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Hydroclimatology of Illinois: A comparison of monthly evaporation estimates based on atmospheric water balance and soil water balance

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Abstract. Here we describe the regional-scale hydrological cycle of Illinois, including both the land and atmospheric branches, using a data set on most of the hydrological variables, i.e., precipitation, streamflow, soil water content, snow depth, groundwater level, and atmospheric flux of water vapor. Since direct observations of evaporation are not available, two different approaches, soil water balance and atmospheric water balance, were applied to estimate the regional evaporation over Illinois from 1983 to 1994. The availability of a comprehensive hydrological data set covering the large area of Illinois facilitated a comparison between these two approaches for estimation of evaporation. To our knowledge, this is the first time such a comparison has been made. The climatologies of the monthly evaporation estimates from the two approaches agree reasonably well and within a 10% error; however, substantial differences exist between the two estimates of evaporation for individual months. The seasonal variability of the evaporation estimates based on soil water balance is largely balanced by the seasonal pattern of subsurface storage, whereas the seasonal variability of evaporation estimates from the atmospheric water balance is almost entirely balanced by the seasonal pattern of lateral fluxes of water vapor. This contrast reflects a fundamental difference in the hydrology of the land and atmospheric branches of the regional water cycle. In light of the fact that independent data sets were used in the two approaches, our results are encouraging: The atmospheric water balance approach has the potential for the accurate estimation of the climatology of regional evaporation, at least for humid regions at a scale similar to that of Illinois ($\sim 10^5$ km^2). However, sensitivity analysis suggests that the accuracy of atmospheric water balance computations is rather poor for the scale smaller than 10^5 km^2. For the calculation of evaporation using the soil water balance approach in regions where the groundwater table is rather shallow, the incorporation of the change in groundwater storage is indispensable since groundwater aquifers provide a significant portion of water storage at the monthly timescale.

1. Introduction

Evaporation refers to the part of precipitation converted into water vapor and subsequently transported into the atmosphere [*Shuttleworth*, 1979; *Brutsaert*, 1982]. Evaporation together with precipitation govern the amount of runoff available to the planning and management of water resources; hence accurate knowledge of evaporation is required in the design of water resources systems such as storage reservoirs, agricultural irrigation schemes, municipal and industrial water supply systems, etc. In addition, evaporation links the land and atmospheric branches of the hydrological cycle by dictating exchanges of heat and moisture between the atmosphere and the land surface, thus exerting a strong influence on the patterns of atmospheric water vapor transport that are essential in shaping our climate. Evidence from general circulation model (GCM) experiments has suggested that the Earth's climate is sensitive to variations in regional evaporation [*Shukla and Mintz*, 1982; *Delworth and Manabe*, 1988]; hence it is important for climate change studies to model evaporation accurately. However, since evaporation is difficult to estimate at a regional scale, GCM studies have used parameterizations of evaporation that are difficult to validate owing to the lack of data at a compatible scale.

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Traditionally, point measurements of evaporation can be performed with lysimeters [*Pruitt and Angus*, 1960] or evaporation pans [*Gangopadhyaya et al.*, 1966]. While such ground-based measurements of evaporation could be expensive and laborious, they become especially impractical when regional-scale coverage is of interest. Estimation of regional-scale evaporation still relies on modeling techniques. Perhaps the most conceptually simple tool for modeling regional evaporation is through a classical water balance computation. *Thronthwaite* [1948] and *Penman* [1950] pioneered studies of land-surface water balance for small catchments. *Baumgartner and Reichel* [1975] studied the world water balance and developed global maps for the distributions of precipitation, evaporation, and runoff. These and other water balance computations have been discussed by *Street-Perrott et al.* [1983] and *Van der Beken and Herrmann* [1985]. Although numerous previous studies have attempted to estimate land surface water balances (e.g., *McGowan and Williams* [1980], *Alley* [1984], *Ernstberger and Sokolleck* [1985], *Kattelmann and Elder* [1991], *Vandewiele et al.*

[1992], *Lesack* [1993], *Xu and Halldin* [1996], *Roberts and Harding* [1996], and *Malek et al.* [1997], among others), most of these studies have focused on watersheds with the scale from 10^{-1} km^2 to 10^3 km^2. Primarily due to the paucity of data with adequate resolution and enough length, classical land surface water balance has seldom been applied to the estimation of evaporation at a regional scale (10^4 km^2–10^6 km^2).

As an alternative, numerous studies have attempted to estimate large-scale evaporation from the relatively abundant aerological data, utilizing either the atmospheric water balance concept or the similarity theory of planetary boundary layer properties. *Benton et al.* [1950] and *Benton and Estoque* [1954] pioneered studies of large-scale atmospheric water balance. *Benton et al.* [1950] focused on the water balance of the Mississippi Basin, considering both the atmospheric and the land surface branches of the hydrological cycle. *Benton and Estoque* [1954] studied the transfer of water vapor over the North American continent and estimated a water balance for this region. This approach was followed by *Rasmusson* [1967, 1968, 1971], who estimated the continental-scale water balance for the region of North America using both the atmospheric and the terrestrial control volumes. Monthly values of evapotranspiration and change in storage of soil moisture were estimated on the basis of observed values of atmospheric water vapor divergence, precipitation, and streamflow. *Rasmusson* [1977] reviewed the studies on the atmospheric water vapor transport, recommending the general application of vapor flux data in the routine computation of regional water balances, and the subsequent development of monthly, seasonal, and long-term regional water balance climatologies. In addition, the atmospheric water balance approach has been applied at continental or global scale by *Starr and Peixoto* [1958], *Rosen and Omolayo* [1981], *Peixoto et al.* [1982], *Peixoto and Oort* [1983], *Salstein et al.* [1983], *Bryan and Oort* [1984], *Savijarvi* [1988], *Brubaker et al.* [1994], *Roads et al.* [1994], *Oki et al.* [1995], *Higgins et al.* [1996], *Mo and Higgins* [1996], *Gutowski et al.* [1997] and *Ropelewski and Yarosh* [1998], among others. A complete introduction on the theoretical basis of the atmospheric water balance computations was given by *Peixoto and Oort* [1992, chapter 12]. Furthermore, in recent years there has been interest in estimation of evaporation from large areas using the planetary boundary layer measurements [e.g., *Mawdsley and Brustaret*, 1979; *Abdulmumin et al.*, 1987; *Munley and Hipps*, 1991; *Sugita and Brutsaert*, 1991]. For a detailed discussion on the estimation of evaporation using these techniques, see the review by *Parlange et al.* [1995].

Apparently, there is an inconsistency in the scale to which the soil water balance (SWB) approach and the atmospheric water balance (AWB) approach have been applied. Although the AWB approach has been employed since *Benton et al.* [1950], it has never been validated through a comparison of regional evaporation estimates from direct measurements or other independent studies, for example, regional-scale water balance computed from the terrestrial branch of hydrology. To our knowledge, the only efforts made in this direction are those of *Abdulla et al.* [1996] and *Berbery et al.* [1996]. *Abdulla et al.* [1996] found a favorable agreement between the evaporation estimates based on a land-surface parameterization scheme VIC-2L [*Liang et al.*, 1994] and the estimates based on the AWB approach. *Berbery et al.* [1996] applied the National Center for Environmental Prediction's Eta model to study the seasonal variability of the atmospheric branch of the hydrological cycle in continental North America. Evaporation estimates obtained from the AWB approach were compared to the estimates produced from the Eta model. Still, the scarcity of data from ground truth measurements (especially soil moisture measurements) has limited the validation of the AWB approach using direct observations.

The soil moisture data collected in Illinois since 1981 [*Hollinger and Israd*, 1994], when used in conjunction with other available data sets (i.e., precipitation, snowfall, groundwater level, and streamflow), provide a unique opportunity to estimate regional evaporation in Illinois using the SWB approach. More importantly, these estimates can be compared to those obtained from the AWB approach. Annual evapotranspiration has been estimated by the Illinois State Water Survey: *Jones* [1966, p. 12] estimated the annual average evapotranspiration varying from 635 mm (25 inches) in northern Illinois to 762 mm (30 inches) in southern Illinois, based on the average precipitation and the average runoff from 43 stream gauging stations in Illinois. However, this paper focuses on estimation of monthly evaporation in Illinois from 1983 to 1994 over a regional scale ($\approx 2 \times 10^5$ km^2). The objective here is to investigate the validity of the AWB approach by comparing its evaporation estimates to those obtained by the SWB approach at a regional scale. Since the scale of interest in this study is comparable to the typical size of a GCM grid cell ($\approx 100 \times 100$ km^2–500×500 km^2), it is believed that if the two independent approaches yield similar estimates of regional evaporation, then future applications of the AWB approach to compute regional evaporation can aid in the validation of current land-surface schemes used in climate models.

The total area examined in this study is shown in Figure 1. The entire study area encompasses four grid cells with a size of $2.5° \times 2.5°$ (87.5°W–92.5°W, 37.5°N–42.5°N), which covers an area of $\sim$240,000 km^2. As shown in Figure 1, almost all of Illinois, except for the southern tip of the state, is covered by the study region along with some portions of surrounding states. The study region was chosen in an effort to strike a balance between using a larger area, which would provide more accurate atmospheric water vapor flux convergence estimates, and limiting the region to Illinois and its immediate area, so that the evaporation estimates from the AWB approach could be compared to those calculated from the SWB approach. *Rasmusson* [1968, 1971] demonstrated how using smaller areas decreases the accuracy of the estimates of water vapor flux convergence. He also argued that the small-scale spatial variations of atmospheric water vapor convergence are significant and can contribute additional inaccuracy when performing a water balance over smaller regions using spatially limited data. Specifically, *Rasmusson* [1968] indicated that the AWB approach is applicable for regions with area $>2 \times 10^6$ km^2. The application over smaller areas are less reliable especially when the size of the area is reduced to $<10^5$ km^2 [*Rasmusson*, 1971]. This lower limit of scale, for the AWB approach to be possibly valid, corresponds to the scale of interest in this study. If the AWB approach can be shown at this scale (i.e., 10^5–10^6 km^2) to estimate the regional-scale evaporation accurately, then those estimates have the potential for providing "observations" for the validation of GCM simulations.

