three layers is also the most appropriate given the share of the total GCM computer resource needs that ought to be devoted to the land surface.

In order to test the model climate response to soil diffusion and seasonal storage of heat and moisture, the Entekhabi and Eagleson (1989b) simple climate model is used to screen the various basic sensitivities. This exercise has proved valuable in that sensitivities have been identified and quantified that would be rather difficult and improbable to isolate in the full three-dimensional GCM.

In appendix A, the results of these sensitivity experiments with a reduced and simplified climate model are presented. The simple climate model has all the major interactions and components available to the GCM, yet it is considerably more efficient. The simple model allows us to perform long-duration simulations, but more importantly, it allows us to isolate the response of the climate to a land-surface change. In a full three-dimensional GCM, the individual response of one land-surface grid to a change in one of its parameters may be masked or lost to the great many other factors that could alter the diagnostics.

One of the important interactions that has been isolated using this screening approach is the strong sensitivity of the surface heat and moisture balance regime (both seasonal and diurnal) to the total soil depth and the top-soil layer thickness (appendix A). The annual range in surface temperature and moisture status is inversely related to the total soil column depth. The sensitivity is reduced when the soil depth is larger than the penetration reach of the annual heat wave in the soil column (about one meter), as evident in the results of appendix A. More interesting, however, is the sensitivity of the mean surface moisture and surface temperature to the amplitude of the regime and the total soil column depth. Another important observation evident in the analysis contained in appendix A is that the diurnal range in temperature and the mean annual temperature and saturation are so strongly dependent on the choice of the topmost soil-layer thickness.

That the mean climate conditions are sensitive to the amplitude of the diurnal and annual heat and moisture regime is due to the presence of thresholds in the system (for example, in the moist convection scheme). The simple model is used to test reasonable values for soil thickness and layering, as well as other basic parameters such as the vertical distribution of plant roots (see appendix A). The determination of total soil-layer thicknesses was in part based on the field capacities used in the GISS model. Porosities based on the weighted soil textures were used to scale the total depth of the column such that the total field capacity remained unchanged from the GISS GCM. Of the three layers, the top layer is 10 cm and the middle layer is 15 cm. The bottom layer then makes up the difference such that the total field capacity for each grid remains unchanged from the original GISS values.
giving a total soil depth that varies over the earth from 50 to 225 cm.

In the field, the fraction of roots in different soil layers is dependent on many parameters including plant type, soil texture, and climatic regime. It is not the purpose of the parameterization of $\beta$ (13) to provide a full treatment of plant physiology. Rather, we sought to give a means of extracting water from lower soil layers to give a simple approximation to the function of plants in regional hydrologic balance. Based on observations compiled by Epstein (1973), it seems reasonable that for many plants the root density in the soil decreases exponentially with depth. Seeking to find a reasonable value for root densities, the simple climate model of Entekhabi and Eagleson (1989b) was utilized to test sensitivity to various weighting schemes (see appendix A). A weighting of 85% of plant roots in the top layer, 10% in the second layer, and 5% in the lowest layer was concluded to be a reasonable estimate.

Finally, it is pointed out that there are shortcomings of the current GISS soil moisture diffusion that will be improved by including realistic soil diffusion (Abramopoulos et al. 1988). Most important is the replacement of the current simplified mass balance equations with physically based equations, yielding more accurate seasonal cycles of heat and moisture balance. In the current GISS GCM diffusion scheme, for example, there is an instantaneous equilibrium of relative soil saturation in both of the two soil layers for vegetated regions when diffusion is in the upward direction (Hansen et al. 1983). This behavior is evident in Fig. 3 (top panel), where, at the initial time, the top layer (layer 1) is assigned to zero relative soil saturation and the second layer is considered saturated. According to the current GISS diffusion scheme, the two instantaneously equilibrate to an intermediate value between saturation and dry condition. Such instantaneous response is unrealistic and processes that depend on this scheme, such as the amplitude of the annual heat and moisture balance, are strongly affected. A more realistic situation would be to distribute moisture in the column according to the diffusion equation discretized with a finite number of layers (three layers in this application). Using the Abramopoulos et al. (1988) scheme, the same boundary conditions (dry topmost layer and saturated lower layers) were allowed to equilibrate. The results shown in the bottom panel of Fig. 3 illustrate the gradual response of the soil column to a state in which there is a hydrostatic equilibrium in the soil-moisture profile. The time constant associated with the response of the column to surface forcing in the two schemes of Fig. 3 will introduce major changes in the surface heat and moisture balance regime of the GCM.

