Representation of Water Table Dynamics in a Land Surface Scheme. Part I: Model Development

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ABSTRACT

Most of the current land surface parameterization schemes lack any representation of regional groundwater aquifers. Such a simplified representation of subsurface hydrological processes would result in significant errors in the predicted land surface states and fluxes especially for the shallow water table areas in humid regions. This study attempts to address this deficiency. To incorporate the water table dynamics into a land surface scheme, a lumped unconfined aquifer model is developed to represent the regional unconfined aquifer as a nonlinear reservoir, in which the aquifer simultaneously receives the recharge from the overlying soils and discharges runoff into streams. The aquifer model is linked to the soil model in the land surface scheme [Land Surface Transfer Scheme (LSX)] through the soil drainage flux. The total thickness of the unsaturated zone varies in response to the water table fluctuations, thereby interactively coupling the aquifer model with the soil model. The coupled model (called LSXGW) has been tested in Illinois for an 11-yr period from 1984 to 1994. The results show reasonable agreements with the observations. However, there are still secondary biases in the LSXGW simulation partially resulting from not accounting for the spatial variability of water table depth. The issue of subgrid variability of water table depth will be addressed in a companion paper.

1. Introduction

Atmospheric general circulation models (GCMs) are widely used for predicting the impacts of natural and anthropogenic perturbations on the earth's climate. With the increased recognition of the importance of feedback between land surface processes and climate (Charney et al. 1977; Shukla and Mintz 1982; Delworth and Manabe 1988), and the interest in evaluation of hydrological and agricultural impacts under the changed climate conditions (Manabe et al. 1981; Rind et al. 1990; Wetherald and Manabe 1995), the development of realistic parameterizations of land surface processes compatible with the scale of a climate model grid cell (~50–500 km) has been an area of active research over the last two decades (Sellers et al. 1986; Abramopoulos et al. 1988; Verseghy 1991; Wood et al. 1992; Dickinson et al. 1993). Moreover, the impact of land surface process representation on GCM simulations of climate sensitivity to the increasing greenhouse gases has become the focus of much public concern (Houghton et al. 1995).

Land surface parameterization schemes (LSPs) are now an important component of numerical weather prediction models and general circulation models. They calculate the water and heat fluxes from land surface to atmosphere and update the surface and subsurface variables affecting these fluxes. The earliest LSP used in climate models was the simple "bucket" model (Manabe 1969). In this model, evaporation was calculated as the product of the potential evaporation and a factor depending on soil moisture, and runoff was assumed to occur whenever the surface water storage exceeded the specified bucket size. Schemes of this type generally assumed a geographically constant value of the bucket size and neglected the impact of vegetation control on the land surface fluxes of water, heat, and momentum. Later in the 1980s–1990s, a new generation of LSPs was proposed (Sellers et al. 1986; Dickinson et al. 1993) to improve the limitations of the bucket

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model. Most of these LSPs were based on the knowledge gained from laboratory experiments or fieldwork at the small scale. These schemes were often referred to as big-leaf models since they assumed horizontal homogeneity of the land surface characteristics within a grid cell. LSPs usually include considerable vertical resolution and physical realism in the transfer of energy, mass, and momentum. However, the soil hydrology treatment is rather simple and spatial variability is ignored altogether in most LSPs. The common practice is to use a single soil column with 2–10 sublayers to simulate the surface heat and water fluxes. The use of such simplified "big leaf–single soil column" models is justified in terms of computational efficiency, a lack of information about the detailed distribution of vegetation types, and the lack of appropriate approaches to accommodate spatial heterogeneity at the land surface.

The Biosphere Atmosphere Transfer Scheme (BATS; Dickinson et al. 1993) and the Simple Biosphere Schemes (SiB; Sellers et al. 1986) are the most commonly used big leaf–single soil column models in global and regional climate studies. Typically, these models solve the energy and water balance equations for only the single soil column and for the big leaf. To characterize the various soil and vegetation properties as well as the hydrological and biogeochemical processes at the earth's surface, these models require a large number of empirical constants that in practice are difficult to estimate. Moreover, there are still unresolved issues as to whether the use of point or small-scale parameters is valid at the atmospheric model grid scale. Because these point parameters might vary over an atmospheric model grid cell, it is not clear how these small-scale, local parameters could be aggregated to provide the grid-scale "representative" values that could account for the nonlinearity of the underlying land surface processes (Wood et al. 1992).

Offline tests such as the Project for Intercomparison of Land Surface Parameterization Schemes [PILPS; see http://www.cic.mq.edu.au/pilps-rice/ and the overview by Henderson-Sellers et al. (1995)] and the Global Soil Wetness Project [GSWP; see http://grads.iges.org/gswp/ and the overview by Dirmeyer et al. (1999)] have demonstrated that small differences in the model physics among various schemes can lead to a wide spread in the simulation results. Although many model parameterizations responsible for the simulation biases were diagnosed and corrected by the individual modeling group participating in the PILPS and GSWP, it is still unclear how to resolve the differences among schemes, and most importantly, how this spread would be affected by coupling these LSPs to their host climate models.

The major objective of this study is to improve the ability of an LSP to model the land surface hydrology at the GCM grid scale. This objective has important implications regarding the evaluation of the impact on regional water resources due to climate change. It has been recognized (Rind et al. 1990; Wetherald and Manabe 1995) that the reliable simulation of any climate change scenario by climate models has been hampered by the inaccuracy in the parameterization of land surface processes. Only by formulating the individual physical components of the LSPs to be as realistic as possible can we expect to make progress in the prediction of the impact caused by climate change. This reasoning hence calls for the most realistic parameterization to be incorporated into the modeling process, which is the main objective of this study.

This paper is organized into six sections. In the next section we summarize the importance of water table representation in climate models. Section 3 presents the results of two 11-yr (1984–94) simulations in Illinois using a traditional land surface scheme without any representation of water table dynamics. The objective is to demonstrate the deficiencies in the model simulations resulting from neglecting the role of the unconfined aquifers in the shallow water table areas. In section 4, a simple unconfined aquifer model is developed and interactively coupled to the land surface scheme. The fully coupled groundwater–land surface model has been tested by using the same 11-yr (1984–94) Illinois data as that used in section 3. The simulation results are presented and compared to the corresponding observations in section 5. This is followed by a summary of conclusions in section 6.

2. The importance of water table representation

The current LSPs designed for use in climate models in general do not include the representation of a shallow water table. According to Zektser and Loaiciga (1993), the global averages of base flow/precipitation ratio and base flow/streamflow ratio are about 10% and 30%, respectively. Why should water table dynamics be included in LSPs used in climate models? Most importantly, since all the LSPs are developed based on the concept of water balance, it is unrealistic to expect that these schemes would reproduce correct water balance without considering the entire set of significant processes involved in the hydrological cycle. At least for the humid climates, the water table usually lies near the ground surface and for these areas, groundwater runoff is often the dominant streamflow generation mechanism. Under such conditions, the regional climate directly interacts with groundwater through the water fluxes near the water table, namely the groundwater recharge and capillary rise (Levine and Sulvucci 1999).

Unlike the relatively steady and uniform soil moisture distribution under the deep water table condition, the presence of a shallow water table would significantly alter the vertical soil moisture profile at least in the lower part of the vadose zone. Given that most of the land surface hydrological processes (e.g., infiltration, evapotranspiration, drainage, runoff, etc.) are highly dependent on soil moisture, the role of the shallow water table must be incorporated in any LSPs in order to realistically simulate the soil moisture dynamics. Compared to the deep water table condition, the presence of a shallow water table results in a decrease of infiltration and an increase of evapotranspiration due to the downward increase of moisture content in the lower vadose zone. Moreover, field evidence points to the close correlation between plant species and water table depth (Nichols 1993, 1994). The proximity of the capillary fringe to the land surface allows plant roots to enter the saturated zone, while the root penetration is constrained by the anaerobic condition of the saturated zone (Miller and Eagleson 1982).

A recent analysis of the hydroclimatology in Illinois (Yeh et al. 1998) has shown that both groundwater storage change and groundwater runoff are significant terms in the monthly water balance for shallow water table areas. In Illinois, the seasonal cycle of the large-scale average water table depth ranges between 2 and 4 m below the surface. The monthly changes in saturated (groundwater) storage and unsaturated (soil moisture) storage are equally significant in shaping the seasonal hydrologic cycles. The monthly water balance in Illinois could not be closed without the consideration of the changes in groundwater storage (Yeh et al. 1998). The annual change of groundwater storage, unlike the soil moisture storage, does not always integrate to zero, especially during the drought and flood years. Furthermore, the water balance analysis conducted by Yeh (2002) indicated that the high evaporation rate during the summer ($\sim$120 mm month^{-1}) in Illinois as reported by Yeh et al. (1998) is contributed to a significant degree by shallow groundwater. Without considering groundwater storage in the water balance computation, summer evaporation would be underestimated by about 25%. Therefore, the incorporation of groundwater component is indispensable for an LSP to be applied in shallow water table areas such as Illinois.

Figure 1 shows the scatterplots of the observed average monthly streamflow versus the observed average precipitation and water table depth in Illinois from 1984 to 1994. These data are the spatial averages from the hydrologic observational networks in Illinois (Yeh et al. 1998). As shown in this figure, although very little correlation can be noted between average precipitation and streamflow, there exists a significant, nonlinear relationship between average water table depth and the corresponding streamflow at the monthly time scale in Illinois. In addition, the study of hydrometeorological anomalies in Illinois by Eltahir and Yeh (1999, 1205–1206) indicated that the correlation scale of the observed water table level anomalies is about 12 months, much longer than the 3-month correlation scale for both the soil moisture and the streamflow anomalies

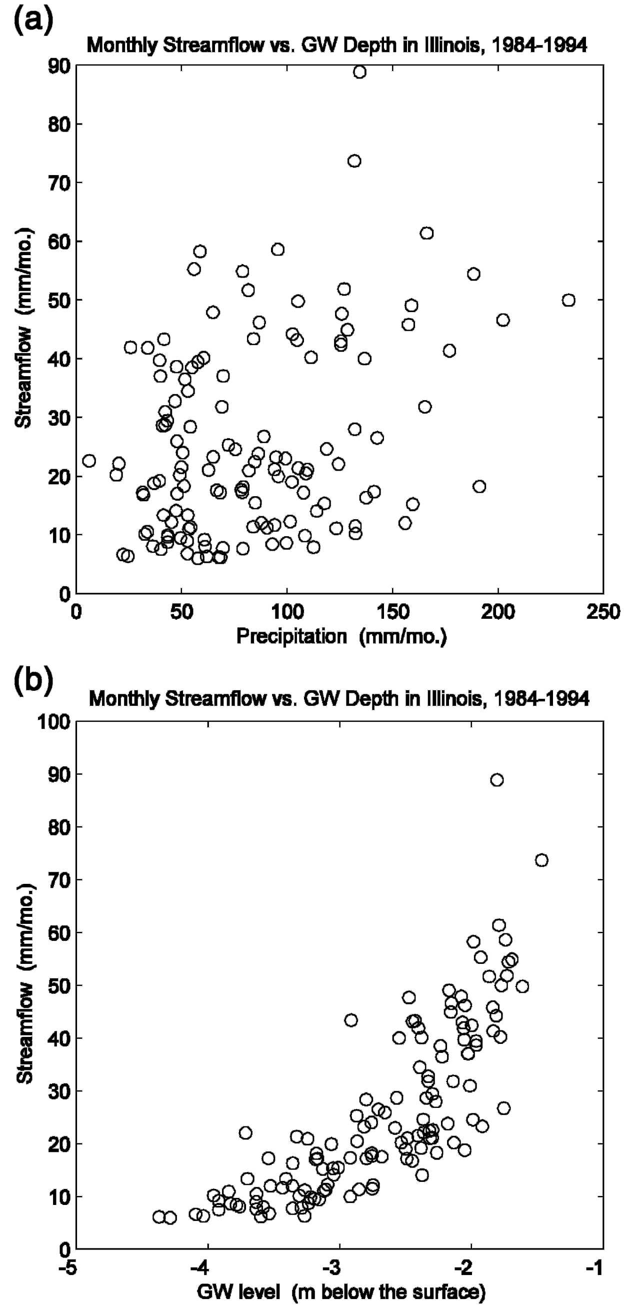

FIG. 1. Plots of monthly (a) precipitation and (b) water table depth vs monthly streamflow during 1984–94 in Illinois.