This paper is organized as follows: Section 2 describes the data sets used in this study. Section 3 introduces the theory of the water balance approach to the estimation of regional evaporation. Results and comparisons of the evaporation estimates made using the two approaches are presented in sections 4 and 5, respectively. Section 6 investigates the sensitivity of the re-

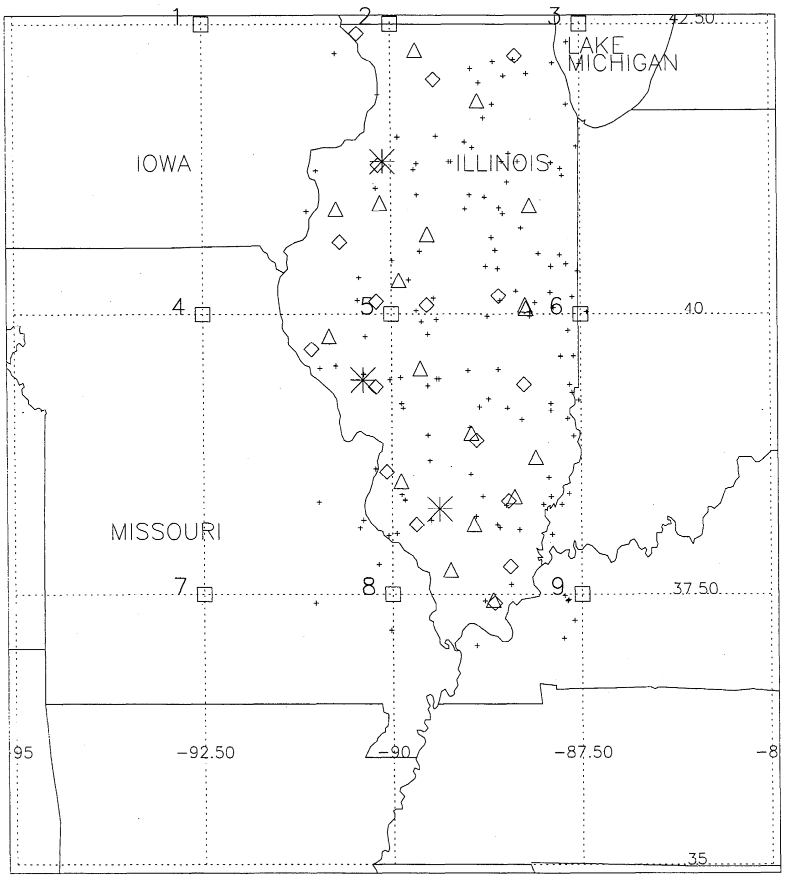

Figure 1. Location of the measurement sites of hydrological variables: precipitation (pluses), soil moisture (triangles), groundwater level (diamonds), streamflow (stars). The nine squares at the boundaries are the 2.5° × 2.5° grid points of National Center for Environmental Protection/National Center for Atmospheric Research (NCEP/NCAR) specific humidity and wind velocity data. The study region is encompassed by these nine grid points with an area of ~240,000 km^2.

gional evaporation estimates from the atmospheric water balance computations to the domain size. The sensitivity of specified parameters to the evaporation estimates from the soil water balance computations is also presented in this section. The conclusions are given in section 7.

2. Data

The data used in this study consist of the following variables: atmospheric specific humidity, wind velocity, precipitation, streamflow, snow depth, soil moisture content, and groundwater level. The locations of all various measurement stations are shown in Figure 1. A brief summary of the data used in this study is given in Table 1.

The data on specific humidity and wind speed is a subset of the National Center for Environmental Prediction/National Center for Atmospheric Research (NCEP/NCAR) global reanalysis data [*Kalnay et al.*, 1996]. This data set has daily and 6 hour temporal resolutions and a 2.5° × 2.5° horizontal resolution. It includes the mean specific humidity at eight pressure levels and wind velocity at seventeen pressure levels in the atmosphere. In this study, we use data of 6 hour resolution at the following eight pressure levels: 1000, 925, 850, 700, 600, 500, 400, and 300 mb. Above 300 mb, the atmospheric water vapor content is negligible. The NCEP/NCAR global reanalysis data have been used by *Higgins et al.* [1996] and *Gutowski et al.* [1997] to evaluate the moisture budget in the Central United States.

The data on precipitation have been supplied by two sources: Midwest Climate Center (MCC) [*Kunkel et al.*, 1990] and EarthInfo. Inc. The MCC data set consists of daily precipitation at 129 NWS cooperative observe weather stations within Illinois. The EarthInfo. data set consists of hourly precipitation at 52 stations within the same areas. Area average precipitation observations from 181 stations were used in this study.

Table 1. A Summary of the Data Sets Used in This Study

Data	Number of Stations	Resolution	Sources
Precipitation	129	daily	MCC
	52	hourly	EarthInfo. Inc.
Wind/Humidity	9 grids (global 2.5° × 2.5°)	6 hours	NCEP/NCAR
Snow	129	daily	MCC
Soil moisture	16	1–4 times/month	ISWS
Ground water	15	monthly	ISWS
Streamflow	3	daily	USGS

MCC, Midwest Climate Center; ISWS, Illinois State Water Survey; NCEP, National Center for Environmental Prediction; and NCAR, National Center for Atmospheric Research.

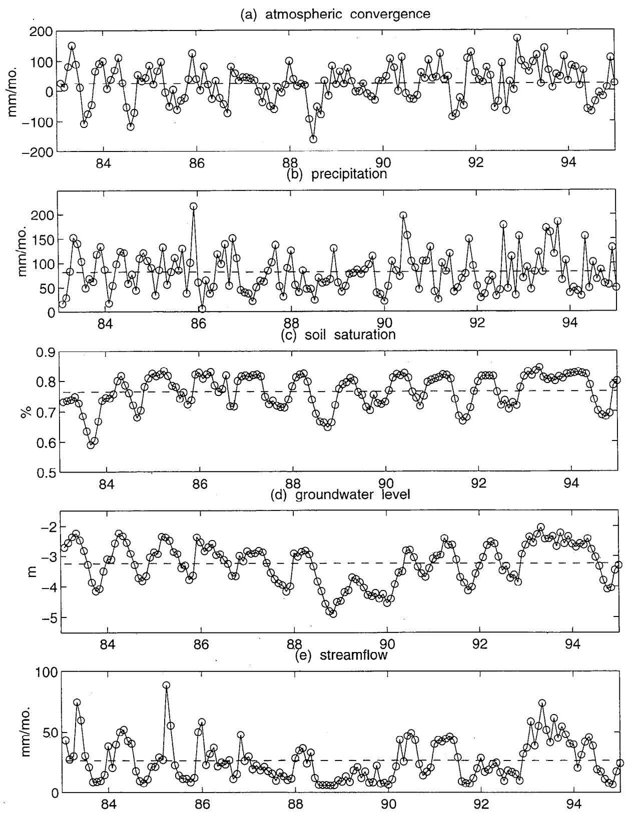

Figure 2. Twelve year time series of (a) atmospheric water vapor convergence, (b) precipitation, (c) soil relative saturation, (d) groundwater level, and (e) streamflow in Illinois. The dash lines denote the long-term average from 1983 to 1994.

The snow data were also provided by MCC. The 129 stations of MCC that reported daily precipitation reported daily records of snowfall as well. The snow data included both records of the snowfall and snow depth.

The daily streamflow records collected by the U.S. Geological Survey (USGS) were used in this study. Daily discharge measurements at three hydrological stations were used in this study: Illinois River at Valley city, which drains ~69,237 km^2; Rock River near Joslin, which drains ~24,721 km^2; and Kaskaskia River near Venedy Station, which drains ~11,373 km^2. These three streamflow-gauging stations were selected as downstream as possible with the largest drainage areas along the three major rivers in Illinois. The station at Valley City represents the outflow from the total Illinois River Basin, consisting of the upper and lower Illinois River Basins, which encompass 28,032 km^2 and 45,978 km^2 of central and western Illinois, respectively. Rock and Kaskaskia River Basins represent northern and southern Illinois, respectively. Their integrated monthly discharges were weighted by drainage areas to give an estimate of average streamflow in Illinois. The total drainage area of these three rivers exceeds two thirds of the total area of Illinois.

The data on soil moisture were collected by the Illinois State Water Survey (ISWS). Since 1981, ISWS has been collecting measurements on the soil moisture content at 8 grass-covered sites around Illinois using the neutron probe technology. Seven additional sites were added in 1982; two more were added in 1986; and by 1992 the total increased to 19. Sixteen of these 19 sites cover the period 1983–1994 and hence will be used in this study. Biweekly (March through October) and monthly (November through February) measurements of soil wetness were taken at 11 different soil layers with a resolution of about 20 cm down to 2 m below the surface. The data on soil porosity, field capacity, and permanent wilting point were also provided in this data set which enables us to estimate the water-holding capacity of soil layers. The details about this extensive soil moisture data set has been published elsewhere [*Hollinger and Isard*, 1994]. *Findell and Eltahir* [1997] used this data set to investigate the soil moisture-rainfall feedback mechanism in Illinois.

The data set on groundwater level was provided by the Groundwater Division of ISWS, which consists of monthly measurements of shallow groundwater level at 18 wells scattered throughout the state of Illinois from the 1960s till now. These 18 wells are located far away from pumping centers and streams. All of these 18 wells are under unconfined conditions, where the average water table levels range between 1 to 10 m below the surface. Fifteen wells with complete records from 1983 to 1994 were used in this study to represent the regional groundwater system in Illinois. This data set has been used to investigate the 1980–1981 drought in southern Illinois by *Changnon et al.* [1982] and the statistical relationships between precipitation and groundwater level in Illinois by *Changnon et al.* [1988]. Figure 2 shows the 12 year time series for monthly spatial averages of atmospheric water vapor convergence, precipitation, soil relative saturation, groundwater level, and streamflow.