3. Results of simulations with the GISS GCM

a. Simulation descriptions

The implementation of the advanced land-surface hydrology schemes into the GISS GCM is performed in a stepwise fashion in order to isolate the marginal changes due to each scheme. Four simulations with the three-dimensional GISS GCM will be compared as listed in Table 1. The numerical experiments will use the $8^\circ \times 10^\circ$ GISS GCM with seasonally fixed climate sea surface temperatures. The seasonally fixed ocean temperatures are desirable when performing land-surface hydrologic sensitivity tests. Low-frequency transients introduced by a mixed-layer ocean or a coupled circulation model may complicate the validation of the model climate. Run G-0 is the control case with the GISS-II GCM (Hansen et al. 1983). Run G-I differs from G-0 only due to the replacement of the Hansen et al. (1983) empirical runoff and evapotranspiration efficiency functions with the Entekhabi and Eagleson (1989a) expressions. Run G-2 is further modified to include the Abramopoulos et al. (1988) soil moisture
TABLE 1. Simulations performed with the GISS GCM.

<table>
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<tr>
<th>Simulation name</th>
<th>Description</th>
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<tr>
<td>G-0</td>
<td>GISS Model II as described in Hansen et al. (1983) using $8 \times 10$ degree grid resolution and fixed ocean temperatures (control run).</td>
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<tr>
<td>G-1</td>
<td>Same as G-0 except for new formulation of runoff coefficient $R$, bare-soil evaporation efficiencies, $\beta_v$, and transpiration efficiency, $\beta_t$ (after Entekhabi and Eagleson 1989a). Fractional wetting parameter $\kappa$ set equal to 1.0 for large-scale supersaturation rainfall and 0.6 for moist-convective rainfall (abbreviated &quot;space&quot;).</td>
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<tr>
<td>G-2</td>
<td>Same as G-1 except new soil moisture diffusion scheme of Abramopoulos et al. (1988) is used with transpiration from lower layers. No instantaneous upward diffusion or prescribed growing season as in GISS GCM (abbreviated &quot;space/soil&quot;).</td>
</tr>
<tr>
<td>G-3</td>
<td>Same as G-2 except fractional wetting parameter $\kappa$ for moist-convective rainfall set to 0.15 (abbreviated &quot;space/soil/storm&quot;).</td>
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Diffusion scheme. The final simulation experiment G-3 differs from G-2 only by modifying the fractional wetting parameter $\kappa$ to reflect conditions expected over the GISS GCM discretization. The fractional wetting by precipitation in run G-3 is $\kappa = 0.15$, assuming that the average convective storms cover one-fourth of the $8 \times 10$ GCM grid and, of that area, the mesoscale structure delivers rain to only 60% of the storm-affected region (Eagleson et al. 1987).

The precipitation forcing of the new land-surface module differentiates between the grid precipitation due to the large-scale condensation (resulting from supersaturation) and the moist convection (see Hansen et al. 1983 for description). The surface flux of heat, mass, and momentum are computed using the surface-layer parameterization of GISS-II GCM (Hansen et al. 1983). In this module, the flux of sensible and latent heat are iteratively solved such that the flux from the surface to a height of about 30 m above the surface equals the flux between the 30-m level and the lowest atmosphere level of the GCM. The transfer coefficients are corrected for stability (Hansen et al. 1983). The latent heat flux from the surface is the product of the potential evaporation rate, the evapotranspiration efficiency functions ($\beta_v$ and $\beta_t$ weighted by the vegetation canopy coverage and land fraction of the grid), and the specific latent heat of vaporization. The potential rate is determined by assuming saturated conditions at surface boundary. The temperature defining this saturation level is taken as the ground temperature similar to the GISS-II treatment. Milly (1992) raises important concerns about the consistency of the use of the ground temperature in estimating $e_p$ and the evapotranspiration efficiency function of the type corresponding to (8).

