Precipitation Infiltration in the Simplified SiB Land Surface Scheme

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Abstract

A global two-dimensional implementation of the simplified Simple Biosphere (SSiB) land surface scheme is integrated offline for two years as part of the Global Soil Wetness Project (GSWP). A climatology of soil wetness and surface fluxes has been produced. This climatology is compared to a number of sensitivity studies that have been performed to investigate how the partitioning of precipitation between runoff and evapotranspiration is affected, when aspects of the soil parameterization and the treatment of convective precipitation are altered.

The control integration has a reasonable spatial distribution of the surface hydrologic balance components, and shows realistic seasonal and interannual variations. Evaporation from the soil surface accounts for a majority of the water fluxes from the soil over all but heavily forested areas, where transpiration dominates. The sensitivity studies show that in general the most sensitive terms on seasonal time scales appear to be runoff, direct evaporation from the soil, and the seasonal change of water storage in the soil matrix. A realistic distribution of convective precipitation in space and time is necessary to simulate at the grid scale adequately high values of runoff, and to not over-represent direct evaporation of rainfall intercepted by the canopy. Sensitivity is found to the choice of thickness of the surface soil layer — a parameter often assigned arbitrarily in land surface models. Little sensitivity is found when imposing a vertical profile of soil porosity intended to account for surface soil aeration and deep soil compaction.

1. Introduction

Land surface models are prevalent today in the coupled climate system — models are used to examine issues of climate predictability, as well as environmental sensitivity to various anthropogenic changes. These land surface models are applied either regionally or globally, depending on the domain of the atmospheric component of the climate system model. However, the large domains of interest present a problem: complete validation of land surface models can only be conducted at single points or at relatively small plot scales where field campaigns have been conducted, such as has been done in Phase 2 of the Project for Intercomparison of Land Surface Parameterization Schemes (PILPS; Henderson-Sellers *et al.*, 1995). How can validation of land surface models be conducted globally? Remote sensing can provide large scale coverage and measurement of some quantities, but it cannot pro-

vide all of the fields needed to validate land surface models. Also, since errors and biases exist in atmospheric models, it is important to decouple the land surface from atmospheric models in order to isolate the traits of the land surface model for validation.

The Global Soil Wetness Project (GSWP; IGPO, 1995) attempts to address these issues by providing a framework and data sets to run land surface models offline globally at 1° resolution for a two-year period (1987–1988). The project is also a test bed for developing large-scale validation methods over the land surface. Central to GSWP are the global data sets made available by the International Satellite Land Surface Climatology Project (ISLSCP) Initiative I (Meeson *et al.*, 1995, Sellers *et al.*, 1996a). This collection includes global data sets of vegetation and soil properties in forms commonly used by many of today's land surface models, as well as observed and analyzed global fields of meteorological data that can be used in place of an atmospheric model as a one-way coupling to drive the land surface model.

Uncertainties still exist. Sources of uncertainty include: the variable quality and accuracy of the data

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sets used to specify soil and vegetation properties globally; shortcomings in our understanding of the physics involved at the land surface, which affect our model parameterizations; uncertainties in how to deliver to the land model the information available from meteorological observations or simulated quantities produced by atmospheric models. Each potential source of uncertainty may not have equal consequence for the accurate simulation of the heat and moisture balance of the land surface. By examining each potential source of uncertainty in turn via sensitivity studies, one can begin to understand which ambiguities are most problematic for accurate simulation of the land surface. The GSWP framework provides a ready means for conducting such sensitivity studies on a global scale.

In this study, we examine how the simulation of infiltration may be affected by the problems described above. We focus on the division of precipitation between runoff and the components of evapotranspiration in the simplified Simple Biosphere model (SSiB; Xue *et al.*, 1991; 1996) implemented in an offline, two-dimensional mode (Dirmeyer and Zeng, 1997). Specifically, we try to discover the sensitivity of infiltration to the treatment of convective precipitation (which is characterized by heterogeneous distribution within the resolvable grid size used in climate simulations) and to choices made in the land surface model regarding the vertical distribution of soil properties and layers. This study is by no means exhaustive, but is meant to provide direction for future model development.

An outline of the land surface model and its implementation in GSWP is presented in Section 2. Section 3 of this paper briefly describes the climatology of the hydrologic cycle of this model in the GSWP framework. Section 4 presents results of several sensitivity experiments. Conclusions are given in Section 5.