3. Theoretical Background

3.1. Soil Water Balance (SWB)

The large-scale water balance equation for the unsaturated soils can be written as follows:

$$\overline{nD}\,\frac{\partial \bar{s}}{\partial t} = \bar{P} - \bar{E} - \overline{R_S} - \overline{P_G} \tag{1}$$

where $\overline{nD}$ (millimeters) is the available storage depth of the soil: the product of soil porosity and root zone depth. (Throughout this paper the overbar denotes the regional-scale monthly mean value.). Here $\bar{s}$ (percent) is the soil relative saturation (i.e., soil moisture content divided by soil porosity), $\bar{P}$ (mm/month) is the precipitation, $\bar{E}$ (mm/month) is the evaporation, $\overline{R_S}$ (mm/month) is the surface runoff, and $\overline{P_G}$ (mm/month) is the downward percolation flux into the groundwater aquifer. Illinois has a rather uniform terrain with the local relief in most counties <60 m (200 feet) [*Leighton et al.*, 1948], hence the interflow is not a significant mechanism of runoff generation. The regional-scale evaporation can be estimated if all the other terms in (1) are known,

$$\bar{E} = \bar{P} - \overline{nD}\,\frac{\partial \bar{s}}{\partial t} - \overline{R_S} - \overline{P_G} \tag{2}$$

However, usually $\overline{R_S}$ and $\overline{P_G}$ are not measured. The water balance equation for the unconfined groundwater aquifer is given by

$$\overline{S_y}\,\frac{\partial \bar{H}}{\partial t} = \overline{P_G} - \overline{R_G} \tag{3}$$

where $\overline{S_y}$ (percent) is the specific yield, $\bar{H}$ (millimeter) is the groundwater level, and $\overline{R_G}$ (mm/month) is the groundwater discharge. Here $\overline{S_y}$ is the fraction of water volume that can be drained by gravity in an unconfined aquifer [*Domenico and Schwartz*, 1990].

Since the direct measurement of percolation flux and surface runoff are difficult to obtain, the estimation of regional-scale percolation and the separation of streamflow into surface runoff and groundwater runoff have been a difficult task. However, this difficulty can be avoided by adding (1) and (3) together to get

$$\overline{nD}\,\frac{\partial \bar{s}}{\partial t} + \overline{S_y}\,\frac{\partial \bar{H}}{\partial t} = \bar{P} - \bar{E} - \bar{R} \qquad (4)$$

where $\bar{R} = \overline{R_S} + \overline{R_G}$ (mm/month) is the total (measured) runoff. This equation describes the soil water balance in a drainage basin or a physiographic unit, where there is no water movement between the adjacent hydrologic units. From (4), regional evaporation can be estimated as

$$\bar{E} = \bar{P} - \left(\bar{R} + \overline{nD}\,\frac{\partial \bar{s}}{\partial t} + \overline{S_y}\,\frac{\partial \bar{H}}{\partial t}\right) \qquad (5)$$

which facilitates the soil water balance computations by avoiding the necessity of hydrograph separation and percolation estimation.

For regions where the snow accumulation and ablation are important hydrological quantities, the combination of unsaturated and saturated water balances in (1) and (3) into (4) are not sufficient to accurately describe water balance. The snow accumulation and ablation can be incorporated into water balance computations by adding another equation representing the surface water storage as follows:

$$\frac{\partial \overline{W_S}}{\partial t} = \bar{P} - \bar{I} - \overline{V_S} - \overline{V_M} \qquad (6)$$

where $\overline{W_S}$ (millimeter) is the accumulated depth of snowpack (liquid equivalent), $\bar{I}$ (mm/month) is infiltration rate, $\overline{V_S}$ (mm/month) is the sublimation rate, and $\overline{V_M}$ (mm/month) is snow melting rate. Under such a condition, (1) should be modified into the following:

$$\overline{nD}\,\frac{\partial \bar{s}}{\partial t} = \bar{I} + \overline{V_M} - \bar{E} - \overline{R_S} - \overline{P_G} \qquad (7)$$

Therefore the equations for the water balance of the surface and the subsurface can be combined (by adding (3), (6), and (7)) and rearranging to result in

$$\bar{E} + \overline{V_S} = \bar{P} - \left(\bar{R} + \overline{nD}\,\frac{\partial \bar{s}}{\partial t} + \overline{S_y}\,\frac{\partial \bar{H}}{\partial t} + \frac{\partial \overline{W_S}}{\partial t}\right) \qquad (8)$$

where the left-hand side denotes the combination of the rates of evaporation and snow sublimation.

Figure 3 illustrates the 12-year spatial average seasonal variability of snow accumulation depth $\overline{W_S}$, soil saturation $\bar{s}$, and groundwater level $\bar{H}$ in Illinois. These hydrological quantities correspond to the water storage in three distinct reservoirs: unsaturated soil reservoir, aquifer reservoir, and surface reservoir. For the two subsurface reservoirs, a similarity in the seasonal cycle is clearly demonstrated in Figure 3 with a peak of surface water storage occurring in March and a trough in August or September. The snow depth shown in Figure 3 is the 129 station average snow accumulation depth on the last day of each month. Figure 4 shows the average days within a month when there are snow accumulations and the seasonal cycle of snowfall in Illinois. The snow data shown in Figures 3 and 4 have been converted into the water equivalent by multiplying 0.1. The long-term (1983–1994) average annual snowfall is ~115 cm/yr, which is less than the average annual snowfall for North America 35°N × 45°N (i.e., 140 cm/yr) [see *Groissman and Legates*, Table 3, 1994]. Actually, Illinois is not a snowy state even by Midwestern standards (J. Angel, climatologist, ISWS, personal communication, 1997). The major reason is that Illinois locates at the upwind direction of Lake Michigan such that the lake-effect snow is irrelevant. Although the snow amount comprises a significant fraction of winter precipitation in Illinois, close examinations of the snow data reveal that after a day with snow occurrence the snow accumulation lasted only through the subsequent 1–5 days for most of the cases. Therefore we inferred the snow storage effect in Illinois is insignificant in comparison with other hydrological quantities involved

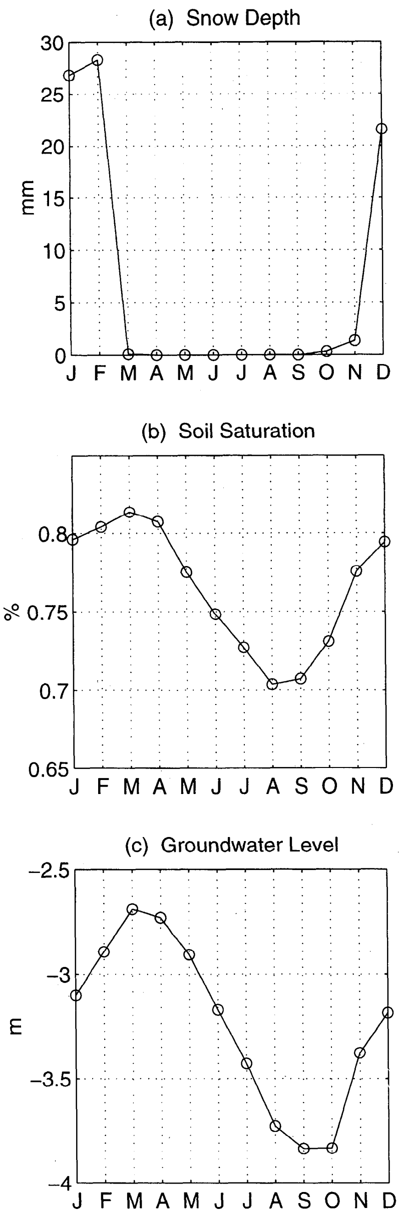

Figure 3. Twelve year average seasonal cycles of (a) snow depth, (b) soil saturation, and (c) groundwater level in Illinois from 1983 to 1994.

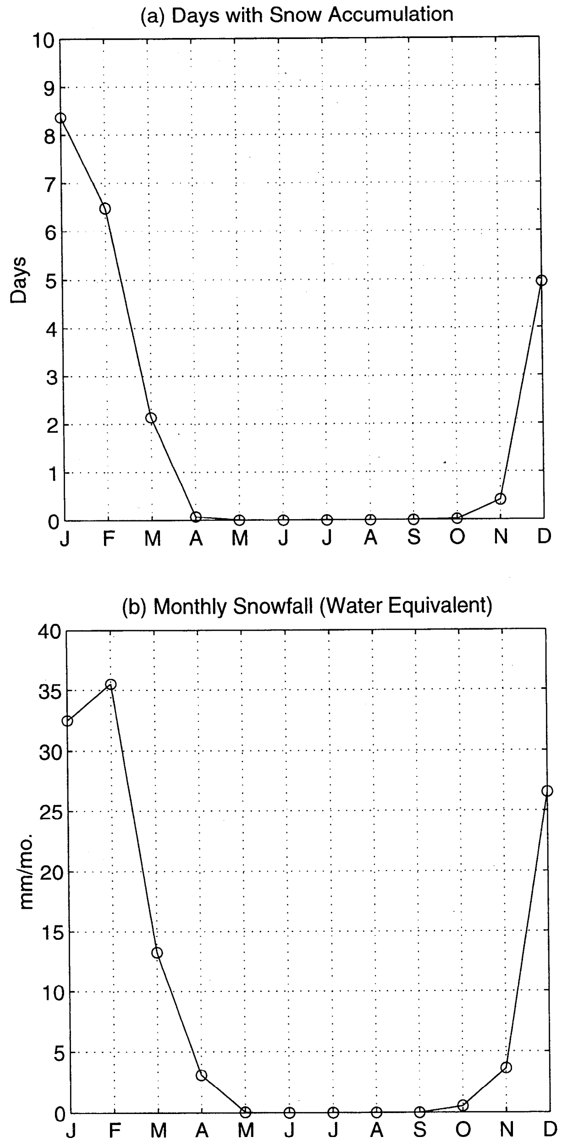

Figure 4. (a) The average days within a month with snow accumulations in Illinois; (b) seasonal cycle of snowfall in Illinois from 1983 to 1994.

in monthly water balance computations. Given the limited role of snow in Illinois, we will employ (5) in conjunction with observations of $\bar{P}$, $\bar{R}$, $\bar{n}$, $\bar{s}$, and $\bar{H}$ to estimate the regional evaporation $\bar{E}$ in Illinois during the period 1983–1994.