The snowmelt and ground-freezing formulation is identical in all experiments in lieu of better parameterizations and in order to isolate the marginal changes due to replacing the runoff and evapotranspiration models. Improvements in the modeling of snow and ice given subgrid-scale spatial variability are currently under research.

Each simulation is for at least 4 years; diagnostics are collected over the last 3 years. Initial conditions for each of the simulations are taken from an earlier standard GISS Model II $5^\circ \times 10^\circ$ experiment.

b. Model land-surface boundary conditions

The GISS GCM land-surface boundary conditions are based on the archived $1^\circ \times 1^\circ$ vegetation dataset of Matthews (1983) and the archived $1^\circ \times 1^\circ$ soil texture dataset of Zobler (1986). Eight vegetation types were defined by Matthews (1983), and each vegetation type was associated with a value for albedo, roughness length, "masking depth" (a variable describing the effect of vegetation on snow albedo), and field capacity. The values of the parameters used over a particular grid in the model are weighted averages based on the percentage of each vegetation type over that grid.

Soil textures (given by percentage of sand, silt, and clay) are assumed to have no variation in the vertical dimension. From the soil textures, values of the soil hydraulic properties are taken as weighted averages over the three soil types. Entekhabi and Eagleson (1989a) give the soil hydraulic properties in each textural class.

c. Global hydrologic balance

In the analysis of the simulation results and their comparison with the available observed data a progression is followed from the very largest scale comparisons (globally averaged quantities), to more detailed (zonal averaged and continentally averaged

**Observations**

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**FIG. 4.** Global hydrologic balance (in centimeters per year) for all simulations with observations from Henning (1989).
quantities), and finally to mapped distributions over the earth.

The most basic picture of hydrologic performance in the model is the globally averaged hydrologic cycle. There are several sources of compiled values for the globally averaged precipitation and evaporation over the oceans and land surface and the exchange of moisture between those two regions (Baumgartner and Reichel 1975; Budyko 1978; Henning 1989). In the long-term mean this latter exchange must equal the runoff from the land. As a comparison, a schematic of the global hydrologic balance is plotted and the simulation values are tabulated below the one set of observed values (Henning 1989) in Fig. 4. The values are all in annual depths [centimeters] per unit area, and the two exchange values (atmospheric divergence over land and runoff) are normalized by total land surface area. Run G-3 is slightly out of balance due to incomplete drainage of initial storage from the land surface. Because of this, it is expected that the runoff value in run G-3 is slightly higher than that which would result from a longer-duration simulation. The ocean-evaporation value is essentially constant for all runs due to the use of prescribed ocean-surface temperatures.

First, notice that the GISS model (control run) has underestimated runoff and overestimated both precipitation and evaporation over the global land surface (Fig. 4). The range of values of global land-surface runoff based on the available observations cited earlier is from 27 to 35 cm annually. The GISS value does not fall in this range. Further, neither the GISS precipitation over land nor evaporation over land fall into the ranges of the observed for these quantities (precipitation over land observations range from 72 to 80 cm annually, whereas GISS gives 88 cm; evaporation over land observations range from 41 to 48 cm annually, whereas GISS generates 65 cm, roughly 50% more than estimated based on observed data). Similarly, runs G-1 and G-2 grossly overestimate land-surface evaporation and underestimate runoff.