2. Model description

This section describes a two-dimensional offline implementation of the SSiB model of Xue *et al.* (1991). This implementation, called 2DSiB, is designed for testing and calibration of parameters and parameterizations in a regional or global framework. It is also suitable for land data assimilation, such as creating global climatologies of soil wetness. It produces time means and instantaneous values of surface fluxes, soil state variables, temperatures, water and energy budget terms on a regular grid over ice-free land points. Detailed descriptions of the standard offline version of SSiB (0DSiB) can be found in Xue *et al.* (1991, 1996). A more complete description of this two-dimensional implementation of the model, including a historical review of published research since the development of the original SiB model of Sellers *et al.* (1986), can be found in

Dirmeyer and Zeng (1997).

The current structure of the offline version of SSiB used here performs four main tasks: 1) determining the amount of radiant energy and water that will enter and interact with the canopy-soil system; 2) calculating the physical and physiological parameters used in the flux calculations; 3) calculating the fluxes of heat; 4) updating the moisture state variables. These tasks are performed at each grid point independently over the region of interest — in this case, all ice-free land areas of the globe. There are no lateral transports of heat or water in 2DSiB, since the model is designed to be integrated on spatial grid scales where the total lateral transports in the soil matrix are small compared to the total vertical transports.

There are four main loops in the driver program. The outermost loop is the *decad loop*, which is executed three times per month. In this loop an entire 10-day period of forcing data is read in, and preliminary units conversions are performed on the forcing data where necessary. At the end of this loop the instantaneous and decad mean grids of output variables are written to files. Within this loop is the *grid-point loop*, which iterates once per each ice-free land point. The first task within this loop is to set all of the SSiB vegetation and soil parameters to the correct GSWP values for the grid point under consideration. The final task is to update the running means and accumulations for that grid point before going on to the next. Within this loop is the 6-hour loop. As the name implies, this loop is invoked once every 6 hours at each grid point. The meteorological forcing data are at 6-hour intervals — this loop reads the appropriate forcing data and interpolates it to a one-hour interval (the chosen interval of integration for 2DSiB). The innermost loop is the *1-hour loop*, or more accurately the one-time-step loop. Within this loop the required forcing variables are set to the appropriately interpolated values, and the 0DSiB model described earlier is invoked.

3. Baseline simulation

3.1 Global climatology

A two-year simulation using all of the specified GSWP data sets and specifications was made to evaluate the performance of 2DSiB, and to a control case for further sensitivity studies. Global fields of soil properties (texture, albedo, surface slope, and total soil column depth), time invariant vegetation properties (vegetation type distribution), and monthly varying vegetation properties (leaf area index, greenness and green leaf fraction) are specified from the $1^\circ \times 1^\circ$ data sets on the ISLSCP Initiative I CD-ROM (Sellers *et al.*, 1996a). Near surface meteorology and surface radiation data are taken from the same source and used to drive 2DSiB for the two-year period encompassing all of 1987 and

Fig. 1. Two-year average precipitation and evapotranspiration (mm y^{-1}), and components of all water fluxes out of the soil for the same two-year period, expressed as a percentage of the mean precipitation.

1988. Details about the procedure followed by all GSWP participants may be found in IGPO (1995), and a comparison between the models is described by Sato and Nishimura (1998).

Figure 1 shows global time-mean fields of precipitation forcing, model simulated evapotranspiration, and specific components of the surface water budget which remove water from the vegetation-soil matrix. The magnitude of these components are expressed

as a percentage of the total precipitation. In this format, one can see how different terms of the water balance predominate over different regions. Total runoff is the sum of surface and subsurface runoff, and the bottom four panels sum to equal total evapotranspiration.

Surface runoff accounts for a very small percentage of total runoff. This may be due to the method of interpolating 6-hourly precipitation data adopted

for GSWP, and the use of a one-hour time step which is considerably longer than that used in climate model integrations. This point is explored in the following section. Subsurface runoff accounts for a sizeable fraction of the dispensation of precipitation over most regions of the globe. The fraction is especially high over the Amazon Basin, the monsoon regions of southern Asia and Indonesia, and much of the high latitudes.

Transpiration is a major part of total evapotranspiration over the forested midlatitudes, the tropical forests of Africa, and the southern and eastern margins of Amazonia. There is some spurious signal over the Sahara and Arabia, but mean evaporation there is extremely low (the unshaded regions over Africa have annual rainfall of less than 10 mm yr^{-1}). Transpiration is otherwise low over regions of sparse vegetation.