Before the implementation of evaporation estimation procedure, two soil parameters in (5), the root-zone depth $\bar{D}$ and specific yield $\overline{S_y}$, should be determined. Constant values representing the regional-scale average soil properties are estimated and used in this study since the paucity of data excludes the possibility to characterize the spatial patterns of these parameters.

Root-zone depth, from which the soil water is available for evapotranspiration, can extremely vary within the range of a few centimeters to a few meters according to the different soil textures and vegetation characteristics. The determination of mean root-zone depth was initially performed by calculating the change in soil water storage for all of the available surface layers. It was found that the water stored below 1 m soil layers has much less variations than those stored within top 0.5 m soil layers. However, since the data are available, the mean root-zone depth was taken as 2 m in this study in order to avoid underestimation of the available storage volume of soil.

The mean specific yield $\overline{S_y}$ in (5) is also considered as a constant. Several sources of information can be utilized to infer the representative value of $\overline{S_y}$. The first source is the study of hydrological budgets from three small watersheds in Illinois by *Schicht and Walton* [1961]. Their analysis of groundwater recessions of different duration yielded a steady value of $\overline{S_y}$ = 0.08. Next, judging from the soil logs taken from 19 access holes of neutron probe in Illinois [*Hollinger and Isard*, 1994, Table 2], the soil textures at the soil moisture measurement sites are silt loam with only two exceptions: One of nineteen stations is silt clay loam, and the other one is loamy sand. Further, according to the soil map provided by *Food and Agriculture Organization (FAO)-UNESCO* [1975], the U.S. Department of Agriculture (USDA) soil texture class in Illinois is "fine silt" with the drainage characteristic as "imperfect, moderate to moderate slow permeability." According to the compilation of typical values of specific yield by *Johnson* [1967; also see *Domenico and Schwartz*, 1990, p. 69] for various soils, the representative specific yield in the unconfined silt loam aquifers in Illinois is specified as 0.08 in this study, which is exactly the value found by *Schicht and Walton* [1961]. Therefore $\overline{S_y}$ = 0.08 was used in this study. Scientists from ISWS (A. Visocky, ISWS, personal communication, 1997) have suggested that $\overline{S_y} \cong 0.08 - 0.1$ is an appropriate estimate for the unconfined aquifers in Illinois. Sensitivity analysis was also conducted that indicates the variation within a possible range of $\overline{S_y}$ (0.05–0.10) has negligible effect on the changes in monthly evaporation estimates, as discussed later in section 6.

3.2. Atmospheric Water Balance (AWB)

The atmospheric water balance equation can be written as [*Peixoto and Oort*, 1992]

$$\frac{\partial \overline{W_a}}{\partial t} = \bar{E} - \bar{P} + \bar{C} \tag{9}$$

where $\overline{W_a}$ is the mean precipitable water (or water vapor storage) within the atmospheric column, $\bar{C}$ (equals $-\nabla \cdot \bar{Q}$) is the mean convergence of lateral atmospheric vapor flux. Here $\bar{Q}$ is the vertically integrated mean total water vapor flux. Values $\overline{W_a}$ and $\bar{Q}$ can be calculated by integrating the profiles of specific humidity q, zonal and meridional wind components u and v, from the pressure at the ground surface p_s to that above which water vapor content becomes negligible p_u (i.e., p_u = 300 mbar in this study) as follows [*Rasmusson*, 1967, 1968; *Peixoto and Oort*, 1992]:

$$\overline{W_a} = \frac{1}{g}\int_{p_u}^{ps} \bar{q}\, dp \quad \overline{Q_\lambda} = \frac{1}{g}\int_{p_u}^{ps} \overline{qu}\, dp \quad \overline{Q_\phi} = \frac{1}{g}\int_{p_u}^{ps} \overline{qv}\, dp \tag{10}$$

where g is the acceleration of gravity and λ and ϕ represent the longitude and latitude, respectively. The most convenient way of calculating the mean atmospheric vapor flux convergence $\bar{C}$ is through the application of the Gauss Theorem:

$$\nabla \cdot \bar{Q} = \frac{1}{A}\oint_S \bar{Q} \times n ds \tag{11}$$

Table 2. Average Monthly Soil Water Balance and Atmospheric Water Balance Components in Illinois From 1983 to 1994

	P, mm	E_{soil}, mm	E_{soil}/P, %	R, mm	R/P, %	s, %	H, m	$nD\ \partial s/\partial t$, mm	$S_y\ \partial H/\partial t$, mm	E_{atmo}, mm	E_{atmo}/P, %	C, mm	$\partial W_a/\partial t$, mm
Jan.	39.2	1.50	3.8	27.7	70.7	79.6	−3.10	7.2	2.9	0.4	1.0	41.4	2.6
Feb.	50.2	−2.5	−5.0	28.7	57.2	80.4	−2.89	7.5	16.5	−0.4	−0.9	50.2	−0.5
March	77.1	12.3	15.9	39.5	51.2	81.4	−2.69	8.9	16.4	12.6	16.3	66.6	2.0
April	90.0	59.0	65.6	41.2	45.8	80.8	−2.73	−6.9	−3.3	51.7	57.4	45.8	7.5
May	94.8	108.7	114.7	32.3	34.1	77.5	−2.91	−31.9	−14.3	96.9	102.2	2.8	4.9
June	89.3	112.0	125.4	24.6	27.5	74.9	−3.17	−26.2	−21.0	122.5	137.2	−22.4	10.8
July	97.3	117.5	120.8	20.9	21.5	72.7	−3.43	−20.6	−20.5	126.9	130.4	−32.0	−2.4
Aug.	83.5	114.6	137.3	14.8	17.7	70.4	−3.73	−21.9	−23.9	120.4	144.3	−42.4	−5.5
Sept.	89.0	80.5	90.5	15.2	17.1	70.7	−3.83	1.8	−8.6	74.1	83.3	7.9	−7.0
Oct.	83.0	42.5	51.2	16.6	20.0	73.1	−3.83	23.8	0.1	50.6	61.0	31.1	1.3
Nov.	112.5	9.9	8.8	22.7	20.2	77.6	−3.38	43.5	36.4	21.0	18.7	82.9	−8.6
Dec.	69.0	3.8	5.5	30.0	43.5	79.4	−3.19	19.8	15.3	5.5	8.0	62.1	−1.3
Average	81.2	55.0	61.2	26.2	32.2	76.5	−3.24	0.42	−0.33	56.9	63.2	24.5	0.09
Total	974.8	659.8	—	314.1	—	—	—	4.98	−4.01	682.2	—	293.7	1.07

where A is the corresponding area over which the convergence is being calculated. Hence $\bar{C}$ can be calculated by taking the line integral of the water vapor flux around the area under study.

By using the aerological data to calculate the convergence $\bar{C}$ and change in atmospheric water vapor storage with respect to time $\partial \overline{W_a}/\partial t$, (9) can be solved for the unknown regional evaporation:

$$\bar{E} = \bar{P} - \bar{C} + \frac{\partial \overline{W_a}}{\partial t} \tag{12}$$

The mean precipitable water and mean zonal and meridional water vapor fluxes are calculated at each of the nine grid points shown in Figure 1 by integrating vertically with respect to pressure using the trapezoidal method. Once these vertically integrated fluxes are determined, the convergence can be calculated for a given area by (11). The approach used here was essentially identical to that used by *Rasmusson* [1968, 1971].

By averaging (5) and (12) over long time series, all the derivative terms can be assumed negligible. Thus, by equating the two estimates of long-term evaporation, we derive

$$\bar{R} = \bar{C} = -\nabla \times \bar{Q} \tag{13}$$

which is an expression of the fact that for any climate equilibrium the long-term convergence of atmospheric moisture toward any hydrologic unit has to be balanced by the long-term net discharge of water out of the same hydrologic unit. Therefore (13) can be conceived as a criterion for evaluating the agreement between the atmospheric and hydrological data sets.

Equations (5) and (12) constitute the basis of the mass balance methods for the estimation of regional evaporation [*Brutsaert*, 1982]. These two equations are valid for all scales. However, the accuracy by which each of the terms on the right-hand sides of (5) and (12) can be evaluated varies depending on the spatial and temporal resolutions of the data used. Particularly, it should be noted that precipitation measurements from rain gauges tend to have a systematic negative bias, in that they underestimate the actual precipitation occurring over an area due to the local wind effects around the gauge. This can be especially pronounced during the winter months when snow occurs, because the snowflakes are much more prone to wind deflection than raindrops. According to *Larson and Peck* [1974], rain gauges can underestimate rainfall up to 20% for a wind of 32.18 km/hour. For snowfall this value is likely go up by at least another 10%. *Groisman and Legates* [1994] estimated the average gauge undercatch bias of precipitation measurements is 9% of the annual precipitation for the continental United States, with a seasonal maximum of 15% in winter and a minimum of 5% in summer. The negative bias of precipitation measurement may result in the underestimation of evaporation, especially during the winter. However, inspection of (5) and (12) reveals that the bias affects both evaporation estimates from the SWB and AWB to an equal extent.

The AWB approach has been applied by several previous studies to estimate evaporation for large areas [*Benton et al.*, 1950; *Rasmusson*, 1967, 1968, 1971]. However, the SWB approach has not been applied at such a large scale. In this study, we utilize the extensive hydrological data set of Illinois to estimate regional evaporation using the SWB (5) and the AWB (12). Thus we will be able to compare the two estimates of regional evaporation in Illinois. The results are presented in section 4.

4. Estimation of Evaporation

In this section we present the estimates of monthly evaporation derived from the SWB and AWB computations. Table 2 summarizes the 12 year average annual soil and atmospheric water balances. Annual evaporation is about 660 mm, which falls within the range of the annual evaporation estimate (i.e., 635–762 mm/yr) by *Jones* [1966]. Annual total runoff is about 314 mm, which comprises about 32% of the annual precipitation (975 mm). The annual convergence of atmospheric moisture into this region is about 294 mm (Table 2). The difference between runoff and convergence is about 6.3% of runoff, which is comparable to the accuracy of the estimates themselves. This suggests that any additional lateral and downward fluxes (leakage) across the boundary of the study region are indeed negligible (see (12)).