Major improvements occurred with run G-3 (with scaled fractional wetting parameter of moist convective events to match the GISS GCM resolution). Here precipitation over land and evaporation over land are seen falling within their observed ranges. Global runoff is slightly high, but as mentioned earlier, the runoff value will likely decrease as soil storages are depleted to their equilibrium states. The land-surface evaporation in this model climate is comparable to the observed value.

d. Zonal hydrologic balance

Figures 5a–c show the zonally averaged values of precipitation, evaporation, and surface runoff over the

FIG. 5. (a) Zonally averaged annual land-surface precipitation. (b) Zonally averaged annual land-surface evaporation. (c) Zonally averaged annual surface runoff.
land surface for each of the simulations, as well as observed values compiled by Zubenok (1970) and Henning (1989). Although Hansen et al. (1983) plotted zonally averaged precipitation and evaporation and found very good agreement with observations, their comparison included the oceans. Since two-thirds of the earth is covered by oceans and the hydrologic fluxes over oceans are considerably larger than over land, the contribution of the land-surface alone was masked. Here it is seen that the zonally averaged hydrology over land surfaces in the GISS Model II is actually in rather poor agreement with data (run G-0). It should be noted that in zonally averaged plots, each latitude belt is given equal weight. In reality, however, the actual fluxes depend on the amount of land surface in the latitude belt. Since most of the land mass over the earth is in the Northern Hemisphere, it must be kept in mind that errors in zonal plots should be interpreted accordingly. In latitude bands where land comprises a small fraction of the total area (e.g., around 50° to 70° south), the estimates will have larger variance.

As presented here, it is seen that in general, precipitation (Fig. 5a) is modeled quite well in the tropics, but agreement with observations deteriorates in the middle and high latitudes. Particularly in the Northern Hemisphere, midlatitudes (where data are generally more readily available) precipitation is overestimated. The only strong deviation from the control run is run G-3 in the tropics. Here it is seen that although run G-3 had a very good globally averaged precipitation over land, it is actually underestimating in the tropics and overestimating in middle and high latitudes.

In Fig. 5b, the simulations G-1 and G-2 are essentially equivalent to the control (run G-0) in the zonal evaporation over land. Run G-3, however, gives a remarkable agreement with observations in the tropics. In Fig. 5c, the same pattern holds: runoff in run G-3 comes much closer to the observed values in the tropics. The spike at 50° south in the observed runoff is due to the fact that there is very little land in this belt.

The effects of the reduced fractional wetting in moist convective rainfall in run G-3 are strongest by far in the tropics. The humid tropics have a high percentage of moist-convective and intense rainfall. As evident in Fig. 1, the runoff ratio is sensitive to the intensity of the precipitation.

The zonally averaged land-surface figures indicate that the model climate is improved for the GISS GCM equipped with the Entekhabi and Eagleson (1989a) land-surface hydrology parameterization with subgrid-scale spatial variability and the Abramopoulos et al. (1988) soil-diffusion scheme. In particular, the improvement of the evaporation on all scales—global, zonal, and as will be seen, continental and river basin scales—is emphasized.

e. Continental water balance

As the next step in our progression to finer spatial detail, the water balance over the continents is considered. Shown in Fig. 6 are the precipitation, evaporation, and surface runoff over continents (excluding Antarctica) for the reduced set G-0 and G-3 simulations, as well as observations by Henning (1989). In moving to comparisons of greater detail, only the essential control (G-0) and modified (G-3) simulation experiments are

![Continental Precipitation](image)

![Continental Evaporation](image)

![Continental Runoff](image)

*Fig. 6. Annual per unit area (a) precipitation, (b) evaporation, and (c) runoff for runs G-0 and G-3 over the continents.*
retained. In run G-3, there are major changes on all fronts. On all continents, runoff was increased, while precipitation and evaporation were decreased. By comparison to observations, this was a significant improvement for all continents for evaporation, and an improvement was seen in all but South America for precipitation. The runoff was only improved for two continents (South America and Africa), however, while the other four are overestimated. The problem with runoff over continents is not solely a result of the land-surface parameterization but is partly due to there being an excess of precipitation in some key regions.

f. Major river basin water balance

As a further step in evaluating the model hydrologic performance, the runoff over major river drainage systems will be compared to observations. Russell and Miller (1990) are followed in discretizing the GISS model into 33 river basins around the earth. In their paper, a $4^\circ \times 5^\circ$ grid spacing was used in the model, as opposed to an $8^\circ \times 10^\circ$ spacing used here. Further, their results were plotted by volume, whereas here results are plotted per unit area values. The reason for this shift is that the modifications made act on a per unit area basis over grid squares, and the pattern of the effects of the modifications is more readily apparent in this form. Furthermore, ranking according to per unit area runoff helps in isolating the changes according to climatic type.