Interception loss, which is evaporation from rain and snow that collects on foliage, is significant over many of the same forests where transpiration is large. Interception loss is lowest over desert regions. This quantity is highly correlated with leaf area index.

Direct evaporation of moisture from the soil surface is a major component of total evapotranspiration over all but the most heavily forested areas. Over sparsely vegetated areas, soil evaporation naturally dominates. Direct snow evaporation is significant in areas that receive ample winter snow, but are not humid.

The globally high values of subsurface runoff are due to aspects of the baseflow parameterization. In SSiB, the baseflow is proportional to soil wetness of the lowest soil layer, with an effective hydraulic conductivity derived for large humid river basins. This assures that grid-average subsurface runoff is representative of the observed supply of water to major rivers (Xue *et al.*, 1996). There is no explicit water table in SSiB, which limits its soil representation to the vadose zone. In addition, a minimum rate of gravitational drainage is specified despite the value of deep soil conductivity. The value chosen ensures that a small fraction of the total water in the deepest layer of the soil drains each day, even in the driest conditions. This parameterization was instituted to ensure flow to supply river routing schemes during dry seasons. However, this approach illuminates a model inconsistency with relationships between precipitation, net radiation and evaporation developed by Budyko (1974; *cf.*, Koster *et al.*, 1998). It states that on long time scales (annual or longer), evaporation must be limited by the minimum of water supplied by precipitation and available radiative energy. Where precipitation is the limiting factor, $E \rightarrow P$ as radiation increases. Figure 2 shows the relationship between the two-year mean latent heat flux and net radiation (each scaled by precipitation) for

all grid points in the experiment. E/P does not appear to approach the asymptotic limit of unity in dry regions as predicted by Budyko. The limit is closer to 0.85. This limit is probably too low, especially for hot arid and semi-arid regions where virtually no runoff is expected. The globally constant effective hydraulic conductivity used for baseflow in SSiB is inappropriate for arid regions, and may mask real variability between river basins. When the minimum gravitational drainage term is removed, the limit increases to about 0.90.

The preponderance of points above the curve in the energy-limited region of the spectrum skew the curve for SSiB above the theoretical limit for values of $R/\lambda P$ below 0.5. Most of these points are at high latitudes where snow is prevalent. It is possible that sensible heat flux into the surface from warm air advection may be providing extra energy for evaporation. However, problems in the energy balance in the snow formulation of SSiB are known to exist, and may contribute to this effect.

3.2 Regional variability

Stark contrasts exist between the rainfall observed during the corresponding seasons of 1987 and 1988 over many parts of the globe. To investigate the effect of the observed variability of precipitation on regional water balances, we examine several large areas that had significant variations in precipitation between the two years. Figure 3 shows six areas chosen for further study. Three-month seasons were chosen for each region, as indicated by the month initials in parentheses. The irregular shapes of the regions are the result of our selection criteria. In each area, grid boxes where the difference in seasonal precipitation was greater or equal to 30 % are included. This helps maximize the contrast when examining terms of the surface water balance. Three cases during boreal spring/austral fall, and three cases during boreal summer/austral winter have been chosen.

Conspicuously absent is the region of central and eastern North America that experienced a major drought during 1988. That drought was predominantly a springtime event. However, precipitation during spring 1987 was also somewhat below average over much of the same region, so the 1988-minus-1987 differences do not highlight that region.

Table 1 shows the rates (mm per season) during 1987 and 1988 for the various flux terms, as well as the change in storage of soil moisture. Also shown are the 1988-minus-1987 differences. Five of the six regions had significant increases in observed precipitation in 1988 compared to 1987, in the range of 100–200 %. Central South America was the one region to experience a large decrease.

There appears to be little in common among all regions in terms of the reaction of the components of

Fig. 2. Comparison of two-year average net radiation and evaporation, scaled by precipitation, for all ice-free land points. Thick solid curve shows the average relationship from the model data. Solid lines show the theoretical limitations on evaporation from radiant energy (diagonal) and precipitation (horizontal).

the water balance to large precipitation variations. With the exception of South Africa, most regions show little impact in transpiration. Many areas are sparsely vegetated and have low values of transpiration both years. Alaska and central South America are generally well watered during the seasons of interest, and conditions in Alaska are still quite cool during the spring, suppressing transpiration. There is a strong correlation between subsurface runoff and soil moisture change. This is to be expected because gravitational drainage depends heavily on deep-layer soil moisture. In five of the six regions, the largest absolute perturbations are found in the soil moisture storage. Perturbations in subsurface runoff are among the top two in magnitude over five of the regions, and rank third over Australia. Bare soil evaporation also buffers much of the rainfall perturbation over arid and sparsely vegetated regions. Interception generally plays a minor role, and over Alaska the perturbation in snow evaporation is quite small compared to the precipitation anomaly.