and 1-month correlation scale for the precipitation and atmospheric moisture convergence anomalies. The soil water reservoir and the groundwater reservoir act as two adjacent low-pass filters of the atmospheric forcing as it propagates downward through the land surface hydrologic cycle. Because of the longer memory compared to soil moisture, groundwater would create a significant persistence of anomalies and feedback upward to impact the future climate.

Despite its importance, the water table depth, which is a basic hydrologic variable, has not been a standard component in the LSPs. One possible explanation for this absence is that the current generation of LSPs views the soil column as the fundamental hydrologic unit. The large grid scale and thin soil layers (usually 1–5 m) typically considered in LSPs render the groundwater dynamics a seemingly insignificant hydrological process. Furthermore, currently there are no large-scale or global databases that provide information about water table depth, which makes it difficult to implement any large-scale subsurface water table scheme.

From the atmospheric point of view, it might be argued that because water table dynamics only impact evaporation on a small fraction of a GCM grid element, its significance can be neglected. However, this is also possible as a result of simplicity consideration rather than reasonable scale arguments (Chen et al. 1996). At least for the shallow water table regions, such an inaccurate (simplified) representation of subsurface hydrological processes might result in significant errors in the predicted runoff and hence affect the accuracy of the predicted evaporation in climate models.

Because of the absence of water table representation in the current LSPs, the specification of the lower boundary condition of soil moisture (which predominately dictates the subsurface runoff generation in the model) is rather difficult. The common practice in the LSPs is to apply a gravity drainage (i.e., free drainage) condition; that is, the drainage flux equals E the unsaturated hydraulic conductivity at the bottom of the lowest soil layer, or a linear function of the unsaturated conductivity with an empirical coefficient accounting for other factors affecting drainage such as the topographic slope and amplitude, or the location of bed rocks (Boone and Wetzel 1996, p. 166). In fact, 13 out of the total 16 LSPs participating in phase 2c of the PILPS project in the Red–Arkansas River basin applied the gravity drainage condition for subsurface runoff generation (see the summary by Lohmann et al. 1998, their Table 1). The application of the gravity drainage condition assumes that the vertical moisture profile is uniform near the soil bottom so that the upward diffusion flux is negligible. However, this assumption is valid only when water table is deep compared to the thickness of the soil column (typically 2–5 m in LSPs) such that the unsaturated–saturated zone interactions are insignificant. If the water table is shallow, the sharp moisture gradient right above it would significantly enhance the upward diffusion flux. The outcome is a wetter soil moisture condition resulting in a decreased infiltration and an increased evapotranspiration. Moreover, for those schemes specifying the lower boundary condition of soil moisture as a linear function of the gravity drainage condition, the multiplying coefficient (or function; see the summary in Table 3 of Boone and Wetzel 1996) has rather ambiguous physical reality. For example, in the SiB model (Sellers et al. 1986) this coefficient is defined as the sine of the slope angle. Apparently, it makes little sense to specify a single slope angle to represent the drainage condition for a model grid with the size as large as Illinois. Moreover, in the BATS model (Dickinson et al. 1993), the drainage at the bottom of the soil column was empirically assumed as a constant (4×10^{-4} mm s^{-1}) independent of soil type multiplied by a nonlinear function of soil saturation because "it is difficult to relate the drainage at the bottom of the subsoil layer (at 5–10 m) to soil properties at the surface . . ." (Dickinson et al. 1993, p. 37).

More importantly, the specification of gravity drainage–like conditions for the bottom soil moisture cannot allow for the occurrence of negative soil drainage (i.e., upward water fluxes from the aquifer to the soils). The evidence for net upward groundwater flux has been estimated in Illinois by Yeh (2002) and also has been reported elsewhere by Tschinkel (1963), Daniel (1976), and Zecharias and Brutsaert (1988). Figure 2 shows the 11-yr (1984–94) monthly time series and the average seasonal cycle of the state-average groundwater recharge in Illinois estimated by Yeh (2002) based on the large-scale water balance analysis. As seen, negative groundwater recharge occurs in 38 months (most of them in summer) during the period of 1984–94 (probability of ~30%). The seasonal cycle indicates that negative recharge occurs during summer months with a maximum of 14 mm month^{-1} in August. During the summer when the deficit in the root-zone soil moisture is large, upward water flux from the shallow aquifer replenishes the root-zone soil moisture, resulting in a steep decline of the water table (Yeh et al. 1998).

As a first attempt to address this issue, Yeh (2002) developed a lumped aquifer model to represent the regional unconfined aquifer as a nonlinear reservoir and incorporated it into an LSP. The model testing results are summarized in this paper. According to our best

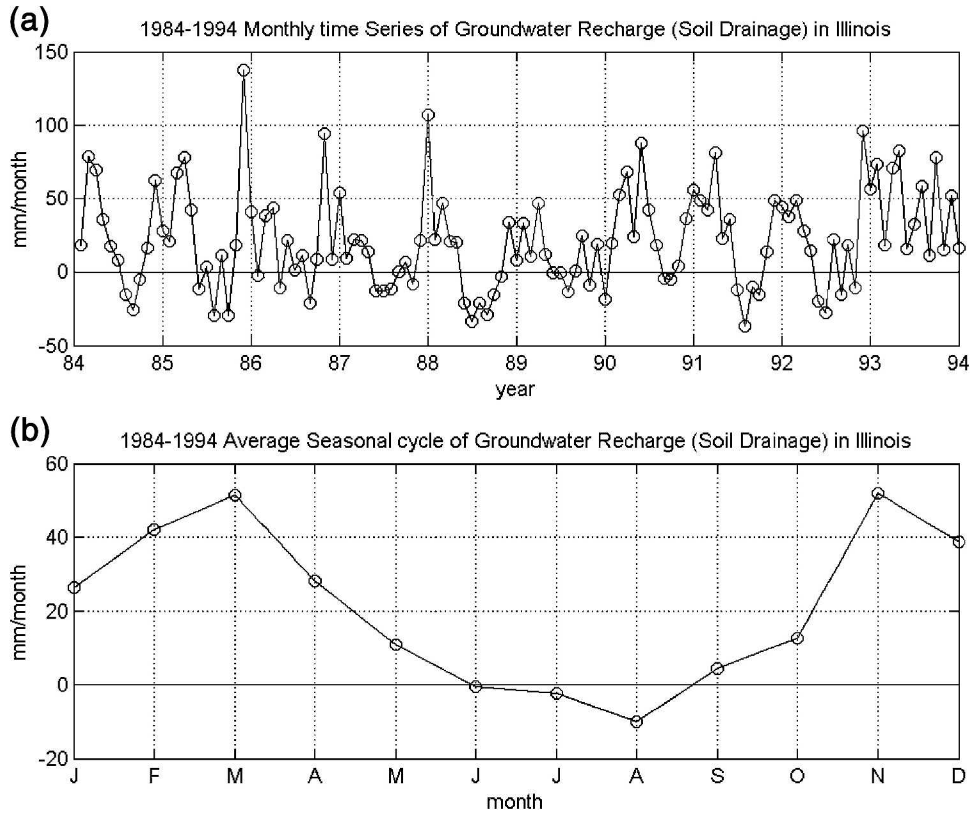

FIG. 2. The large-scale average 11-yr (1984–94) (a) monthly time series and (b) seasonal cycle of the groundwater recharge (soil drainage) flux estimated by Yeh (2002) based on water balance analysis.

knowledge, two studies addressing the importance of surface groundwater interactions in LSPs were found in the literature. York et al. (2002) studies the aquifer–atmosphere interactions on a decadal time scale by coupling the physically based U.S. Geological Survey (USGS) Modular Three-Dimensional Groundwater Flow Model (MODFLOW; McDonald and Harbaugh 1988) with a land surface model and a single-column atmosphere model. Their simulation results for a catchment in northeastern Kansas indicate that annually, as large as 20% of evapotranspiration can be drawn from an aquifer. The groundwater-supported fraction of evapotranspiration is higher in drier years, especially when evapotranspiration exceeds precipitation. Another study was undertaken by Liang et al. (2003), who developed a new parameterization to represent surface–ground-water interactions and implemented it into the three-layer variable infiltration capacity (VIC-3L; Liang et al. 1994, 1996; Liang and Xie 2001) model. They tested their new model (called VIC-ground) in Pennsylvania and found significant impacts of groundwater aquifer on the partitioning of water budget components, primarily through changing the vertical soil moisture profile.

3. LSX model simulations (without water table representation)

The land surface parameterization scheme [Land Surface Transfer Scheme (LSX)] developed by National Center for Atmospheric Research (NCAR) scientists (Pollard and Thompson 1995; Thompson and Pollard 1995) is used as the modeling tool in this study. LSX is also the land surface scheme used in the dynamic biosphere model, the Integrated Biosphere Simulator (IBIS; Foley et al. 1996). The performance of LSX has been evaluated by offline simulations conducted at several locations around the world by Delire and Foley (1999). The LSX computes exchanges of momentum, energy, and water mass in the soil–vegetation–atmosphere system with explicit accounts for vegetation effects. The design of the LSX borrows heavily from BATS and SiB, and is intermediate in complexity between these two models. There are two vegetation layers in LSX, an upper layer (“trees”) and a lower layer (“grass”). The main LSX equations simulate vegetation temperature, canopy air temperature, and specific humidity. Another simple set of equations predicts the amount of intercepted water and the precipitation through the canopy. A multilayer soil model is used to simulate the fluxes of energy and water in the upper few meters of soil. For each layer, there are three prognostic variables: temperature, liquid water content, and ice content. The physics governing the movement of liquid water is described by the Richards equation:

$$n\frac{\partial s}{\partial t} = \frac{\partial}{\partial z}\left(-K_{\mathrm{unsat}}(s) + D(s)\frac{\partial s}{\partial z}\right)$$

$$K_{\mathrm{unsat}}(s) = K_S s^{2B+3}; \quad D(s) = K_S \psi_S B s^{B+2}, \qquad (1)$$

where n is the soil porosity; s is the soil saturation degree; K_S and K_{unsat} are the saturated and unsaturated hydraulic conductivity, respectively; ψ_S is the saturated soil suction; B is the empirical soil exponent; and D is the soil diffusion coefficient. The two terms inside the parenthesis of Eq. (1) are the gravity drainage flux and the diffusion (capillary) flux, both of which have high nonlinear dependence on s. The boundary conditions at the bottom of the soil model could be gravity drainage, no flux, or a flux in between determined by specifying an empirical drainage coefficient between 0 and 1. The upper boundary is specified by the effective surface infiltration rate, which equals to rainfall minus evaporation. When the surface layer is saturated, any excess rainfall minus evaporation becomes surface runoff. For other details of the LSX model, see the appendix of Pollard and Thompson (1995) and Thompson and Pollard (1995).