Figure 7 shows the runoff of the control run G-0, run G-3, and observations plotted in descending order of per unit area runoff; the first river basins shown are thus generally the most humid ones and the ones with the highest runoff ratios. This is where improvement of run G-3 is most evident.

In some cases, such as the Yellow River, the extreme runoff values are due to a severe model overestimation of precipitation. The observed value of runoff for this narrow basin is also prone to underestimation due to overbank flooding at times of high flow.

The striking improvement due to run G-3 apparent in the tropical river basins depicted in Fig. 7 confirms the improvements seen earlier in the zonally averaged runoff plot for tropical regions. The GISS coarse-resolution GCM generally overestimates the precipitation over the world's arid and semiarid climates; the fraction of the precipitation that is partitioned to runoff is accordingly overestimated.

g. Heat balance

The hydrologic balance over land-surfaces has a strong effect on the regional heat balance. The mechanisms include: 1) atmospheric transport of latent heat, 2) the radiative absorption and emittance by atmospheric water vapor, 3) the reflective properties of clouds, 4) the reflective properties of snow on the earth's surface, 5) the increase of soil heat capacity due to the presence of soil moisture, and 6) the release of latent heat from the surface by evaporation. There are also secondary effects such as those due to the dependence of vegetation characteristics (transpiration and roughness) and the influence of the heat budget on wind patterns. Of all these mechanisms, it is the evaporation that has the strongest effect on the surface temperature; it is a major cooling mechanism of the land surface when moisture is available. As moisture in the soil is depleted, the heat partitioning at the surface shifts toward higher (relative) rates of sensible heat flux. Since the sensible heat flux is a much less efficient transfer process, temperatures at the surface may rise signifi-
Annual Mean Temperature
(Landsurface Only)

Monthly Runoff Ratio
Central Argentina Grid [Control Run (G-0)]

Latitude

DJF Mean Temperature
(Landsurface Only)

Runoff Ratio
(runoff/precipitation)

G-0 (control)
G-1 (space)
G-2 (space/soil)
G-3 (space/soil/storm)
Observations (Legates & Willmott 1990)

Relative Surface Soil Saturation

G-0 (control)
G-1 (space)
G-2 (space/soil)
G-3 (space/soil/storm)
Observations (Legates & Willmott 1990)

Monthly Runoff Ratio
Central Argentina Grid [Run G-3 (sp/so/st)]

Latitude

JJA Mean Temperature
(Landsurface Only)

Runoff Ratio
(runoff/precipitation)

G-0 (control)
G-1 (space)
G-2 (space/soil)
G-3 (space/soil/storm)
Observations (Legates & Willmott 1990)

Relative Surface Soil Saturation

Fig. 9. Comparison of (a) control run G-0 and (b) run G-3 (space/soil/storm) of the monthly runoff ratio over a Central Argentina grid.

cantly under reductions in soil moisture. Comparisons of model heat balance with data are given here for surface air temperature only since it is a result of all heat balance components and has been measured most extensively and accurately.

Figures 8a–c show zonal mean surface air temperatures over the land-surface (oceans excluded) for all simulations along with observations compiled by Le-
The runoff ratio (the ratio of runoff to precipitation) at a point may be averaged over time and plotted against the average relative surface soil moisture. Figure 9 shows monthly averaged runoff ratios in both the control run G-0 and run G-3 over a central Argentina grid square chosen for its large seasonal range of soil-moisture values. It is seen that the plot for the control run G-0 deviates only slightly from the empirical $R = \frac{1}{s}$ line, which is prescribed in the model (2). This is due to ponding on the surface (giving a slightly higher value of $R$) and also to the surface storage of snow (giving a slightly lower value of $R$). The GISS scheme has no dependence on precipitation intensity, soil type, or spatial characteristics but only on relative surface soil moisture.