4. Sensitivity studies

A number of sensitivity studies have been conducted, where specific parameterizations in 2D-SiB have been altered to affect the way precipitation is

partitioned between evapotranspiration and runoff. These changes alter both aspects of the project design and the model itself. The results are presented here.

4.1 Convective precipitation

Convective precipitation is assumed to fall in a spatially and temporally heterogeneous pattern over the grid areas and time intervals used in GSWP. This is consistent with the treatment in most dynamical weather and climate models of the atmosphere, as well as larger-scale hydrologic models. This is in contrast to large-scale or stratiform precipitation, which is assumed to fall relatively evenly and steadily over large regions. Most atmospheric models distinguish between these two forms of precipitation. Because different physical processes are involved in their generation, different parameterizations are used to calculate their occurrence and intensity. Thus, they can deliver separate quantities to the land surface model for convective and large-scale precipitation. This is important, as these precipitation types have different infiltration properties when considered over grid-scale areas.

Convective precipitation is characterized by intense showers in one or many small cells (10^0 –

Fig. 3. Regions and seasons of consideration for seasonal moisture budgets.

10^2 km^2) surrounded by regions of lesser rainfall rates, or even rain-free regions. Generally, these storms are short lived or fast moving, meaning that at a specific location the intense rainfall may last for a relatively brief period (a few minutes to an hour). The intense rainfall rates can exceed the maximum infiltration rate for soil during that period, or quickly saturate wet or shallow soils, producing runoff. The area of these storms is generally much smaller than the area of a model grid box, and is certainly smaller than the 1° grid used in GSWP. Thus, different regions of the same grid box may experience infiltration excess and no rainfall whatsoever. Large-scale precipitation may drop an identical mass of water onto a grid box, but because of its gentler rate and uniform distribution, wetting and infiltration will be much more complete.

Many land surface models, including SSiB, parameterize the subgrid scale variability of convective precipitation by assuming a distribution of rainfall intensities within the grid box. In most of the SiB family of models:

$$I_c(x) = a_c e^{-b_c x} + c_c, \quad (1)$$

where $I_c(x)$ is the relative precipitation intensity as a function of normalized grid area x , ($0 \leq x < 1$). The coefficients a , b , and c are chosen to provide the desired distribution of intensities, such that:

$$\int_0^1 I_c(x) dx = 1 \quad (2)$$

See Sellers *et al.* (1996b) for a thorough description.

This approach provides a realistic distribution of convective precipitation within a grid area for time

steps up to an hour or so. However, in GSWP the meteorological data based on global analyses are available only every six hours. Most fields are interpolated linearly to produce hourly data, but such interpolation greatly reduced convective rainfall intensities. We have investigated the effect of this averaging by conducting a parallel integration where a distribution in time of the convective precipitation intensities is also assumed. We suppose that:

$$G_c(t) = 4\sqrt{\frac{2}{\pi}} e^{-8t^2}, \quad (3)$$

where G_c is the intensity as a function of time on the 6-hour interval normalized so that $0 \leq t < 1$, and

$$\int_0^1 G_c(t) dt = 1. \quad (4)$$

The resulting intensities binned over one-hour intervals is then shuffled to produce the moderately skewed distribution shown in Fig. 4. Nearly half of the total precipitation is made to fall during a single hour with this choice of distribution.

In addition to this treatment, another run is made where all precipitation is assumed to be large-scale, so that none of the convective distribution treatments come into play. The effects of these approaches are shown in Table 2 for three of the regions during 1987 and 1988. Increasing heterogeneous parameterizations of convective precipitation generally lead to greater runoff, more transpiration, and less evaporation from canopy interception. Changes in soil storage are lessened in magnitude, regardless of

Table 1. Components of the surface water balance over the regions shown in Fig. 3 during 1987, 1988 and 1988 minus 1987. Storage change is calculated as a residual.