Figure 5a illustrates the mean seasonal cycle of each soil water balance component (precipitation, evaporation, runoff, and storage changes in the unsaturated and saturated zones). The 12 year average annual evaporation and runoff are ~70% and 30% of the annual precipitation (Table 2). For the evaporation estimates, a seasonal cycle is clearly demonstrated with a peak close to 120 mm/month during the summer (June–August) and a trough of almost zero during the winter (December–February). From May to August, the average monthly

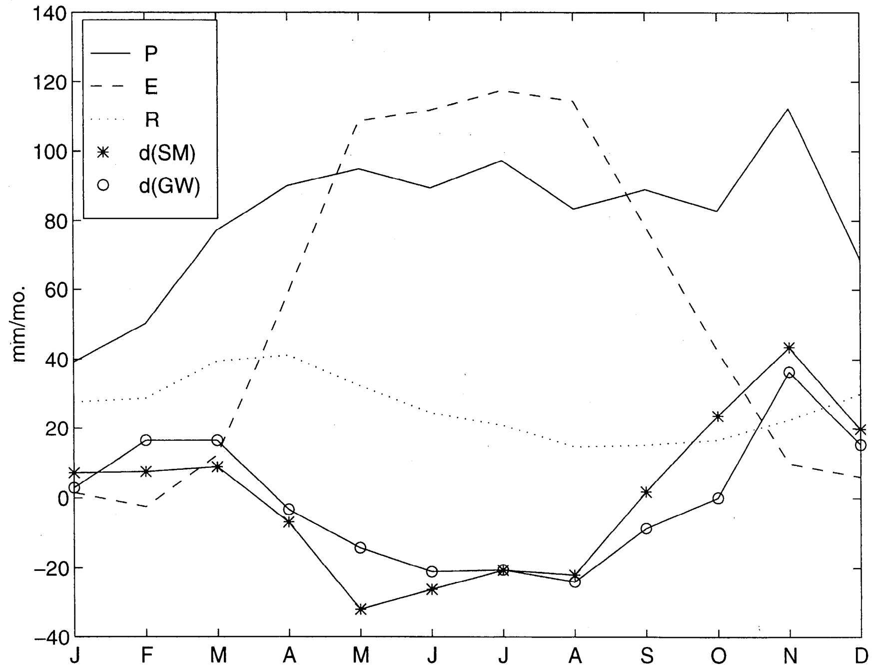

Figure 5a. Seasonal cycle of the soil water balance components in Illinois. Here P is precipitation, E is evaporation estimate from the soil water balance, R is streamflow, d(SM) is change in the soil moisture storage (i.e., $nD\ ds/dt$), and d(GW) is change in the groundwater storage (i.e., $S_y\ dH/dt$).

evaporation exceeds that of precipitation, while during the winter months, evaporation falls well below the average precipitation. Upon examination of each individual year, some evaporation estimates during the late fall and winter months appear to be below zero because of their limited accuracy.

The contributions of changes in both unsaturated and saturated storage to the annual water balance are small (i.e., 0.4–0.5%, see Table 2). However, from month to month these storage changes can be significant because of their apparent seasonal cycles (Figure 5a). For some months, the sum of the changes in soil moisture storage and in groundwater storage amount to almost 70% of the corresponding monthly precipitation. Considering the unsaturated and saturated zone as a whole, the seasonal cycle of the change in monthly subsurface

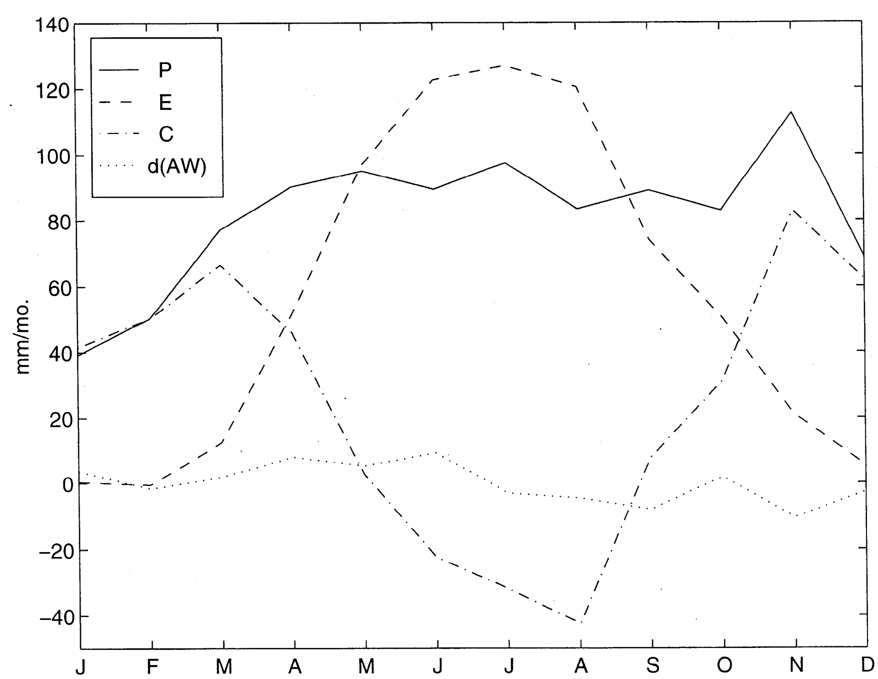

Figure 5b. Seasonal cycle of the atmospheric water balance components in Illinois. Here P is precipitation, C is atmospheric water vapor convergence, E is evaporation estimate from the atmospheric water balance, and d(AW) is change in the atmospheric water storage (i.e., dW_a/dt).

storage has a maximum of 80 mm/month in November and a minimum of −47 mm/month in June, which constitutes a large contribution to the monthly water budget. Rainfall is uniform throughout the year, although conditions during the spring, summer, and fall are relatively wet in comparison with the relatively dry winter conditions (Figure 5a). The strong seasonal cycle of incoming solar radiation (not shown here) forces a similar cycle in evaporation, which then propagates through the soil down to the groundwater aquifer to dictate similar patterns of seasonal variability in soil moisture and groundwater levels. The seasonal cycle of evaporation is reflected in the changes of monthly unsaturated and saturated storage. Because of the same order of magnitude of both subsurface storage changes, the incorporation of the change in groundwater storage is indispensable for monthly soil water balance computation in humid areas such as Illinois, since fluctuations in the shallow groundwater table level (i.e., 2–5 m on average) cause a significant variation in saturated storage at the monthly timescale.

The evaporation estimates from the AWB approach are presented as follows: On average, the total evaporation E_{atmo} over a year is 682 mm, and the net convergence of water vapor is 294 mm (Table 2). Again, the annual evaporation estimate is close to that estimated by *Jones* [1966]. As expected, the change in atmospheric storage $\partial \overline{W}_a/\partial t$ is small (usually <10 mm/month) and integrates to zero over the annual timescale (Table 2). The mean monthly values over the 12 year record of evaporation E_{atmo}, along with precipitation P, convergence C, and change in storage $\partial \overline{W}_a/\partial t$, are shown in Figure 5b. A seasonal cycle for the evaporation estimates with a peak more than 120 mm/month during the summer (June–August) is clearly seen in Figure 5b. Evaporation is lowest in the winter months with a trough almost equal to zero (December–February). Similar to the SWB computation, during the months of May, June, July, and August the evaporation rate exceeds the precipitation rate. Atmospheric moisture diverges away from this region, and the subsurface storage of water acts as a significant source of water to the atmosphere. However, these patterns are reversed during the remaining months of the year: Precipitation exceeds evaporation, significant amounts of atmospheric moisture converges toward the region, and this helps to replenish the subsurface storage of water before the onset of dry conditions in the following summer.

In February, the AWB approach gives an average evaporation equal to −0.45 mm/month. Although small, this is the only negative value among the twelve monthly evaporation estimates. Interestingly, February is also the only month when the SWB approach yields a negative average evaporation estimate (Table 2). Upon examination of the individual years, some values of E_{atmo} during the late fall and winter are negative. In most cases the negative values are within 0–−10 mm/month. However, in an extreme case such as November 1991 the evaporation estimate E_{atmo} is of the order of −49 mm/month. In his study *Rasmusson* [1971] also found negative evaporation estimates during the fall and winter months. He argued that the error is most likely introduced in the measurement of precipitation and computation of convergence. As discussed earlier (section 3.2), the negative bias of precipitation may be the primary source of error for the underestimation of winter evaporation, with the remaining error most likely resulting from the estimation of convergence. The latter has been discussed in detail by *Rasmusson* [1977].

The impact of the undersampling of winter precipitation on the evaporation estimates can be roughly evaluated by increasing winter precipitation by 15%, according to the study by *Groisman and Legates* [1994]. The resulting average evaporation estimates in winter months increased from 3.8 mm (December), 1.5 mm (January), and −2.5 mm (February) to 16.4 mm, 7.4 mm, and 5.0 mm, respectively.

It is important to look at the AWB equation and note how each component affects the estimate of evaporation. From (12) it can be seen that precipitation $\bar{P}$ and changes in atmospheric storage $\partial \overline{W}_a/\partial t$ affect evaporation $\bar{E}$ positively; that is, an increase in $\bar{P}$ or $\partial \overline{W}_a/\partial t$ will increase $\bar{E}$. Conversely, $\bar{E}$ is negatively related to convergence $\bar{C}$. Therefore, when $\bar{C}$ becomes more positive (negative), $\bar{E}$ decreases (increases). As mentioned earlier, changes in atmospheric storage are small and therefore have little impact on evaporation. While precipitation is slightly lower in the winter, evaporation is expected to be at its lowest because of limited solar radiation, which is consistent with the positive convergence throughout the fall and winter months. During the summer months, precipitation is abundant, and evaporation is expected to be highest due to the seasonal pattern of incoming solar energy. This is consistent with the strong negative peak in convergence for almost every summer. It is interesting to note that, while the seasonal variability of the evaporation estimates based on the SWB approach is largely balanced by the seasonal pattern of subsurface storage rather than by the lateral water fluxes (runoff), the reverse is true for the evaporation estimates based on the AWB approach. The seasonal pattern of evaporation estimates from the AWB approach is almost entirely balanced by the seasonal pattern of lateral fluxes of water vapors and not by the changes in storage. This contrast reflects a fundamental difference in the hydrology of the land and atmospheric branches of the regional water cycle. The differences in the memories of the two systems underscore the importance of the land in creating climate persistence in atmospheric states.