The bottom plot of Fig. 9 shows the runoff ratio of the same grid square but for run G-3. Immediately obvious is the nonlinear relationship between $R$ with $s$. High values of $s$ result in runoff ratios reaching 0.7 in the monthly mean. Actual soil moisture data are unavailable for this region (as it is in general). The important point is that the runoff ratio in run G-3, with the effects of spatial variability and realistic infiltration dynamics, exhibits a nonlinear relationship with soil saturation.

When various hydrologic fluxes are plotted for all land-surface grids, patterns of direction and magnitude of changes may be seen across the globe. Figure 10 gives the distribution of soil saturation versus precipitation intensity. Figures 11 through 13 illustrate the difference between run G-3 and the control run of the diagnostics of runoff and evaporation over land-surface grids plotted against relative soil moisture and precipitation intensity.

Figure 10 shows that in both G-0 and G-3 (but more consistently in G-3) the topmost soil layer is dry for grids that have high precipitation intensities (climatic precipitation rate divided by the fraction of time with precipitation). Figures 11 and 12 compare the differ-

gates and Willmott (1990). Figure 8a is the annual mean, while Figs. 8b and c are the winter and summer means, respectively. As can be seen, all simulations are essentially the same except in the tropics. Here run G-3 diverges from the other three, ranging from $0^\circ$ to $2^\circ$ warmer. In the annual mean, it is seen that run G-3 is in exceptionally good agreement with data from $20^\circ$ south to $20^\circ$ north.

h. Grid-by-grid comparison of surface hydrology

The fundamental changes introduced by the new hydrology parameterizations may be further diagnosed by the examination of individual land-surface grids. Here diagnostics are considered of single points in space averaged over increments of time, as well as distributions of diagnostics of all land-surface grids averaged over the entire simulation period.

![Fig. 11. Grid-by-grid plot of the difference in runoff produced in run G-3 (space/soil/storm) and the control run G-0, plotted against precipitation intensity for conditions in run G-3.](image-url)
The statistical-dynamical parameterization of Entekhabi and Eagleson (1989a) is used for its critical advantage over current GCM hydrological schemes due to the inclusion of subgrid-scale spatial variability. The soil-moisture diffusion parameterization of Abramopoulos et al. (1988) is used to provide more realistic deep soil-water storage and moisture diffusion. These parameterizations are both physically based in contrast with the current empirical land-surface hydrological parameterization of the GISS GCM. The additional computation requirements increase the simulation CPU usage by 25% with inclusion of both parameterizations (10% for the spatial variability parameterization alone). It is important to note that these algorithms may be further optimized in their coding to reduce their cost.

The results shown in section 3 demonstrate the major improvements in the hydrologic balance resulting from the inclusion of spatial variability. Results are compared over a wide range of spatial domains (global land surface, zonal land surface, continental, and river basins) using a number of datasets, and improvements are verified on all fronts. Because of the nonlinear response of runoff to soil moisture and precipitation intensity when spatial variability is included, the strongest improvements in hydrologic budgets occur over the tropics. Improvement in the hydrologic balance further results in improved heat balance verified most distinctly in comparisons of zonal surface air temperature over land-surface areas. A physically based soil moisture and heat diffusion scheme is necessary for maintaining realistic annual cycle amplitudes of heat and moisture content.