In this section, the ability of the LSX to reproduce the observed hydroclimatology in Illinois during an 11-yr period (1984–94) will be tested. All the simulations conducted in this study are in an offline mode; that is, using the prescribed atmospheric forcings to drive the model rather than coupling LSPs to climate models. The test domain is the whole state of Illinois with the scale ~500 km × 300 km, which is comparable to the typical size of a GCM grid cell.

To drive the LSX in an offline fashion, seven input atmospheric forcing (i.e., precipitation, near-surface air temperature, air humidity, air pressure, shortwave and longwave radiations, and near-surface wind speed) are prepared from either direct observations or the product of the (National Centers for Environmental Prediction) NCEP–NCAR reanalysis. For the details of Illinois data networks, please see Yeh et al. (1998). The sampling interval of the atmospheric forcing is 1 h. Precipitation is retrieved from the EarthInfo, Inc., hourly dataset from which 52 rain gauges in Illinois with data coverage above 90% during 1984–94 are selected to derive the mean precipitation in Illinois. Because of the high density of rain gauges, the arithmetic average of the 52-station precipitation records is considered as the mean precipitation in Illinois. To correct the missing data in the EarthInfo hourly dataset, the total precipitation for each day of the 11-yr period is adjusted to be consistent with the arithmetic average of the 129-station daily precipitation data provided by the Midwest Climate Center (MCC). Moreover, the near-surface air temperature, air humidity, and air pressure are taken from the National Climate Data Center (NCDC) Surface Airway hourly dataset. There are 11 NCDC stations located in or near Illinois that are used to derive the state-average values by simple averaging. In addition, the shortwave and longwave radiations are interpolated from the 6-hourly NCEP–NCAR reanalysis data to the hourly resolution, and then their monthly averages are adjusted to be consistent with the National Aeronautics and Space Administration Surface Radiation Budget (NASA SRB) monthly dataset, which contains global information on both net longwave radiation and net solar radiation. Finally, the near-surface wind speed is taken from the 6-hourly NCEP–NCAR reanalysis data and interpolated to the hourly resolution. The 11-yr (1984–94) average climatologies of the seven atmospheric forcings in Illinois are shown in Fig. 3.

To ensure the simulation results independent of the uncertain initial conditions of hydrologic states (e.g., soil moisture, soil temperature, canopy storage, etc.), several spinup years are necessary to be added ahead of the simulation to bring the system to the hydrological equilibrium after the spinup years. Hydrological equilibrium is reached when the deviations in the annual water and heat balance are negligible from year to year. For the present LSX simulations in Illinois from 1984 to 1994, three spinup years with forcings identical to that of 1984 are found to be sufficient to reach equilibrium. This addition of spinup years has been widely used in the offline testing of LSPs during the different phases of the PILPS. The number of spinup years required depends on the memory of the system, but in general it takes longer to reach equilibrium for a dry condition than for a wet condition.

Model validation is based on a comprehensive hydrologic dataset in Illinois. The dataset includes the 11-yr (1984–94) monthly observations of soil moisture, water table depth, and streamflow (see Yeh et al. 1998). The datasets on soil moisture and water table depth were both provided by the Illinois State Water Survey (ISWS). The groundwater data consist of monthly measurements of water table depth at 18 wells scattered throughout Illinois since the 1960s. These wells were designed for monitoring the response of unconfined aquifers to the precipitation forcing and are all located far away from pumping centers. For other details of the data sampling networks in Illinois, see Yeh et al. (1998,

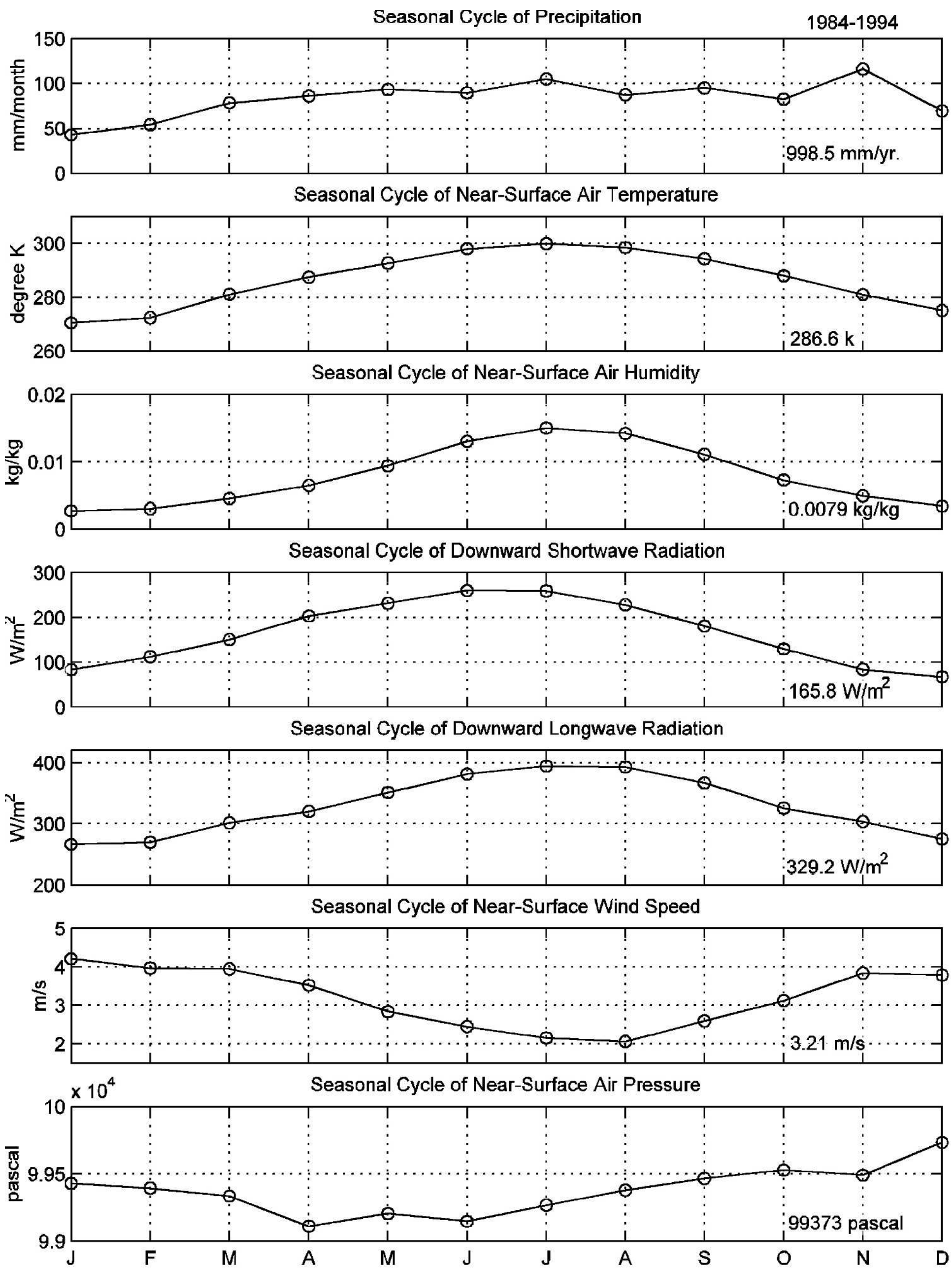

FIG. 3. The 11-yr (1984–94) average climatologies of the seven atmospheric forcings used in the LSX simulations in Illinois.

19 825–19 826). Using the water balance computations from both the soil and the atmospheric branches of hydrology, monthly evapotranspiration (Yeh et. al. 1998) and groundwater recharge (Yeh 2002; see Fig. 2) have been estimated based on the Illinois hydrologic dataset. These estimations will also be used in the model validation process.

The dominant soil type in Illinois is silt loam or silty clay loam inferred from the soil texture descriptions in 19 soil moisture measurement sites provided by Hollinger and Isard (1994). The Illinois soil moisture dataset also includes the measurements of soil porosity, field capacity, and wilting point. The specification of the saturated hydraulic conductivity (K_S), soil exponent (B), and soil water potential at saturation (ψ_S) for the silt loam soils was based on the tabulation of Rawls et. al. (1982). Moreover, the dominant land cover in Illinois is 60% cropland and 20% shortgrass. Since crop is absent in the vegetation lookup table in the LSX, the vegetation cover is specified as the shortgrass with a C3 photosynthetic pathway. This is consistent with Delire and Foley (1999) who used C3 grass in the LSX model to simulate a crop site in the HAPEX-Mobilhy site in France. In addition, a seasonal cycle of leaf area index (LAI) with a maximum in summer and zero in winter is imposed based on the available information of the crop growing characteristics in Illinois (e.g., Vinnikov et al. 1999).

The statistical interception scheme developed by Eltahir and Bras (1993) has been incorporated into the original LSX model. This scheme takes into account the subgrid spatial variabilities of both precipitation and canopy storage. The major outcome of incorporating this scheme is the reduction of the interception loss to a more realistic value, as will be discussed later in this paper.

As discussed earlier, a common weakness in most LSPs is the specification of a gravity drainage condition as the lower boundary condition of soil moisture. In the original LSX model, the bottom boundary condition is specified as the unsaturated conductivity of the lowest soil layer multiplied by an empirical drainage coefficient ranging from 0 (representing no flux, bedrock condition) to 1 (gravity drainage condition). This coefficient dictating the drainage rate has profound impacts on the model's partitioning of precipitation into runoff and evaporation. As a result of the difficulty of measuring this coefficient in the field, it can only be treated as a tuning parameter without any realistic physical meaning. To investigate the sensitivity of the lower boundary condition, two contrasting 11-yr (1984–94) offline simulations, one with the "gravity drainage condition" at the soil bottom (hereafter referred to as case A), and the other with the "no-flux (i.e., bedrock) condition" (case B), were conducted. Both simulations were driven by the identical atmospheric forcings and parameters as summarized above. These two extreme cases are selected here to illustrate the sensitivity of model simulation to the specification of the lower boundary condition of the soil model in LSPs.