5. Comparison of Evaporation Estimates From the Soil Water Balance and the Atmospheric Water Balance

Annually, the difference between the evaporation estimates from the SWB and AWB approaches is 22 mm/yr. Since changes in both the atmospheric and subsurface water storage are negligible over the annual timescale, the difference between two evaporation estimates is thus equal to the difference between the annual totals of atmospheric convergence and runoff. The comparison between the time series of two evaporation estimates, E_{soil} from (5) and E_{atmo} from (12), is shown in Figure 6 for each of 12 years. The average of the difference between the two evaporation estimates is 1.8 mm/month, and the standard deviation is 37 mm/month. The two estimates agree reasonably well in the timing and magnitude of the seasonal pattern of evaporation; however, significant differences are evident in some estimates of evaporation. Twenty-three monthly evaporation estimates during the 12 years (15% of total estimates) have differences >50 mm/month. The largest difference between E_{soil} and E_{atmo} is 108 mm/month in October 1986 when an unreasonable negative $E_{\text{soil}} = -72$ mm/month is obtained from the soil approach. Further, for June 1988, $E_{\text{soil}} = 102$ mm/month and $E_{\text{atmo}} = 197$ mm/month, which has a difference of 95 mm/month. The evaporation estimates are also compared on a monthly basis by plotting the estimates from the SWB approach versus the estimates

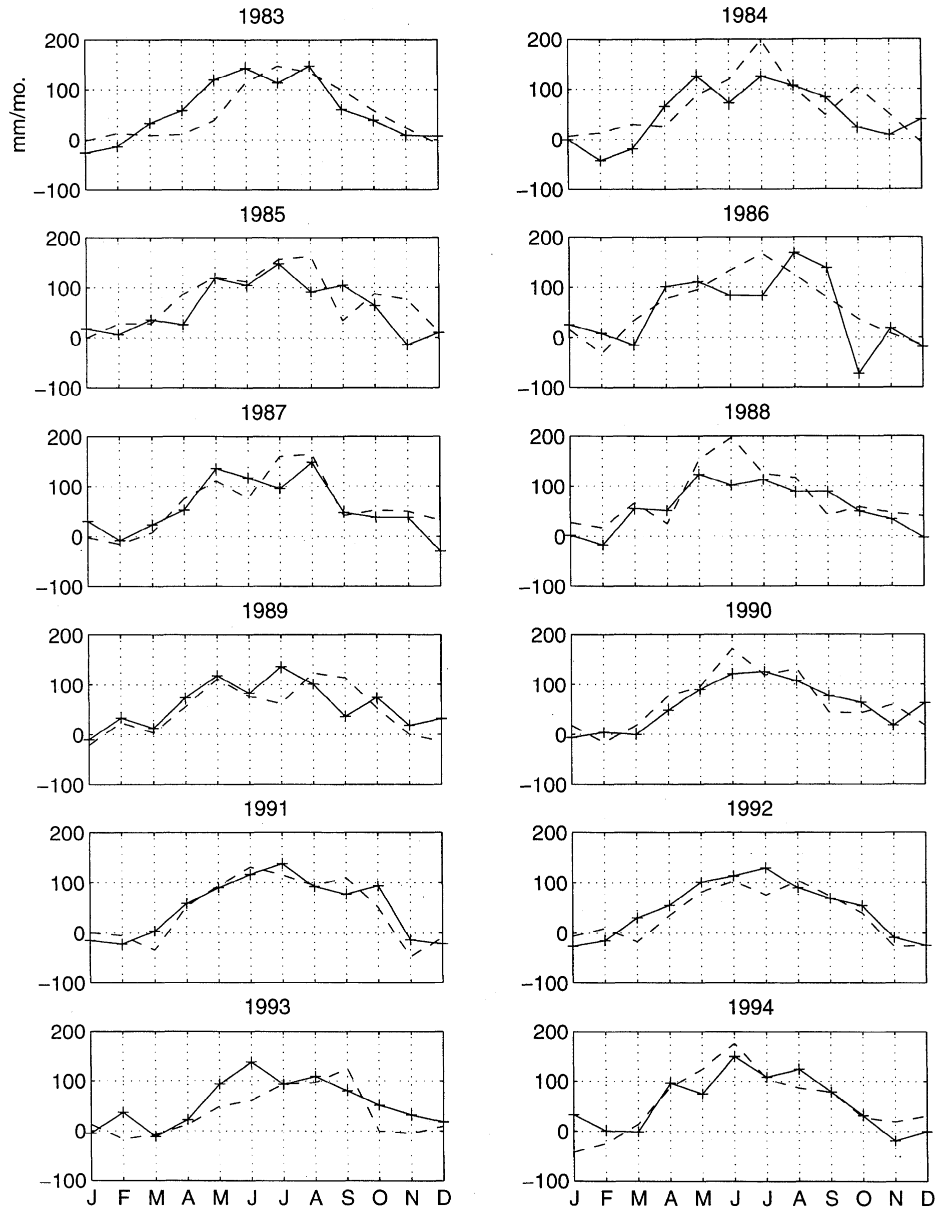

Figure 6. Twelve year time series of evaporations estimated from soil water balance approach (solid line) and atmospheric water balance approach (dash line) from 1983 to 1994.

from the AWB approach (Figure 7). For perfect estimates, the data would lie along the 45° line. The coefficient of correlation between the time series of E_{soil} and E_{atmo} is 0.785. Also shown in Figure 7 is the best fit line derived by linear regression: $E_{\mathrm{soil}} = 0.92 E_{\mathrm{atmo}}$, which reflects the annual 22 mm difference between two estimates of annual evaporation (see Table 2).

The seasonal cycles of two evaporation estimates are presented in Table 2 and plotted in Figure 8 for comparison. The standard errors of the mean evaporation are indicated by the error bars. For climatological estimates using the 12 years of data, the error is calculated as $\mathrm{STD}_i/\sqrt{12}$, where STD_i is the standard deviation of evaporation estimates in month i. Also shown in Figure 8 is the potential evaporation from the monthly Class A pan evaporation compiled by *Farnsworth and Thompson* [1982] for the United States. The evaporation measurements from Class A pans are available only from April to October. These pans are removed each fall around the middle of October and reinstalled around the middle of March. Farnsworth and Thompson suggested that potential evaporation from moist natural surfaces (i.e., shallow lake and wet soil) is roughly 70% of the Class A pan evaporation for the same meteorological conditions. Thus potential evaporation is derived by multiplying a pan coefficient of 0.7 to their pan evaporation data. From Figure 8, the two estimates E_{soil} and E_{atmo} agree reasonably well for most of the months; however, small differences of the order of 10 mm/month occur during late spring and summer. Both estimates predict peak evaporation in July of ~120 mm/month and a minimum but slightly negative evaporation in February. For both estimates the evaporation is less than the precipitation during most of the year except for summer, when evaporation exceeds precipitation (see Table 2). Both estimates predict that evaporation can

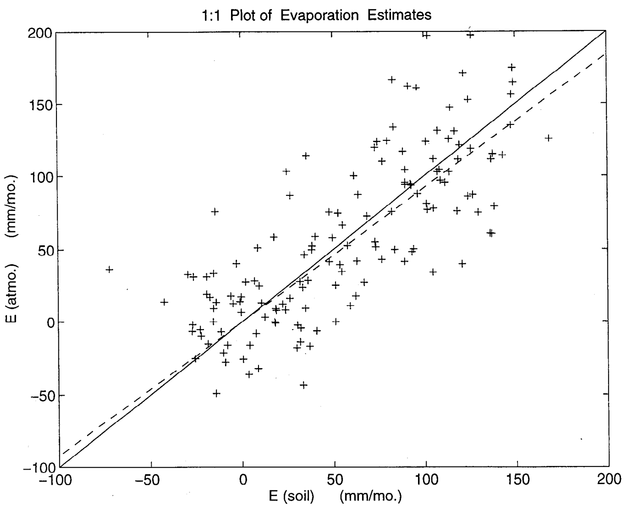

Figure 7. Plot of evaporation estimates from SWB approach versus evaporation estimates from AWB approach. The solid line is the 1:1 line; the dashed line is the best fit line by linear regression.

exceed precipitation by up to 40% in August. Over the entire year, however, evaporation is ~70% of precipitation. The seasonal evolution of both evaporation estimates follows closely to that of potential evaporation. The peak of the estimates occurred in the summer months with a magnitude of ~80–90% of potential evaporation. This is indicative of the atmospheric control on evaporation in Illinois. According to the classification of climate by *Thronthwaite* [1948], Illinois belongs to the region where there is no water deficiency throughout the year. The nonwater-stressed evaporation can be perceived in terms of the close agreement between the seasonal pattern of evaporation with that of net radiation (although not shown here).

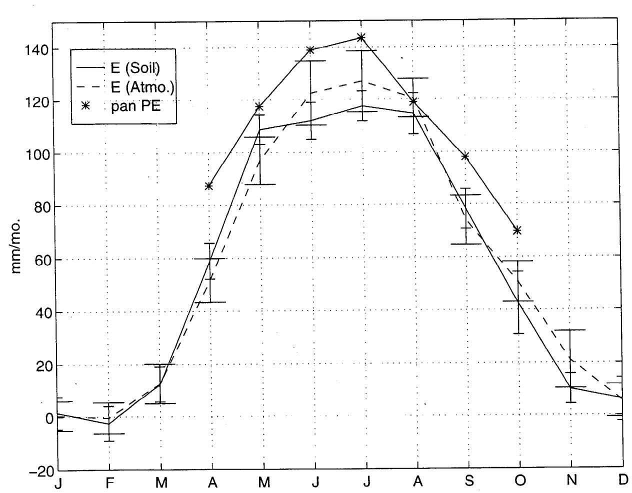

Figure 8. Seasonal cycles of evaporation estimated from SWB approach (solid line) and AWB approach (dash line), and potential evaporation (from April to October) calculated from Class A pan evaporation tabulated by *Farnsworth and Thompson* [1982]. The error bar indicates the standard error of the climatological estimates of evaporation. The wide error bars correspond to the evaporation estimates from the AWB, while the narrow ones correspond to those from SWB.