S Africa (MAM)	1987	1988	Change
Precipitation	0.81	2.10	1.29
Sfc. Runoff	0.00	0.00	0.00
Subsfc. Runoff	0.33	0.79	0.46
Transpiration	0.24	0.48	0.23
Bare Soil	0.86	1.24	0.38
Snow Evap.	0.00	0.00	0.00
Interception	0.09	0.19	0.10
Storage Change	-0.71	-0.60	0.11
Australia (MAM)	1987	1988	Change
Precipitation	0.54	1.62	1.08
Sfc. Runoff	0.00	0.01	0.01
Subsfc. Runoff	0.20	0.37	0.16
Transpiration	0.04	0.04	0.00
Bare Soil	0.69	0.88	0.19
Snow Evap.	0.00	0.00	0.00
Interception	0.04	0.07	0.04
Storage Change	-0.43	0.25	0.68
Alaska (MAM)	1987	1988	Change
Precipitation	0.49	1.13	0.64
Sfc. Runoff	0.00	0.00	-0.00
Subsfc. Runoff	0.28	0.34	0.05
Transpiration	0.01	0.01	-0.00
Bare Soil	0.06	0.02	-0.04
Snow Evap.	0.03	-0.02	-0.04
Interception	0.03	0.08	0.05
Storage Change	0.08	0.70	0.62
Sahel (JJA)	1987	1988	Change
Precipitation	1.62	3.14	1.51
Sfc. Runoff	0.01	0.06	0.05
Subsfc. Runoff	0.19	0.61	0.42
Transpiration	0.12	0.14	0.02
Bare Soil	0.90	1.25	0.36
Snow Evap.	0.00	0.00	0.00
Interception	0.05	0.08	0.03
Storage Change	0.35	0.99	0.64
S Asia (JJA)	1987	1988	Change
Precipitation	1.53	3.47	1.94
Sfc. Runoff	0.00	0.00	0.00
Subsfc. Runoff	0.37	1.03	0.66
Transpiration	0.04	0.06	0.02
Bare Soil	0.97	1.50	0.53
Snow Evap.	0.02	0.03	0.00
Interception	0.04	0.09	0.05
Storage Change	0.08	0.76	0.67
S America (JJA)	1987	1988	Change
Precipitation	1.52	0.51	-1.01
Sfc. Runoff	0.00	0.00	-0.00
Subsfc. Runoff	0.93	0.52	-0.41
Transpiration	0.75	0.74	-0.01
Bare Soil	0.72	0.70	-0.03
Snow Evap.	0.01	0.00	-0.01
Interception	0.22	0.11	-0.12
Storage Change	-1.12	-1.56	-0.44

whether they are positive or negative during the season. It appears that the control formulation optimizes bare soil evaporation — when no convective precipitation is simulated, throughfall is reduced, decreasing the water supply to evaporation at the soil surface. When temporal variation is added to spatial rainfall variability, soil evaporation again decreases because, although plenty of water reaches the ground, more water runs off.

4.2 Soil porosity

Instead of a uniform porosity for all three soil layers, a vertical profile meant to simulate shallow aeration and deep compaction is used in another sensitivity experiment. This approach was expected to allow greater infiltration and slower gravitational drainage due to the alteration of the porosity profile. The new profile is a quadratic fit to an exponential decay of porosity from a value higher at the surface than the mean value specified for the GSWP experiment, to a constant value slightly less than the mean value below $z_c = 1$ m:

$$\theta_s = s_2 z^2 + s_1 z + s_0, \quad z < z_c, \quad (5)$$

$$\theta_s = s_2 z_c^2 + s_1 z_c + s_0, \quad z \geq z_c,$$

where we choose $s_2 = 0.3899$, $s_1 = -0.6847$, and $s_0 = 0.3191$. The resulting profile resembles that shown in Fig. 5. The range of porosities is exaggerated compared to observed profiles to amplify any signal in the sensitivity experiment.

The results for three of the areas are shown in Table 3; the general pattern holds for all areas examined. Surface evaporation increases at the expense of runoff. However, the effects are small, considering that significant changes have been made to the porosities. For more realistic porosity profiles, the effects would be even less.

4.3 Surface layer thickness

The final sensitivity study described here involves the impact of specified surface soil layer thickness on the hydrologic balance. In the GSWP experiment, an arbitrary value of 50 mm was chosen as the surface soil layer thickness for all participating multi-layer models. However, many land surface parameterizations use smaller values for this surface layer when coupled to atmospheric models. The thickness, which represents the depth of soil having significant direct exchanges of heat and moisture with the air, is often specified as low as 20 mm. We chose this lower value for this experiment.