Simulation results of case A are presented in the following. Figure 4a shows the comparison between the 11-yr (1984–94) average seasonal cycle of the simulated total runoff and the observations. Figures 4b–c plot the seasonal cycles of two runoff components in the LSX, namely soil drainage and surface runoff. Figures 4d–e show the 11-yr monthly time series of soil drainage and total runoff in comparison with the corresponding monthly observations. Notice that the "observed" soil drainage plotted in Figs. 4b and 4d is actually the groundwater recharge in Fig. 2 estimated from water balance computations (Yeh 2002) rather than a direct observation. As seen from Fig. 4a, runoff is underestimated from June through October by 5–10 mm month^{-1}. Since surface runoff contributes only a small percentage of total streamflow, the underestimation is primarily caused by the small magnitude of the soil drainage. A notable example observed in Fig. 4e is the extended dry summer span from 1987 to 1989 when streamflow was largely sustained by base flow. Without groundwater representation in the model, the simulated streamflow in summer is significantly underestimated because of the incorrect procedure in LSPs of treating the soil drainage equivalent to the subsurface runoff. Moreover, soil drainage was poorly simulated with the range of its seasonal cycle considerably smaller than the observations. Although the observed groundwater recharge exhibits a negative peak during summer months, the LSX fails to reproduce any upward water flux into the soil bottom during the 11-yr period owing to incorrect specification of the gravity drainage boundary condition. In addition, the simulated soil drainage is significantly underestimated in winter and spring and unable to capture the double peaks that occurred in November and March.

Figure 5 provides similar comparisons with the observations, but for case B (no-flux boundary condition). In this case, runoff is contributed solely from surface runoff since soil drainage is completely shut down. The overall pattern of the negative biases in the total runoff simulation is similar to that in case A, but with even larger deviations from the observations. For example, the streamflow peaks were significantly overestimated (1984, 1986, and 1990), while nearly zero streamflow was simulated during dry periods from 1988 to 1990.

The seasonal cycles of the soil saturation degree in 11 soil layers (consistent with the vertical sampling intervals of soil moisture data in Illinois) are compared to the observations in Figs. 6 and 7 , respectively, for cases A and B. As is clearly shown, the simulated soil moisture in either case is not close to the observation. For case A (Fig. 6), there is a dry bias in the simulated soil saturation for the layers below 70 cm from the surface. Because of the absence of water table, the drainage flux out of the soil column is overestimated, which is responsible for the underestimated soil moisture in these layers. In contrast, a wet bias is noted in each of the 11 soil layers for case B (Fig. 7). As a result of the complete shutdown of drainage, the simulated soil moisture below 1 m from the surface is close to saturation for the entire year. The wet bias in the lower soil layers propagates to the upper layers, which leads to unrealistically large amounts of surface runoff in case B, as already shown in Fig. 5.

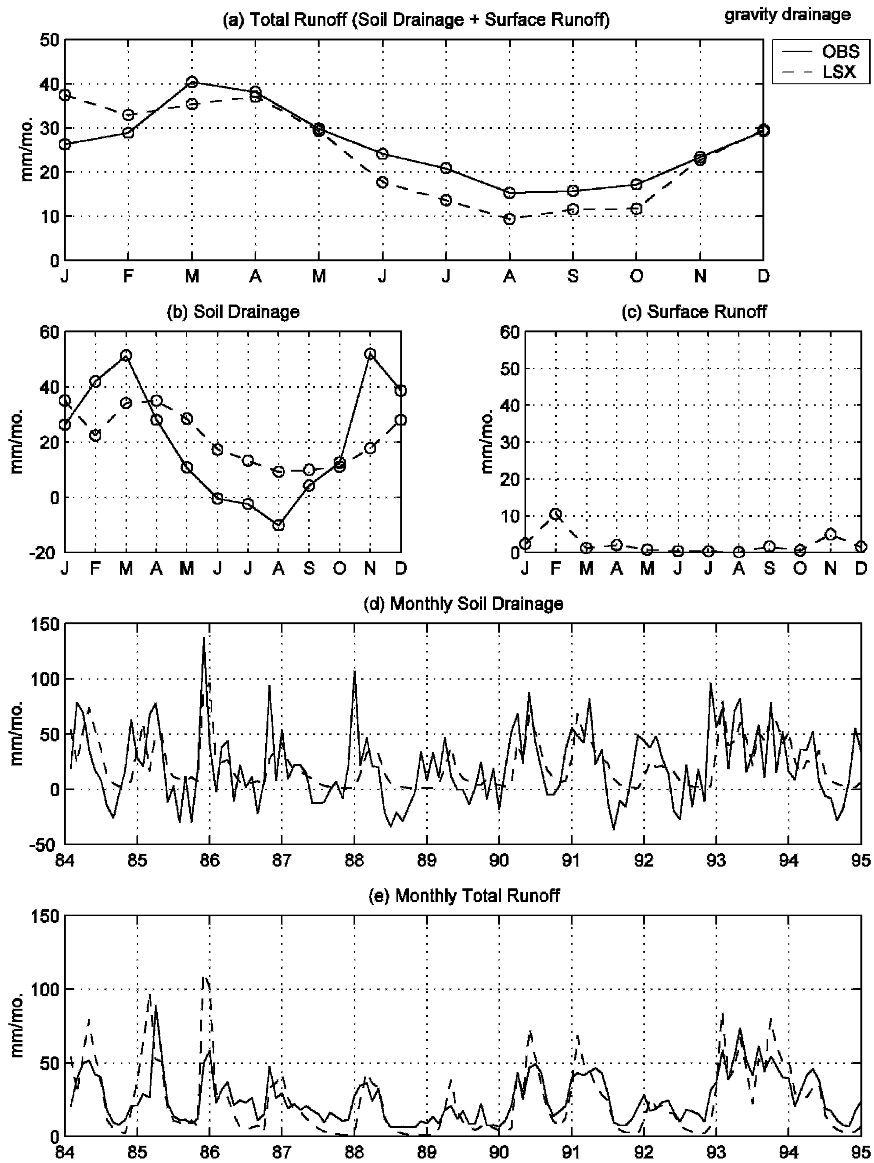

FIG. 4. Case A (i.e., gravity drainage boundary condition): 11-yr (1984–94) average seasonal cycles of (a) the simulated total runoff in comparison with the observations, (b) the simulated soil drainage in comparison with the estimates of groundwater recharge from water balance computations, and (c) the simulated surface runoff. The 11-yr monthly time series of (d) soil drainage and (e) total runoff, respectively, in comparison with the monthly time series of observations.

The model performance is also evaluated at the annual time scale for the gravity drainage case. Figure 8 plots the evaporation ratio (i.e., annual total evaporation/precipitation) and runoff ratio (i.e., annual total runoff /precipitation) from 1984 to 1994 for case A in comparison with the observed ratios. It can be observed from this figure that the LSX with the gravity drainage condition can simulate the correct ratios only in 1988–90 and 1993, while large discrepancies can be noted for the rest of the 11 yr.

As discussed earlier, most of the current LSPs use free drainage to generate subsurface runoff. In the PILPS 2c experiment in the Red–Arkansas River basin, all the participating LSPs were allowed to calibrate any of their own model processes and parameters, but none of those 13 LSPs with the free drainage condition attempted to modify this unrealistic condition (see Table 4 of Wood et al. 1998). The gravity drainage condition in most LSPs has been treated as a standard parameterization regardless of variable subsurface conditions that

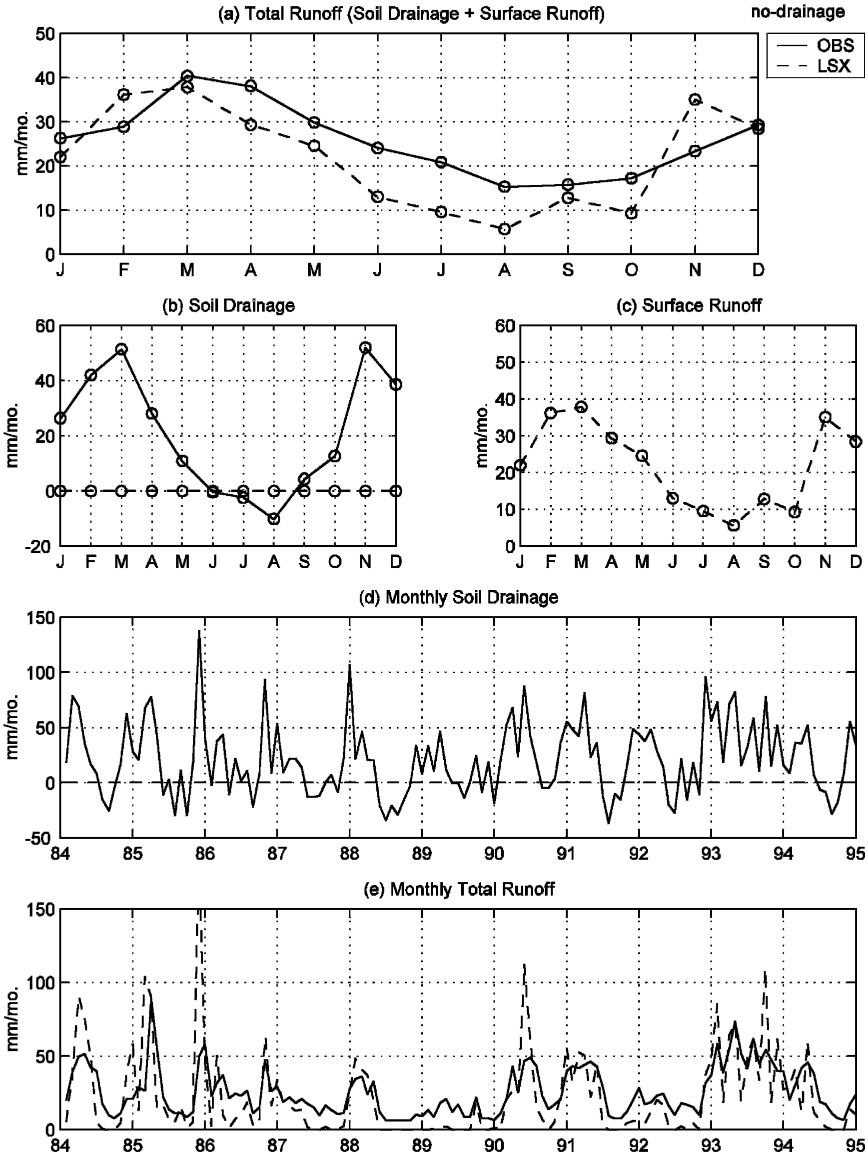

FIG. 5. Same as in Fig. 4, but for case B (i.e., no-flux boundary condition).

might be encountered in reality. Since the LSX uses a tunable drainage coefficient, it would be interesting to explore the feasibility of treating this empirical coefficient as a calibration parameter. Figure 9 compares the soil drainage (groundwater recharge) fluxes simulated by LSX using drainage coefficients 0.25, 0.5, 0.75, and 1 (free drainage) with the corresponding observations. As clearly shown in this figure, the difference between using various coefficients is negligible: the average annual drainage fluxes are 270, 275, 278, and 279 mm yr^{-1} for the drainage coefficients of 0.25, 0.5, 0.75, and 1, respectively. As expected, none of these cases can reproduce the net upward water fluxes that occurred in 38 months of the 11-yr period. The major reason for the lack of sensitivity is the negative feedback between the drainage rate and the bottom soil saturation that originated from their nonlinear dependence: a larger (smaller) drainage coefficient would decrease (increase) the bottom soil saturation, which would subsequently reduce (increase) the drainage rate. A closer inspection (not shown) reveals that the bottom soil moisture is slightly drier for the cases with larger drainage coefficients, while the drainage flux remains nearly constant in all cases.