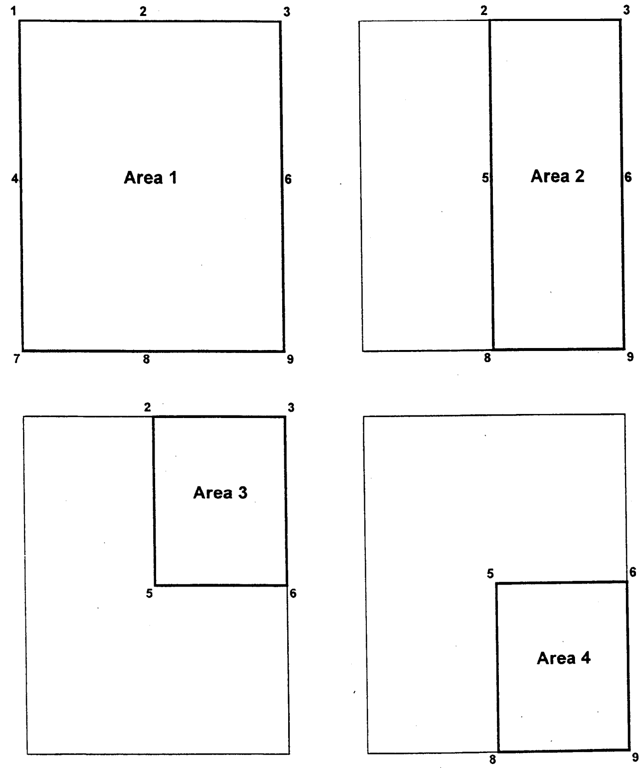

Figure 9. Schematic illustration of area 1 ($\approx 500 \times 500$ km^2), area 2 ($\approx 500 \times 250$ km^2), area 3 ($\approx 250 \times 250$ km^2), and area 4 ($\approx 250 \times 250$ km^2). The numbers marked on the boundaries of areas correspond to the nine grid points of NCEP/NCAR atmospheric data shown in Figure 1.

This implies that the variability of evaporation in Illinois is mainly controlled by the variability of energy supply.

The difference between the climatologies of E_{soil} and E_{atmo} has the mean value of 1.8 mm and the standard deviation of 7.5 mm. The evaporation estimates during the summer are expected to be closer to the true evaporation than that during the winter because of the significant rain gauge bias, particularly during the winter snowfall periods. The two approaches produce almost identical climatologies of evaporation. In light of the fact that independent data sets were used in the two approaches, this result is encouraging: The AWB approach has the potential for the accurate estimation of the climatology of regional evaporation at least for humid regions at a scale similar to that of Illinois (~10^5 km^2).

6. Sensitivity Analysis

So far, the above discussion has centered on estimates made corresponding to the whole study region ($\cong$250,000 km^2, hereafter referred to as Area 1) (see Figure 9). In order to test whether the AWB approach can be applied to even smaller regions, estimates were computed for several areas within the same study region: area 2 ($\approx 500 \times 250$ km^2), area 3 ($\approx 250 \times$ 250 km^2), and area 4 ($\approx 250 \times 250$ km^2) (see Figure 9). To estimate evaporation for these areas, the same AWB approach described above was used. The only changes involved were the precipitation data and flux calculations. For precipitation, only those gauges within each area were used. The mean precipitation over the 12 years for each area is shown in Figure 10a.

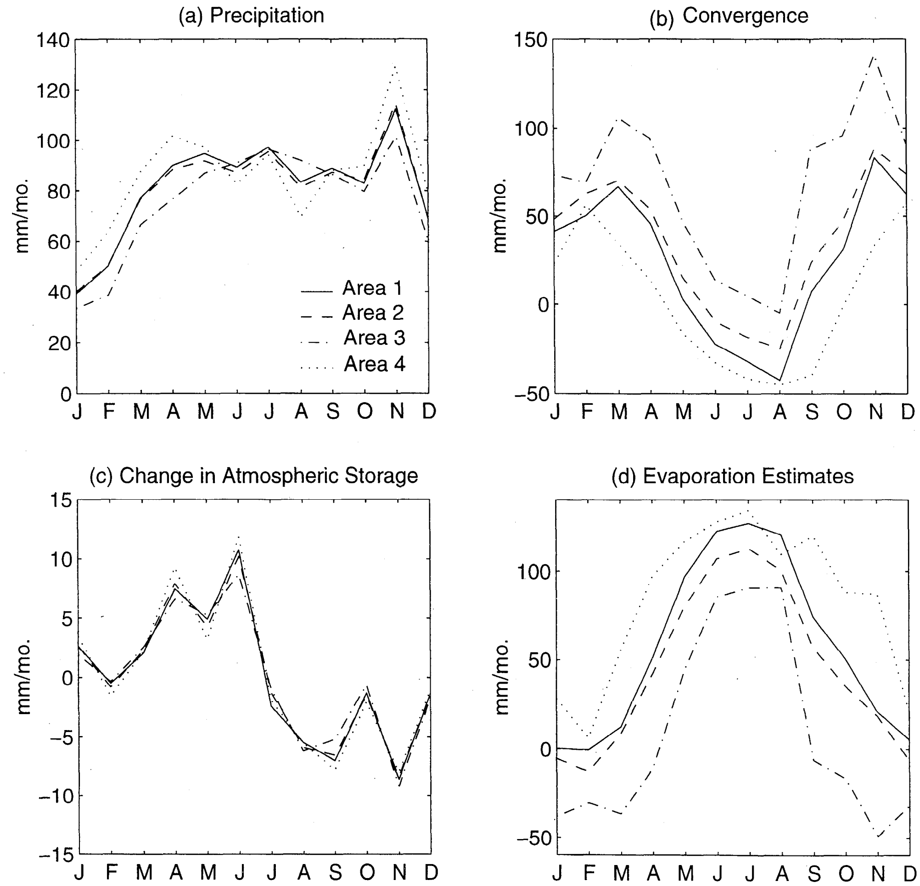

Figure 10. Twelve year monthly averages of (a) precipitation (b) atmospheric water vapor convergence, (c) change in atmospheric storage, and (d) evaporation estimated from atmospheric water balance approach for four different areas shown in Figure 8.

From Figure 10a, precipitation does not change significantly; however, there is more precipitation in area 4 and less precipitation in area 3 except in summer, while area 2 has approximately the same amount of precipitation as area 1. The fluxes which are calculated to determine the convergence also change based on which node makes up the boundary of a specific area. For areas 3 and 4 only four boundary nodes are available, so the estimates for these areas are most prone to error. The mean convergence over the 12 years for each of the four areas is shown in Figure 10b. All four of the convergence estimates exhibit a similar pattern in terms of two maximum peaks in spring and November and a large negative peak in August. For areas 1 and 2 the results are very similar, however, for areas 3 and 4 the results differ significantly. The mean change in storage $\partial \overline{W_a}/\partial t$ over 12 years shows little sensitivity to changes in area (Figure 10c). The differences in convergence between the different areas stem from two factors: the limited accuracy of land extent of the available observations and the pattern of atmospheric circulation in this region. Specifically, the calculations of $\partial \overline{W_a}/\partial t$ and convergence use the same specific humidity data; therefore it can be inferred that the humidity is not the sensitive variable to the size of area. Instead, wind velocities, the only other atmospheric variable used to calculate the fluxes (see (10)), are the ones which cause the difference in convergence as shown in Figure 10b. As mentioned earlier, *Rasmusson* [1968, 1971] stated that the AWB approach is more accurate for large areas, and this statement is indeed valid when such a small area is examined ($\cong$62,500 km^2).

The mean monthly evaporation for each area is shown in Figure 10d. The evaporation results are similar for areas 1 and 2. The evaporation estimates for area 3 are significantly lower owing to the differences in the convergence as shown in Figure 10b. For 8 out of 12 months the evaporation estimates for area 3 are negative, which underscores the error involved in the estimates.

Three principal factors controlling the accuracy of atmospheric flux computations were identified by *Rasmusson* [1968, p. 721] who stated that "in order to obtain satisfactory results, an adequate aerological network must exist, the region considered must not be too small, and the time period over which the observations are averaged must be of sufficient length to render the effect of random errors negligible." Since the atmospheric flux field exhibits significant diurnal variations [*Rasmusson*, 1967], using a finer temporal resolution of wind velocity measurements may alleviate the inaccuracy of atmospheric flux computations. From the study of the upper Mississippi Basin, where the water transport by the nocturnal low-level jet is a significant source of water vapor, *Gutowski et al.* [1997] suggested that 6 hourly rawinsonde data appear to be the minimally acceptable frequency for the characterization of the temporal variability of water transport. This resolution is consistent with that used in this study; thus we believe the temporal resolution of rawinsonde data is not the main source of the estimate errors in area 3. The 12-year average period used in this study is expected to be long enough to give a reliable estimation of average atmospheric convergence. The most probable source of error thus lies in that the spatial resolution of aerological network (2.5° × 2.5°) cannot resolve the significant feature of atmospheric circulation in area 3. Therefore we concluded that the evaporation estimates for area 3 are not consistent with other areas. Also, whatever causes the error in area 3 may in fact bias the estimates for other areas using node 3. This fact reiterates the need to use as large an area as possible in order to yield more accurate estimates of convergence. Since the estimates of evaporation for area 3 were deemed inaccurate, our conclusion drawn from the sensitivity analysis is that the accuracy of the AWB approach is rather poor when the scale of the study area is smaller than $\sim 10^5$ km^2.