Table 4 shows the effect of this change on the same three areas in Tables 2 and 3. With the reduction in surface thickness, there is a general reduction in direct evaporation from the soil. This is seen to be true globally where soil evaporation comprises a significant fraction of the surface water balance, mostly

Fig. 4. Redistribution of 6-hourly convective precipitation in one hour bins to simulate temporal variability.

over sparsely vegetated areas of the low and midlatitudes. The decrease is due to the reduction in the fraction of soil water exposed to the atmosphere. In most regions, the decrease is offset by an increase in runoff, although transpiration does increase in areas of shrub or sparse forest (*e.g.*, southern Africa). In many regions, surface runoff increases to account for 2–10 % of the total precipitation. The thinner surface layer has a much smaller capacity, causing infiltration excess to be more likely during heavy rainfall or snowmelt events.

Each GSWP integration begins from an equilibrium soil moisture state, attained after a multi-year spin-up process where 1987 forcings are used to drive the model repeatedly until certain soil moisture convergence criteria are met. We find that changing the surface layer thickness has a profound effect on this equilibrium state, particularly in the deepest layer. Figure 6 shows for each of the three soil layers the difference in the equilibrium soil wetness using 20 mm versus 50 mm surface layer depth. Over many areas, there is an increase in soil wetness. In the deep layer, this increase averages about 5 % of the total soil capacity over sparsely vegetated regions. The pattern in the root zone is very similar to the deep layer, although relative fractions are less. Variations in the surface layer are small and somewhat incoherent. Increased soil moisture might be expected; because less precipitation is lost to evap-

oration, more water can infiltrate to the deep layer where it is relatively isolated from direct evaporation or transpiration. An increase in baseflow accompanies the higher soil moisture.

However, several areas over North Africa, North America and central Asia show decreased soil moisture in both subsurface levels, especially in the root zone, when the surface layer is made thinner. This appears to be due to changes in the effect of surface heating and solar radiation on the thinner upper layer. These areas tend to have more surface evaporation, rather than less. This appears true especially during the spring. There is a coincident increase in surface temperature during these episodes. It appears that in these arid regions, where solar radiation is plentiful and vegetation is sparse, a shallower surface layer dries quickly through surface evaporation. This reduction in soil moisture decreases the effective heat capacity in the force-restore calculation of surface temperature, allowing that temperature to rise. Because of the strong soil moisture gradient between surface and root zone, diffusion of water can overcome gravity and replenish the surface layer, where it evaporates quickly in the warmer soil. This positive feedback pumps water from the root layer to the surface and out via direct evaporation much more efficiently than in the control case, where the thicker surface layer buffers against rapid drying and warming. It appears that this mechanism

Table 2. Components of the surface water balance during 1987 and 1988 for cases of “no convection,” control (with sub-grid spatial variation), and both spatial and temporal variation.

S Africa (MAM)	1987			1988		
	No Convect	Control	Space & Time	No Convect	Control	Space & Time
Precipitation	0.81	0.81	0.81	2.10	2.10	2.10
Runoff	0.31	0.33	0.34	0.74	0.79	0.80
Transpiration	0.20	0.24	0.25	0.40	0.48	0.49
Soil Evap.	0.80	0.86	0.82	1.16	1.24	1.19
Interception	0.20	0.09	0.06	0.42	0.19	0.14
Storage Change	-0.71	-0.71	-0.66	-0.61	-0.60	-0.52
Sahel (JJA)	No Convect	Control	Space & Time	No Convect	Control	Space & Time
Precipitation	1.62	1.62	1.62	3.14	3.14	3.14
Runoff	0.19	0.20	0.36	0.68	0.67	0.97
Transpiration	0.08	0.12	0.13	0.10	0.14	0.15
Soil Evap.	0.85	0.90	0.87	1.21	1.25	1.17
Interception	0.16	0.05	0.04	0.24	0.08	0.05
Storage Change	0.34	0.35	0.22	0.91	0.99	0.79
S Asia (JJA)	No Convect	Control	Space & Time	No Convect	Control	Space & Time
Precipitation	1.53	1.53	1.53	3.47	3.47	3.47
Runoff	0.31	0.37	0.52	0.92	1.03	1.33
Transpiration	0.03	0.04	0.04	0.05	0.06	0.07
Soil Evap.	0.93	0.97	0.93	1.47	1.50	1.41
Interception	0.10	0.04	0.03	0.20	0.09	0.07
Storage Change	0.13	0.08	-0.02	0.80	0.76	0.56

Fig. 5. Vertically varying porosity profile, and its representation in three discrete soil layers.

may be highly dependent of meteorological and soil conditions, and occurs in an irregular fashion. It is quite likely that this is a model artifact — we plan to investigate whether this situation exists when a 3-layer heat diffusion scheme replaces the force-restore method in SSiB.