The soil drainage flux between the unsaturated and saturated zones is not always downward in shallow water table areas. Rather, it has a strong seasonal cycle with the upward water flux occurring when capillary

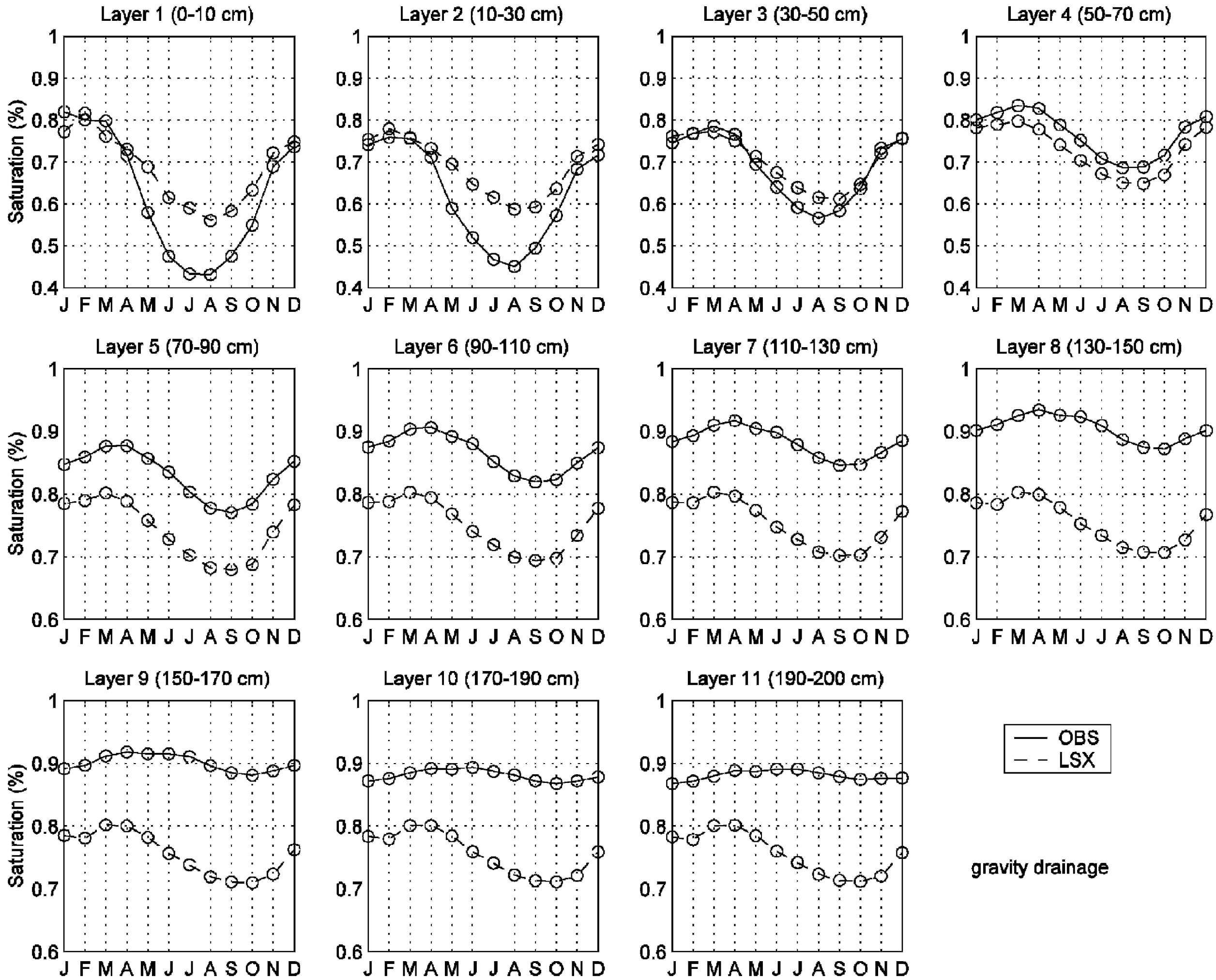

FIG. 6. The 11-yr (1984–94) seasonal cycles of the simulated soil saturation degree for case A (i.e., gravity drainage boundary condition) in 11 soil layers from 0 to 2 m below the surface in comparison with the observations.

diffusion outweighs gravity drainage as a result of the strong soil moisture deficit. Most LSPs, however, do not allow for the occurrence of the upward water fluxes at the soil bottom. The most critical outcome caused by the inappropriate boundary condition specification is the erroneous soil moisture profile that would affect the prediction of land surface water and energy fluxes. Compared to a deep water table condition, the presence of shallow water table in general results in a decrease of infiltration and an increase of evapotranspiration due to downward increase of moisture content from the root zone to the water table. Therefore, we conclude that only by explicitly incorporating the water table dynamics in LSPs can we have a realistic representation of land surface hydrology and hence a reliable partitioning of the land surface water and heat budgets. For this purpose, a simple groundwater model designed for use in climate models will be proposed in the next section, and the procedures of its coupling with the LSX will be presented. The coupled model will be tested in Illinois using the same atmospheric forcing and validation data as used in this section in order to investigate the improvement in model performance resulting from the introduction of water table representation.

4. Development of an unconfined aquifer model

A physically based groundwater flow model (e.g., MODFLOW; McDonald and Harbaugh 1988) usually requires the discretization of the problem domain into a certain number of cells and numerically solves the multidimensional partial differential equations governing groundwater movement. Because of the large scale of a typical grid cell in climate models, such complicated numerical computations may be too expensive to be applied in climate models and also too detailed given the objective of the model development. An alternative

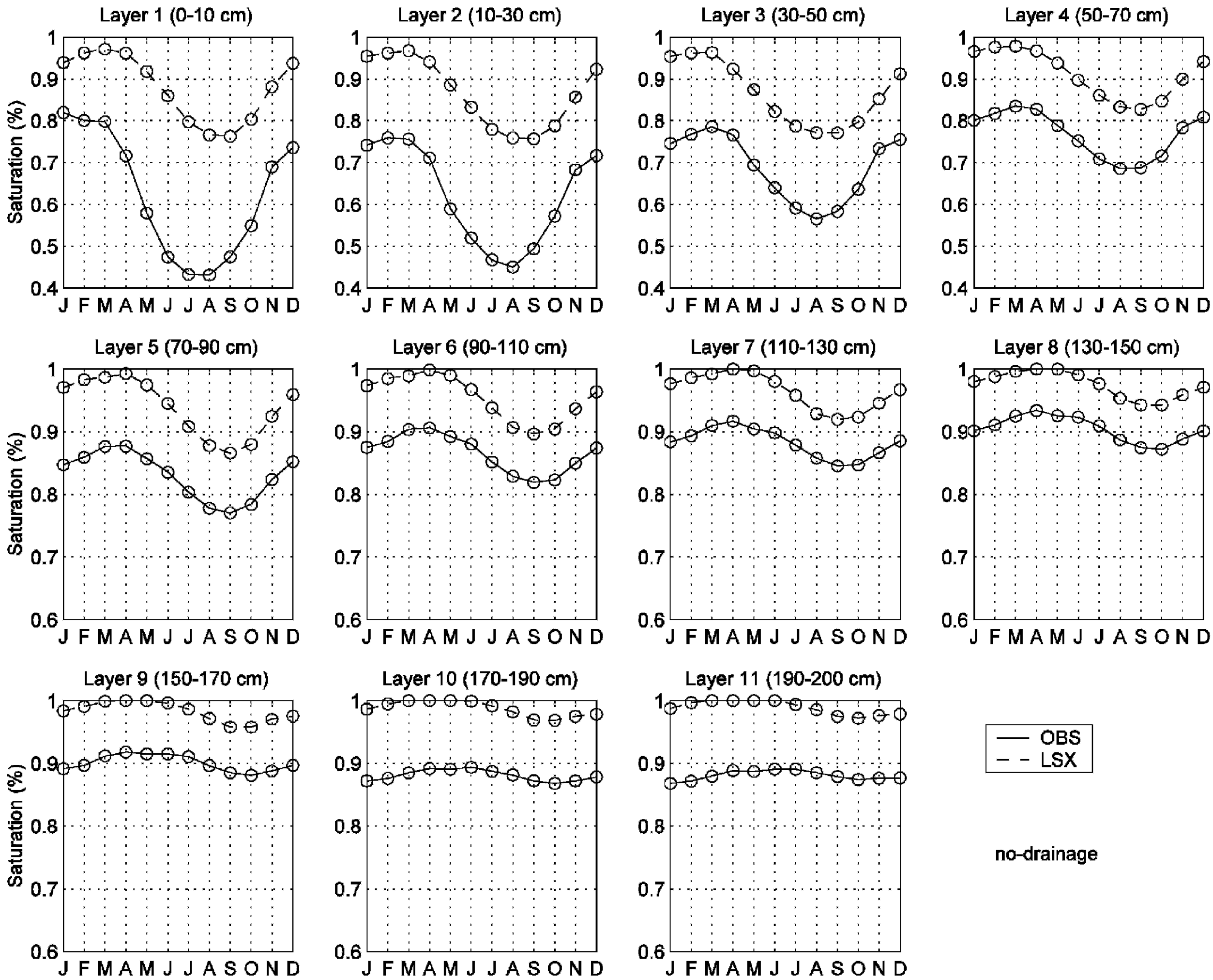

FIG. 7. Same as in Fig. 6, but for case B (i.e., no-flux boundary condition).

with less computational cost and parameter requirement is the lumped-parameter water balance model. For an unconfined aquifer, the water balance equation can be written as:

$$S_y \frac{dH}{dt} = I_{gw} - Q_{gw}, \tag{2}$$

where S_y is the specific yield of the unconfined aquifer, H is the groundwater level above the datum, I_{gw} is the groundwater recharge flux, and Q_{gw} is the groundwater discharge to streams (i.e., groundwater runoff). For the silt loam soil typical in Illinois, the typical value of S_y is specified as 0.08 based on the specific yield data compiled by Johnson (1967). Notice that I_{gw} is also the soil drainage flux at the interface between the unsaturated and the saturated zone [i.e., the terms inside the parentheses of Eq. (1)], which is the sum of the gravity drainage flux from the soil to the aquifer and the diffusion flux from the water table to the soil. Here, I_{gw} is the key process controlling the unsaturated–saturated zone interaction since it is the linkage connecting these two zones.