Improvements in the land-surface hydrologic balance particularly over the tropics are obtained by the inclusion of spatial variability. The poor agreement of the current $8^\circ \times 10^\circ$ GISS Model II hydrologic balance (G-0) can be seen to stem mainly from a lack of runoff generation, giving simulated values much lower than those observed over land-surface areas. This allows evaporative fluxes to exceed by a large margin the evaporation values derived from observations of hydrologic balance. Based on this research, it seems most likely that the lack of runoff generation in the current GISS Model II in the tropics is due to the fact that precipitation is currently modeled as being uniform over the entire grid square. This results in low-average intensities that generate far less runoff (if physically based infiltration equations are employed) than would be obtained by spatially heterogeneous rainfall, having some areas of concentrated rainfall producing larger amounts of runoff and some areas of lesser rainfall intensity producing less runoff. This, coupled with spatial variability in soil moisture in the formulation of Entekhabi and Eagleson (1989a), has been shown here to produce results in better agreement with observations. Regardless of the levels of detail that may be pursued in modeling land-surface hydrology, however, without the element of spatial variability, it seems un-
likely that the global hydrologic balance, with its regional differences in flux partitioning, will be represented adequately in GCMs.

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APPENDIX A

Determination of Soil Column Specifications

The computational expense associated with sensitivity experiments with the full three-dimensional GCM necessitates the use of a simpler climate model to screen the many possible simulations. With screening, comprehensive sensitivity testing is reduced to a few well-defined and focused experiments for the three-dimensional GCM. Another important reason to use a simple climate model is that diagnostics collected from simulation experiments in the full GCM are affected by an enormous number of physical and spatial interactions in the model climate. It is often difficult to isolate the marginal response of the model to a changing parameter. Furthermore, due to the expense, long-duration simulations are prohibitive with the full GCM. Therefore, a simple climate model, as long as it preserves the basic hydrologic and thermodynamic interactions and feedbacks of the full three-dimensional GCM, is a rather valuable tool in screening sensitivity experiments.

The first use of such simplified models for sensitivity analyses may be traced to the independently developed schemes of Koster and Eagleson (1990) and Warrillow et al. (1986). The model used here is based on the work of Entekhabi and Eagleson (1989b), which is an extension of the model originally proposed by Koster and Eagleson (1990). The reader is referred to Johnson et al. (1991) and Entekhabi and Eagleson (1989b) for the technical specifications of the scheme used in this appendix.

GCM physics, such as radiative transfer, moist convection, and large-scale condensation, are all one-dimensional processes. They are also the key areas of hydrologic and thermodynamic coupling in the model climate. They are identically included in the simplified climate model as in a full GCM that has discretized atmospheric profiles over each surface type. The only difference between the GCM and the simple climate model is the parameterization of heat and mass convergence in the model atmosphere. The GCM solves the equations of motion and mass transfer over a discrete grid, whereas the simple models (Warrillow et al. 1986; Koster and Eagleson 1990; Entekhabi and Eagleson 1989b) parameterize the convergence rates. The earlier models prescribed the moisture and heat convergence rates; Entekhabi and Eagleson (1989b) use a simple linear reservoir scheme for the outflow of heat and moisture between two adjacent ocean and land air columns. The advantage of this improved scheme is that the surface runoff rate (which at equilibrium exactly equals the atmospheric moisture convergence rate) is free to change if the surface parameters are modified. Thus, true hydrologic sensitivity may be tested with this simple climate model.

To test the role of soil layer thicknesses in the three-layer discretized soil moisture diffusion scheme, a "nominal" arrangement was defined with layer-1 (the top layer) thickness 10 cm, layer-2 thickness 15 cm, and layer-3 thickness 50 cm (roughly a geometric progression). Alternate cases were examined wherein the proportions of the three soil layers to each other remained the same but total depth was changed to $\frac{1}{2}X$, $2X$, and $4X$ the total nominal depth of 75 cm. A final case used a nominal top-layer thickness (10 cm) and...

![Diagram](image.png)

**Fig. A1.** (a) Mean annual and (b) diurnal ranges of surface ground temperature versus total soil-column thickness and top soil-layer thickness.
thickness is thus regulated by the depth of the diurnal penetration of the heat wave. The annual range of the surface temperature is also dependent on the magnitude of the surface heat capacity, hence total soil depth. But when this depth is increased beyond the penetration depth of the annual heat wave, the sensitivity is reduced (Fig. A1).