5. Conclusions

Forcing a two-dimensional implementation of the land surface scheme SSiB with ISLSCP Initiative I data as part of the Global Soil Wetness Project

(GSWP) has resulted in a global two-year climatology of soil wetness and surface fluxes. Examination of the water balance shows that surface runoff accounts for a small percentage of total runoff, and surface evaporation from the soil accounts for a majority of the water fluxes from the soil over all but heavily forested areas where transpiration dominates. Artifacts of the model baseflow parameterization are evident in the high values of subsurface runoff. Examination of interannual variability of seasonal water balances over specific regions shows

Table 3. Components of the surface water balance during 1987 and 1988 for the control case (porosity constant in the vertical), and porosity decreasing with depth.

	1987			1988		
	Control	Var. Poros	Diff.	Control	Var. Poros	Diff.
S Africa (MAM)						
Precipitation	0.81	0.81	0.00	2.10	2.10	0.00
Runoff	0.33	0.32	-0.01	0.79	0.75	-0.04
Transpiration	0.24	0.25	0.01	0.48	0.49	0.01
Soil Evap.	0.86	0.90	0.04	1.24	1.27	0.04
Interception	0.09	0.09	0.00	0.19	0.19	0.00
Storage Change	-0.71	-0.75	-0.04	-0.60	-0.61	-0.01
Sahel (JJA)						
Precipitation	1.62	1.62	0.00	3.14	3.14	0.00
Runoff	0.20	0.17	-0.03	0.67	0.59	-0.09
Transpiration	0.12	0.13	0.00	0.14	0.15	0.00
Soil Evap.	0.90	0.97	0.07	1.25	1.32	0.07
Interception	0.05	0.05	0.00	0.08	0.08	0.00
Storage Change	0.35	0.30	-0.05	0.99	1.01	0.02
S Asia (JJA)						
Precipitation	1.53	1.53	0.00	3.47	3.47	0.00
Runoff	0.37	0.36	-0.01	1.03	1.00	-0.03
Transpiration	0.04	0.04	0.00	0.06	0.06	0.00
Soil Evap.	0.97	1.00	0.03	1.50	1.53	0.03
Interception	0.04	0.04	0.00	0.09	0.09	0.00
Storage Change	0.08	0.06	-0.02	0.76	0.76	0.00

Table 4. Components of the surface water balance during 1987 and 1988 for the control case (50 mm surface layer), and 20 mm surface layer.

	1987			1988		
	Control	20mm Sfc.	Diff.	Control	20mm Sfc.	Diff.
S Africa (MAM)						
Precipitation	0.81	0.81	0.00	2.10	2.10	0.00
Runoff	0.33	0.34	0.01	0.79	0.83	0.04
Transpiration	0.24	0.29	0.04	0.48	0.53	0.05
Soil Evap.	0.86	0.85	-0.01	1.24	1.21	-0.03
Interception	0.09	0.09	-0.00	0.19	0.19	-0.00
Storage Change	-0.71	-0.75	-0.04	-0.60	-0.66	-0.06
Sahel (JJA)						
Precipitation	1.62	1.62	0.00	3.14	3.14	0.00
Runoff	0.20	0.30	0.10	0.67	0.92	0.25
Transpiration	0.12	0.13	0.00	0.14	0.14	-0.00
Soil Evap.	0.90	0.79	-0.10	1.25	1.14	-0.11
Interception	0.05	0.05	-0.00	0.08	0.08	-0.00
Storage Change	0.35	0.35	0.00	0.99	0.86	-0.13
S Asia (JJA)						
Precipitation	1.53	1.53	0.00	3.47	3.47	0.00
Runoff	0.37	0.38	0.01	1.03	1.10	0.06
Transpiration	0.04	0.04	0.00	0.06	0.06	-0.00
Soil Evap.	0.97	0.93	-0.04	1.50	1.46	-0.03
Interception	0.04	0.04	-0.00	0.09	0.09	-0.00
Storage Change	0.08	0.11	0.02	0.76	0.73	-0.03

that changes in soil moisture storage often absorb much of the variability in precipitation, with direct implications for subsurface runoff. Changes in bare soil evaporation are also prevalent in sparsely vegetated areas.