Equation (2) is a nonlinear reservoir since both I_{gw} and Q_{gw} have complex, nonlinear dependences on H; I_{gw} depends on the soil saturation degree in the bottommost soil layer calculated by numerically solving the Richards equation has an implicit dependence on H since the vertical soil moisture profile is closely dependent on the location of the water table. As shown in Fig. 1, there is a strong nonlinear relationship between the observed large-scale average streamflow and the average water table depth in Illinois. Examination of the large-scale observed precipitation, water table depth, and streamflow (Fig. 1) suggests that monthly average streamflow is significantly more correlated to ≥ water table depth than to precipitation. While average water

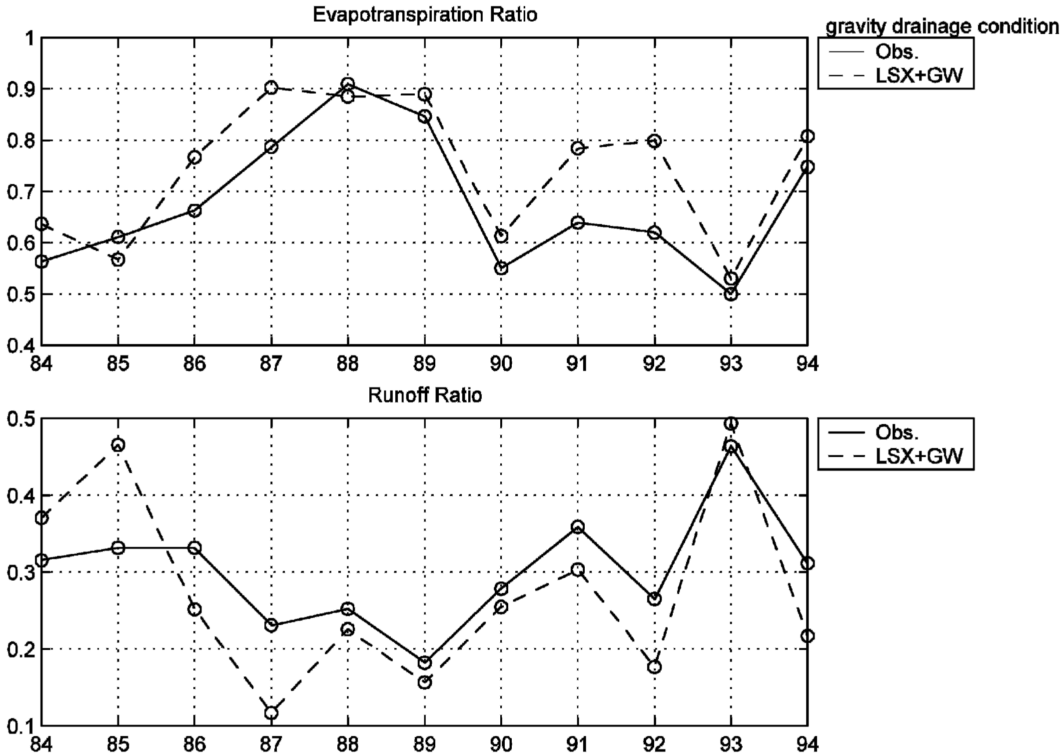

FIG. 8. The evaporation ratio (i.e., annual total evaporation/precipitation) and runoff ratio (i.e., annual total runoff/precipitation) from 1984 to 1994 for case A (i.e., gravity drainage boundary condition) in comparison with the observed ratios.

table depth explains more than 60% of the variance of monthly streamflow, precipitation fails to explain more than 10% of the same variance (Eltahir and Yeh 1999). Moreover, the contribution of surface runoff to streamflow at the scale of Illinois is small primarily as a result of the reduced magnitude of the spatially averaged precipitation. Therefore, the pattern of the observed streamflow should be a good surrogate for that of groundwater runoff; hence it is reasonable to expect that Fig. 1b describes the general features of the nonlinear dependence between Q_{gw} and H reasonably well.

For simplicity, since the large-scale data on water table depth are available, the Q_{gw}–H dependence is sought by using a regression analysis performed with respect to the 11-yr Illinois data in Fig. 1b. The following regression equation between Q_{gw} (in mm month^{-1}) and the depth to water table, D_{gw} (in meters), is derived:

$$Q_{gw} = \frac{164.7}{D_{gw}^2} - 1.4, \tag{3}$$

with the correlation coefficient of 0.84. We have also attempted to change the exponent of the Q_{gw}–D_{gw} relationship from –1 to −4 and found that the inverse square dependence yields the highest correlation coefficient. A rigorous theoretical development of groundwater runoff formulation taking into account the spatial variability of water table depth will be proposed in the companion paper by Yeh and Eltahir (2005).

The developed unconfined aquifer model in Eqs. (2) and (3) is interactively coupled with the LSX. The soil model in the LSX is properly modified to accommodate the unconfined aquifer model. In the original LSX, the boundary condition at the soil bottom is the drainage flux. To locate the lower boundary of the soil column at the water table, the flux boundary condition in the LSX is modified from the flux boundary condition to the head boundary condition by specifying that the soil saturation degree in the lowest soil layer equals unity. Notice that such a boundary is actually where the top of the capillary fringe rather than the water table is located. However, it is trivial to convert the top of the capillary fringe to the actual water table depth by approximating the thickness of the capillary fringe. For the silt loam soils typical in Illinois, the thickness of the capillary fringe is about 50 cm (Philip 1969).

By coupling Eq. (2) to the single soil column model in the LSX, the total length of the unsaturated soil column varies in response to the fluctuation in water table depth. Therefore, the number of the unsaturated soil

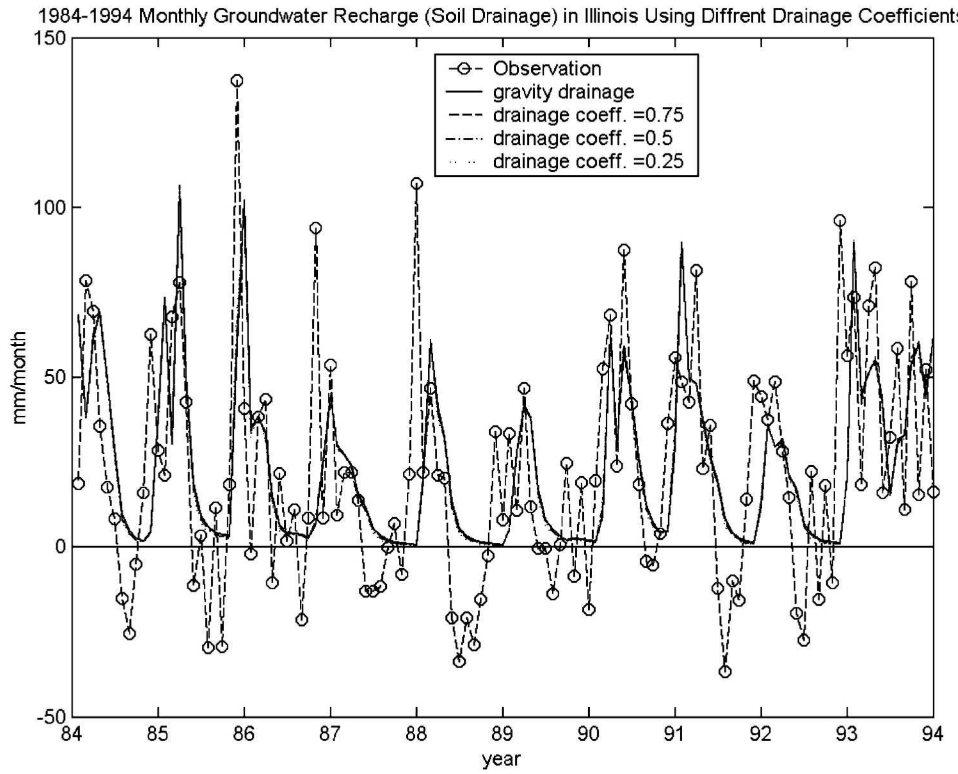

FIG. 9. The 1984–94 simulated monthly time series of groundwater recharge (soil drainage) in Illinois by using different drainage coefficients in the LSX model in comparison with observations.

layers varies with time since the vertical resolution of the soil model is fixed. To locate the water table position more accurately and capture the sharp moisture gradient near the water table, a fine-resolution soil model is adopted. The specification of the soil layer geometry is flexible, and the number of total layers is determined by the water table condition and the bedrock position. Here, a total 50 soil layers are used: 10 cm each for the first 1 m of soils and 20 cm each from below 1 to 9 m. If the water table locates within the layer n for a specific time step, the soil layers from 1 to (n-1) are unsaturated, and the soil moisture in these (n-1) layers would be solved by the Richards equation at that time step. If the water table falls below 9 m, the soil model and aquifer model are basically decoupled and the gravity drainage boundary condition applies. This treatment is advantageous because it does not specify unjustified common boundary conditions such as gravity drainage or no flux conditions. Another reason in favor of the fine-resolution model is because the magnitude of diffusion flux depends on the distance between the centers of two adjacent bottom layers [see Eq. (1)]; thus the thickness of soil layers should be as thin as possible in order to remove this sensitivity.

For each time step, the groundwater model simultaneously receives the drainage flux from the overlying soil column (i.e., groundwater recharge) and discharges runoff into streams. The water table position is updated accordingly at the end of each time step. The direction of the soil drainage flux can be downward or upward depending on the relative magnitude of the gravity drainage and diffusion fluxes. For the next time step, the soil moisture is updated according to the new position of the water table by solving the Richards equation. By doing so, the interactions between the unsaturated and saturated zones are explicitly represented.

From Eq. (2), if I_{gw} is negative (upward flux from the water table), the water table would go down since Q_{gw} is always positive. However, if both I_{gw} and Q_{gw} are positive, the water table may go up or down depending on the relative magnitude of these two fluxes. In either case, it forms a negative feedback that brings the water table back to its equilibrium position. When the water table goes up, Q_{gw} begins to increase and I_{gw} begins to decrease as a result of the stronger upward diffusion flux caused by the steeper moisture gradient. Both mechanisms have the effect of bringing the water table down. On the other hand, when the water table goes down, Q_{gw} decreases and I_{gw} increases because of the weaker diffusion flux as a result of the smoother soil moisture gradient. The effect of both mechanisms is to increase the amount of water in the aquifer and hence to raise the water table.

The coupled model, called LSXGW, has been tested in Illinois using identical atmospheric forcing and model parameters as those used in the LSX simulation presented in section 3. The results for the 11-yr (1984–94) LSXGW simulation will be presented in the next section.

5. LSXGW model simulations (with water table representation)

The coupled model LSXGW has been tested in Illinois using the 11-yr (1984–94) identical atmospheric forcing as used in the section 3. The results are summarized in this section. First, the observed and simulated evaporation ratio and runoff ratio from 1984 to 1994 are plotted in Fig. 10. In general, the LSXGW reproduces the observed interannual variability reasonably well, although small deviations (~10%) from observations remain during 1991–93. The comparison of this figure with Fig. 8 (without water table representation) indicates that the partitioning of precipitation into evaporation and runoff is significantly improved after incorporating the water table dynamics. Figure 11 shows the (11 yr) average seasonal cycles of the simulated water table depth, soil drainage (i.e., groundwater recharge), and the total runoff (as well as its two components, groundwater runoff and surface runoff) in comparison with the observations. Figure 12 shows the seasonal cycles of the total evaporation and its three components (transpiration, soil evaporation, and interception loss) in comparison with the observations. Moreover, the seasonal cycles of the observed and simulated soil saturation in 11 soil layers are compared in Fig. 13.