As shown in Fig. A2, the mean annual relative surface soil saturation and surface temperature are influenced by the top-layer thickness. The special case of the top layer having the nominal thickness and lower layers having 4X the nominal thickness behaves almost exactly as the nominal case. Because these two diagnostics are indicative of the water balance and heat balance, it is concluded that the top soil-layer thickness exerts significant control over the mean modeled conditions.

In Figs. A1 and A2, it is evident that the mean cli-
is used. Using fully vegetated conditions in 2-yr simulations, the cases of root distributions of \( / .85, .10, .05 / \) and \( / .75, .15, .10 / \) are examined for both light textured soil \( [K_{sat} = 7 \times 10^{-3} \text{ (m h}^{-1})] \) and heavy textured soil \( [K_{sat} = 1.67 \times 10^{-3} \text{ (m h}^{-1})] \). The time series of Figs. A3 and A4 show the strong sensitivity of relative saturation in the three soil layers to root distribution.

Based on these simulations, and in the absence of root distribution data for the GISS GCM, the choice was made to use a root distribution of \( / .85, .10, .05 / \) for vegetated regions over the earth. The severe drying of lower soil layers found with larger fractions of roots in lower layers is regarded as unrealistic for most settings.

**APPENDIX B**

**List of Symbols**

- \( c_v \) — Coefficient of variation (standard deviation/mean) of soil moisture (\( \cdot \))
- \( E( \cdot ) \) — Spatial expectation operator
- \( e \) — Evaporation rate \( (LT^{-1}) \)
- \( e_v \) — Potential evaporation rate \( (LT^{-1}) \)
- \( e_o \) — Transpiration rate \( (LT^{-1}) \)
- \( E \) — Combined soil–climate parameter (\( \cdot \))
- \( f( \cdot ) \) — Probability density function (\( \cdot \))
- \( f_c \) — Exfiltration rate \( (LT^{-1}) \)
- \( f^* \) — First soil-layer infiltrability \( (LT^{-1}) \)
- \( I \) — Saturated soil hydraulic conductivity ratio to grid-mean precipitation (\( \cdot \))
- \( K_{sat} \) — Saturated soil hydraulic conductivity \( (LT^{-1}) \)
- \( m \) — Soil pore-size distribution index (\( \cdot \))
- \( n \) — Effective soil porosity (\( \cdot \))
- \( P \) — Point precipitation rate \( (LT^{-1}) \)
- \( Q \) — Grid-mean runoff rate \( (LT^{-1}) \)
- \( R \) — Runoff ratio (\( \cdot \))
- \( s \) — Effective relative soil saturation (\( \cdot \))
- \( s_w \) — Relative soil saturation at vegetation wilting point (\( \cdot \))
- \( s^* \) — Transitional relative soil saturation (\( \cdot \))
- \( T \) — GCM integration time step (\( T \))
- \( w \) — Dimensionless soil parameter measuring the strength of soil capillarity \( \text{[see Entekhabi and Eagleson 1989; (13)](\cdot)} \)
- \( w_{sat} \) — Soil water storage capacity; saturation value of \( w \) \( (L) \)
- \( \psi \) — Ratio of \( E(s) \) to \( s_w \) (\( \cdot \))
- \( \alpha \) — Gamma pdf shape parameter (inverse-squared \( c_v \)) (\( \cdot \))
- \( \beta \) — Grid-mean evaporation efficiency (\( \cdot \))
- \( \beta_v \) — Grid-mean bare soil evaporation efficiency (\( \cdot \))
- \( \beta_g \) — Grid-mean vegetation transpiration efficiency (\( \cdot \))
- \( \gamma(\cdot) \) — Incomplete gamma function
- \( \Gamma(\cdot) \) — Gamma function
κ—Fraction of grid area wetted by precipitation

Ψ_{sat}—“Saturated” soil matric suction parameter (L)

Ω—Combined soil parameter (·)

REFERENCES