A number of sensitivity studies have been per-

formed, based on the control integration of SSiB. The sensitivity studies are designed to investigate how the partitioning of precipitation between runoff and evapotranspiration is affected when aspects of the soil and the treatment of convective precipitation are altered. Specifically, three approaches to

Fig. 6. Difference in equilibrium initial condition wetness (fraction of saturation) for each soil layer between integrations with 20 mm and 50 mm surface soil layers.

treating convective precipitation are examined: no treatment (uniform precipitation in space and time), subgrid variability in space, and variability in both space and in time (within a six-hour interval). Other sensitivity studies involved varying the vertical profile of soil porosity, and choosing a smaller surface soil layer thickness.

In general, transpiration is fairly insensitive to the changes made. Transpiration is controlled by the stomatal resistance. In SSiB, stomatal resistance is parameterized as a function of certain environmental conditions, namely temperature, vapor pressure, and the availability of moisture in the soil. Some other models have a more ecological approach, where sunlight and nutrient availability, in addition to moisture stress, determine transpiration rates as part of the controls on photosynthesis. In a more elaborate model, it is possible that a different sensitivity of transpiration might be found.

The most sensitive terms on seasonal time scales appear to be runoff, direct evaporation from the soil, and the seasonal change of water storage in the soil matrix. Changes in storage are a passive consequence of changes in the fluxes of water out of the soil. On long time scales, storage changes would likely become insignificant. Then, the balance is simplified between variations in runoff and direct evaporation from the soil surface. The soil characteristics, particularly in the surface layer, are strong determinants of both surface evaporation and maximum infiltration rates. The treatment of the partitioning of convective precipitation on subgrid and sub-time-step scales will also greatly affect infiltration and runoff. It would appear from this study that calibration and development efforts should focus on these areas.

Most soil schemes use a small number of layers to represent a soil column of several meters depth. Of the 27 models participating in PILPS as of 1995, only four had more than three layers for soil moisture calculations (Henderson-Sellers *et al.*, 1995). Model developers are recognizing the shortcomings of using a small number of layers: gradients are calculated over large intervals contributing to errors in calculating diffusivity and conductivity, and thick layers make an unrealistically large volume of water available for movement between distant levels of the soil. If one could estimate the proper statistical distribution of subgrid scale precipitation from high-resolution radar over catchments where reliable streamflow data were available, one could find an optimal distribution of layers for a soil model over that region using available data on soil texture, slope, depth, composition, *etc.* As validation and calibration data are becoming available at higher resolution, and interest in land surface models and their

performance spreads among the hydrological community, the time is at hand for improvements in simulating the behavior of water in the soil in weather and climate models. No longer are land surface models merely expected to deliver reasonable estimates of heat fluxes as a lower boundary condition for atmospheric models. Now they must possess a closed and realistic hydrologic cycle.

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SSiB の陸面過程における降水量の浸透について

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GSWP の一環として、陸面過程に SSiB を組み込んだ 2 次元の全球モデルをオフラインで 2 年間積分し、土壌水分量と地表面フラックスの気候値を求めた。そして、この気候値を多くの感度実験の結果と比較した。感度実験では、土壌のパラメタリゼーションと対流性降水の扱いを変えた時に、降水量が流出量と蒸発散量にどのように分配されるかを調べた。

標準実験では、地表面水収支各項のもっともらしい空間分布が得られ、現実的な季節変化と年々変動が見られた。土壌表層からの蒸発は、おおむね地表面からの水フラックスの大部分を占めているが、密林地域では蒸散が卓越している。一般的に、季節程度の時間スケールにもっとも敏感らしいのは、流出量、土壌からの直接蒸発量、土壌マトリックスにおける保水量の季節変化であることが、感度実験から明らかになった。グリッドスケールでの適切な流出量を再現すること、および植生による遮断蒸発量を過大評価しないようにするためには、対流性降水の現実的な時空間分布が必要である。また、陸面過程において適当に与えられることが多いパラメーターの一つである表層土壌の厚さが、実験結果に対して敏感であることが分かった。一方、表層土壌に空隙を作り深層土壌を固める効果がある、土壌の空隙率の鉛直プロファイルは、実験結果に対して敏感ではないことも分かった。