The overall reasonable agreement between the simulation and the observation in Figs. 10–13 indicates that the incorporation of the groundwater model has improved the simulated land surface hydrology. The seasonal cycles of the simulated total runoff as well as soil drainage match the observations significantly better than the original LSX (Figs. 3 and 4). The simulated soil moisture profile by the LSXGW (Fig. 13) has also improved the large biases noted in the LSX (Figs. 6 and 7).

The 11-yr monthly time series of soil drainage (groundwater recharge) simulated by the LSX (gravity

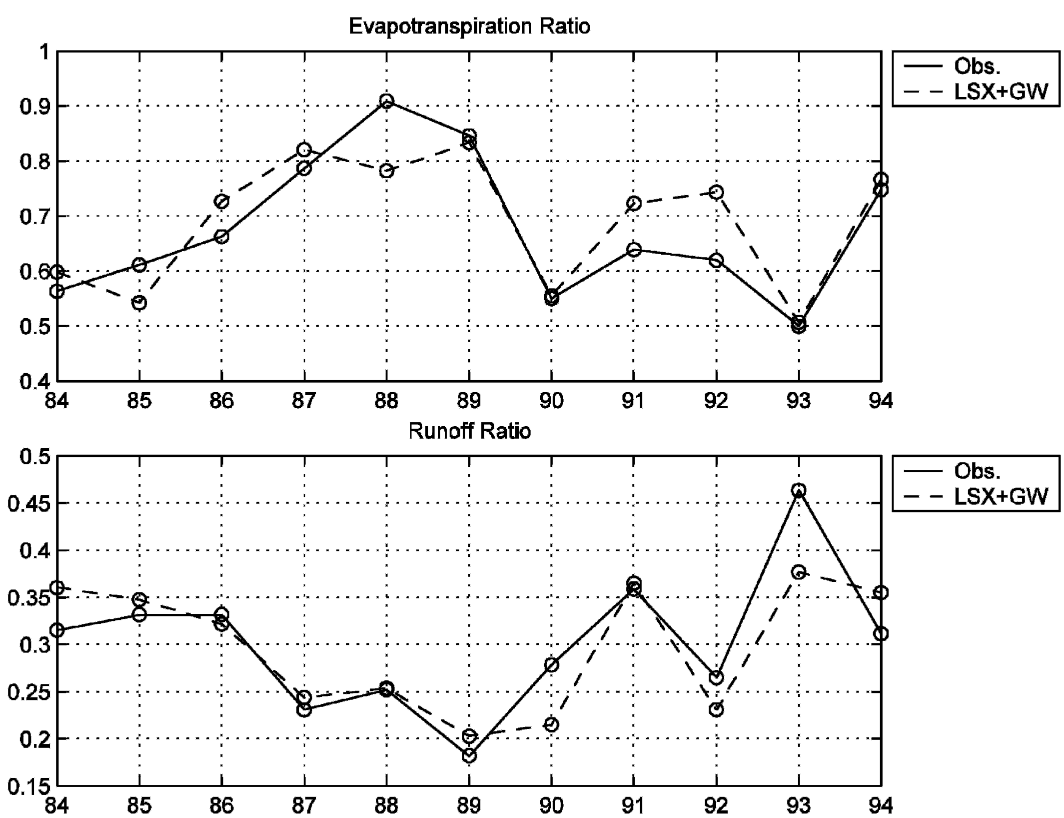

FIG. 10. The simulated evaporation ratio and runoff ratio from 1984 to 1994 in comparison with the observations.

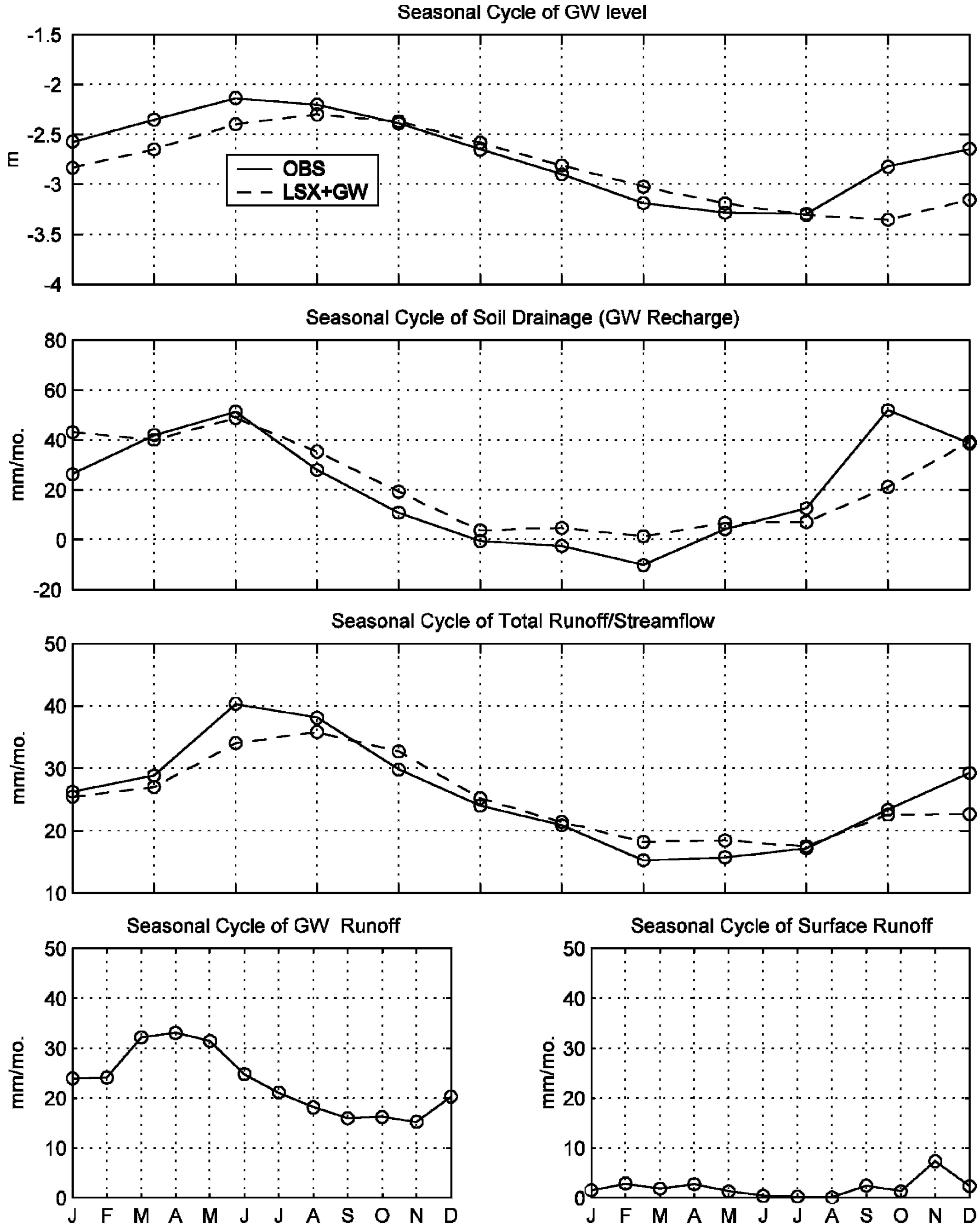

FIG. 11. The 11-yr (1984–94) average seasonal cycles of the simulated water table depth, soil drainage (i.e., groundwater recharge), and total runoff from the LSXGW simulation in comparison with the corresponding observations. The seasonal cycles of two runoff components, groundwater runoff and surface runoff, are also shown in this figure.

drainage) and the LSXGW are plotted together in comparison with observations in Fig. 14. As seen from this figure, the LSXGW indeed reproduces small upward water flux in some months during the 11-yr period. This improvement is attributed to the introduction of exact water table position in determining the vertical soil moisture profile. It should also be noted, however, that the observed upward water flux is still not well simulated by the LSXGW in the summer months of 1985, 1988, and 1991. One possible reason for this bias is that the impacts of the spatial and temporal variability of water table depth were not accounted for in the groundwater runoff parameterization in Eq. (3). The regression relationship in Eq. (3) was derived solely

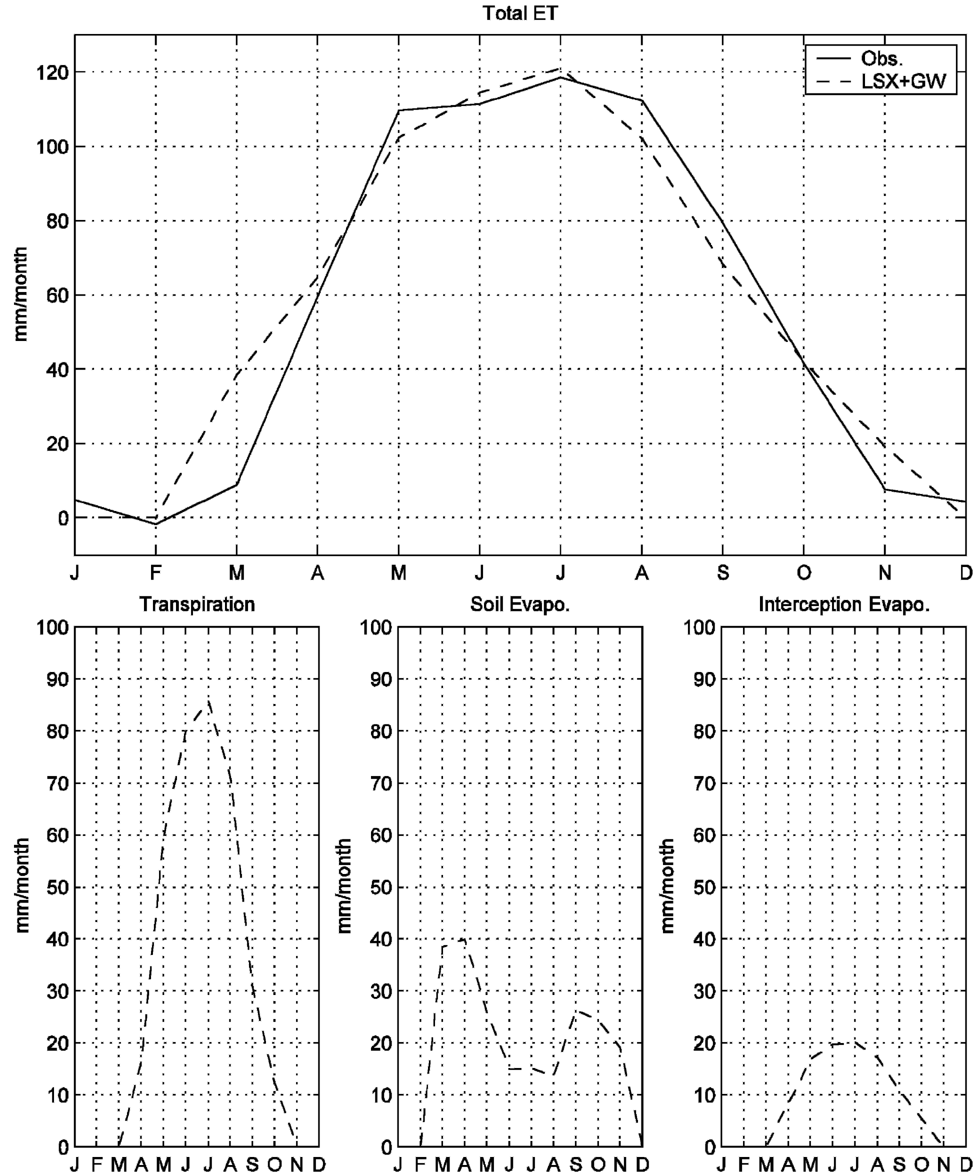

FIG. 12. The 11-yr (1984–94) average seasonal cycles of total evaporation and its three components: transpiration, soil evaporation, and interception loss. The observation in this figure is the evaporation estimate from the water balance computations conducted by Yeh et al. (1998).

based on the state-average, long-term averaged relationship between water table depth and streamflow in Illinois. The deviation of local conditions from the averaged relationship in Eq. (3) needs to be incorporated in order to simulate the correct water table dynamics. For example, the severe drought condition in 1988 resulted in a larger-than-average decrease in groundwater level for some regions in Illinois. Since the LSXGW only simulates the average water table depth over Illinois, the effects of the anomalously low water table depth in some subareas in Illinois cannot be incorporated. Similarly, the upward water fluxes from the aquifer to the unsaturated zone take place only in the shallow water table regions. Without the consideration of the spatial variability of water table depth, the LSXGW cannot simulate the correct average magnitude of upward water fluxes. Moreover, the groundwater rating curve (i.e., aquifer storage–discharge relationship) in Eq. (3) is estimated from the water table depth observations in Illinois, which would not be routinely available in other areas. A rigorous theoretical development and the associated parameter estimation technique of the large-scale groundwater rating curve are needed when the observations are not available.

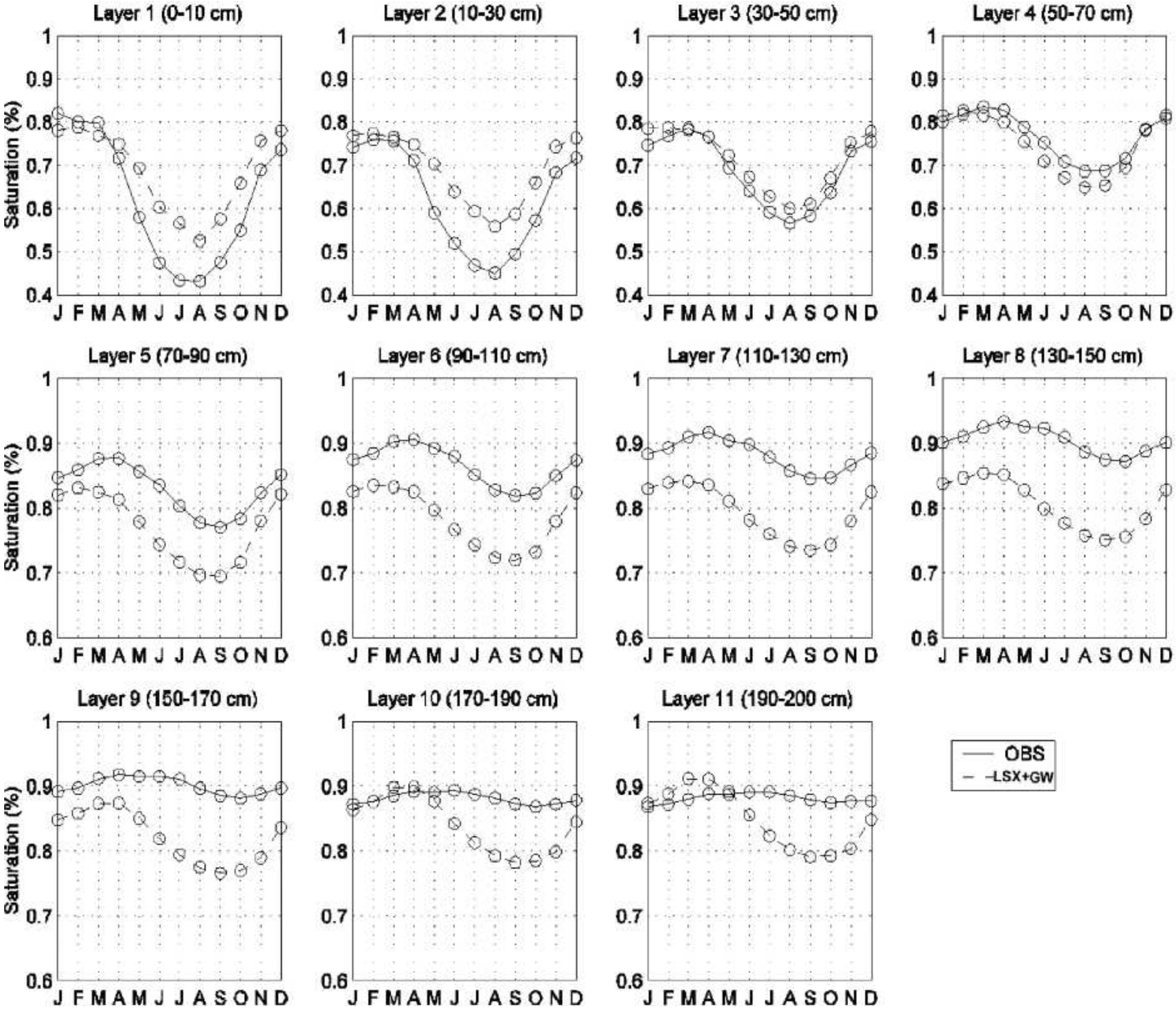

FIG. 13. The 11-yr (1984–94) average seasonal cycles of the observed and the simulated soil saturation degrees in 11 soil layers from 0 to 2 m below the surface.

6. Summary

The current generation of land surface parameterization schemes (LSPs) used in climate models neglects the representation of water table dynamics. For shallow water table areas, such a simplified representation of subsurface hydrology would result in significant errors in the predicted land surface fluxes and states. It has been demonstrated through two 11-yr offline simulations that the LSX, which is the representative of current LSPs without any representation of water table dynamics, fails to accurately simulate the land surface hydrology in shallow water table regions such as Illinois.

In an attempt to incorporate water table dynamics into LSPs, a simple lumped unconfined aquifer model is developed and interactively coupled to the land surface model LSX. This coupled model (called LSXGW) has been tested in Illinois where the availability of a comprehensive hydrological dataset provides a unique opportunity for model validation. The simulation results for an 11-yr period from 1984 to 1994 indicate that most of the simulated hydrological states (e.g., water table depth and soil saturation) and fluxes (e.g., groundwater recharge, runoff, and evaporation) agree with the hydrological observations in Illinois reasonably well. The proposed aquifer model is suitable to be used in climate models in that it is parsimonious in terms of the required parameters and computationally efficient because of its lumped-parameter nature. Moreover, it was designed in a flexible way such that it can be easily coupled to any land surface scheme with very little modifications of the soil model codes. However, there

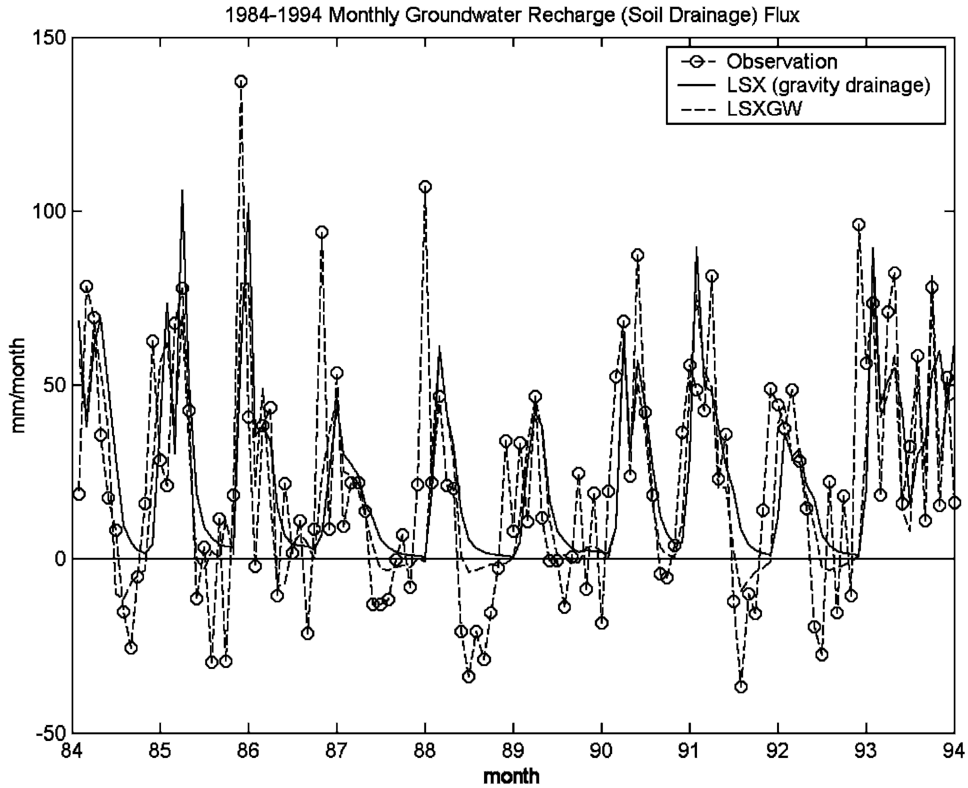

FIG. 14. The 1984–94 monthly time series of groundwater recharge (soil drainage) simulated by the LSX using the gravity drainage and by the LSXGW in comparison with the corresponding monthly observations in Illinois.

are still certain disagreements between the LSXGW simulations and the observations, possibly due to the spatial variability of water table depth. Because of the nonlinearity in most land surface hydrological processes, the spatial variability in the land surface variables needs to be parameterized in the LSPs. The issue of accounting for the subgrid variability of water table depth in the LSXGW will be addressed in a companion paper Yeh and Eltahir (2005).

As a final note, one may argue that the imperfect model process representations in LSPs can be alleviated by a better calibration of associated model parameters or can be disregarded as a result of the limitation in the availability of observational data. The present study denies these arguments by showing the critical importance of the presence of shallow water table in affecting the near-surface soil moisture profile, and hence the numerous hydrological processes associated with the soil wetness condition. For example, the errors caused by the free drainage assumption, which would reverse the direction of flow in the root zone, may not be compensated by parameter calibration.

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