Two lumped parameters required for the estimation of evaporation from SWB, the root-zone depth $\bar{D}$ and specific yield $\overline{S_y}$, were estimated as 2 m and 0.08 as previously described in section 3.1. The sensitivity of evaporation estimates from the SWB computations to these parameters can be investigated by varying them within a reasonable range of values. The results are presented in Figures 11a and 11b. The climatology of the evaporation estimates is insensitive to a specified $\bar{D} > 1$ m (Figure 11a). The range of the monthly evaporation estimates is slightly larger when larger $\bar{D}$ is used. The annual evaporation estimates corresponding to $\bar{D} = 50$ cm, 110 cm, and 200 cm are 663.8 mm, 661.9 mm, and 659.8 mm, respectively. Such a result is anticipated since the soil water storage in the deep root zone (1–2 m) in Illinois maintains close to soil field capacity throughout the year [*Hollinger and Isard*, 1994]. However, the specified yield $\overline{S_y}$ affects the evaporation estimates to a slightly larger extent (Figure 11b). Large $\overline{S_y}$ amplifies the effect of groundwater storage, leading to larger summer evaporation estimates and lower winter estimates, due to the opposite sign of the amount $\bar{P} - (\bar{E} + \bar{R})$ in different seasons. Moreover, unrealistic large negative evaporation estimates during the winter months result from specifying $\overline{S_y}$ as large as 0.15, which suggest the representative $\overline{S_y}$ should range between $\sim$0.05 and 0.1.

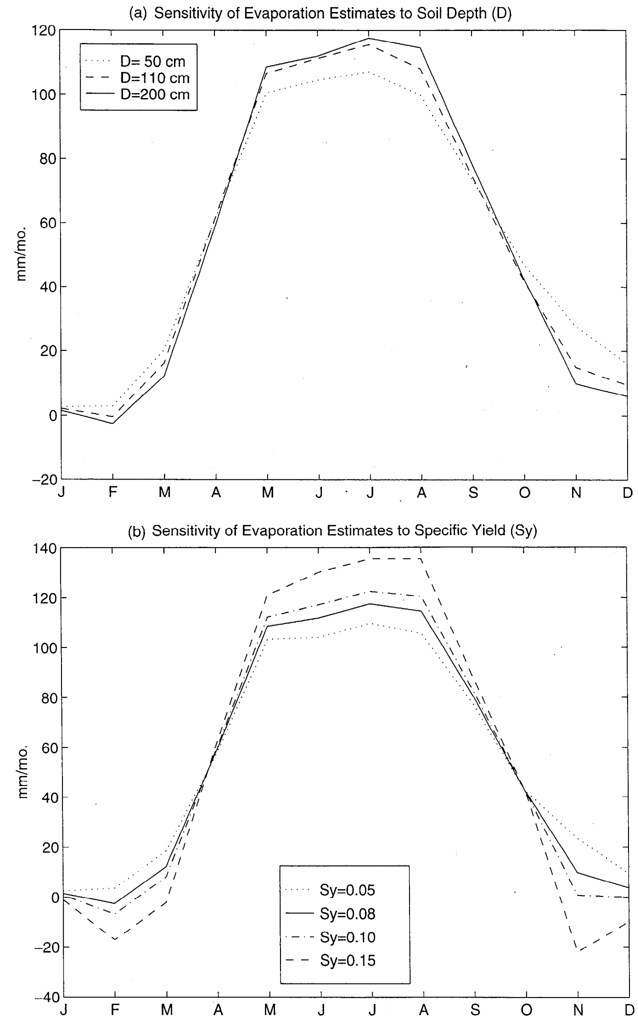

Figure 11. The sensitivity of the seasonal cycle of the evaporation estimates from soil water balance to (a) root-zone depth and (b) specific yield.

7. Conclusions

We have described in this paper the regional-scale hydrological cycle of Illinois, including the land and atmospheric branches, based on an extensive data set of most of the hydrological variables including atmospheric flux of water vapor, precipitation, streamflow, soil water content, groundwater level, and snow depth. Since direct observations on evaporation are not available, two different approaches, based on soil water balance and atmospheric water balance, were applied to estimate the monthly regional evaporation over Illinois from 1983 to 1994. The following conclusions can be drawn from this study:

1. Estimates of evaporation based on the atmospheric water balance approach have been compared to consistent estimates of evaporation based on the soil water balance approach. To our knowledge, this is the first time such a comparison has been made. The annual totals of evaporation estimates from both approaches are consistent with the previous study by the Illinois State Water Survey [*Jones*, 1966]. The climatologies of the monthly evaporation estimates from the two approaches agree reasonably well and within a 10% error. However, substantial differences exist between the two estimates of evaporation for individual months. In light of the fact that independent data sets were used in the two approaches, this result is encouraging: The atmospheric water balance approach has the potential for the estimation of the climatology of regional evaporation, at least for humid regions at a scale similar to that of Illinois ($\sim 10^5$ km^2).

2. Since the magnitude of the change in subsurface storage can be nearly twice as large as runoff at the monthly timescale, the seasonal variability of evaporation estimates based on soil water balance is mainly balanced by the seasonal variability of subsurface storage rather than the lateral water fluxes (runoff). Conversely, the seasonal variability of the evaporation estimates based on atmospheric water balance is almost entirely balanced by the seasonal variability of lateral fluxes of water vapor. On an annual basis, the net storage changes in the soil and atmosphere are both zero. Annual runoff is in close agreement with the convergence of atmospheric water vapor.

3. Although the atmospheric water balance approach has been shown to be promising for the accurate estimation of regional evaporation at a scale of 2×10^5 km^2 (i.e., 500 × 500

km^2), this approach is more prone to error as the size of the study area decreases. This conclusion is obtained from the observed divergence among the evaporation estimates from the atmospheric water balance for small areas, especially for area 3 ($\approx$250 $\times$ 250 km^2), where estimates based on atmospheric water balance approach are not accurate. Therefore the accuracy cannot be warranted if the scale of the study area is $<10^5$ km^2.

4. For the calculation of soil water balance for humid regions such as Illinois, where the groundwater table is rather shallow, the incorporation of the change in groundwater storage is indispensable, since it contributes a significant portion of water storage at the monthly timescale.

As a final note, this regional water balance study has implications for the land-surface parameterization schemes (LSPs) used in GCMs. One important conclusion, obtained from the recent evaluation of 25 LSPs using the observed data from the Cabauw site (i.e., pertaining to phase 2a of Project for Intercomparison of Land-surface Parameterization Schemes (PILPS)), is the existence of considerable inconsistency in the partitioning of annual runoff into its physical components among the various schemes [*Chen et al.*, 1997; *Beljaars and Bosveld*, 1997]. Observational evidence (shallow groundwater table 0–75 cm below surface, deep soil saturated throughout the year, and no overland flow observed) and site topography have indicated that downward drainage from the root zone to a groundwater table and subsequent groundwater discharge is the main mode of runoff at this site. However, more than half of the LSPs generated significant amounts of overland flow and interflow. This is due to the inadequate assumption of free drainage from the root zone commonly used in current LSPs. The identical problem was also found in using Hydrology-Atmosphere Pilot Experiment-Modelisation de Bilan Hydrique-France (HAPEX-MOBILHY) data in PILPS [*Shao and Henderson-Sellers*, 1996]. As emphasized in section 4, the changes in unsaturated and saturated storage in Illinois are equally important with a magnitude comparable to runoff. For the Cabauw site, where similar hydroclimatic characteristics to Illinois exist, measurements of soil moisture, groundwater level, and runoff are not available for model evaluation. The question about the role that groundwater aquifers play in the regional water balance cannot be answered. On the basis of direct observations in Illinois, we have shown from this study that the groundwater aquifer acts as a long-memory reservoir, which stores excessive soil water. For the purpose of improving current LSPs, this study can provide observational support to help in the parameterization of groundwater storage.

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References

Abdulla, F. A., D. P. Lettenmaier, E. F. Wood, and J. A. Smith, Application of a macroscale hydrologic model to estimate the water balance of the Arkansas-Red River basin, *J. Geophys. Res., 101*, 7449–7459, 1996.

Abdulmumin, S., L. O. Myrup, and J. L. Hatfield, An energy balance approach to determine regional evapotranspiration based on planetary layer similarity theory and regularly recorded data, *Water Resour. Res., 23*, 2050–2058, 1987.

Alley, W. M., On the treatment of evapotranspiration, soil moisture accounting and aquifer recharge in monthly water balance models, *Water Resour. Res., 20*, 1137–1149, 1984.

Baumgartner, F., and E. Reichel, *The World Water Balance: Mean Annual Global, Continental and Maritime Precipitation, Evaporation and Runoff*, 179 pp., Elsevier Sci., New York, 1975.

Beljaars, A. C. M., and F. C. Bosveld, Cabauw data for the validation of land surface parameterization schemes, *J. Climate, 10*, 1172–1193, 1997.

Benton, G. S., and M. A. Estoque, Water vapor transfer over the North American continent, *J. Meteorol., 11*, 462–477, 1954.

Benton, G. S., R. T. Blackburn, and V. W. Snead, The role of the atmosphere in the hydrologic cycle, *Trans. A.G.U., 31*, 61–73, 1950.

Berbery, E. H., E. M. Rasmusson, and K. E. Mitchell, Studies of North American continental-scale hydrology using Eta model forecast products, *J. Geophys. Res., 101*, 7305–7319, 1996.

Brubaker, K. L., D. Entehkabi, and P. S. Eagleson, Atmospheric water vapor transport and continental hydrology over the Americas, *J. Hydrol., 155*, 407–428, 1994.

Brutsaert, W., *Evaporation Into the Atmosphere*, 299 pp., D. Reidel, Norwell, Mass., 1982.

Bryan, F., and A. Oort, Seasonal variation of the global water balance based on aerological data, *J. Geophys. Res., 89*, 11,717–11,730, 1984.

Changnon, S. A., G. Achtemeier, S. Hilberg, V. Knapp, R. Olson, W. Roberts, and P. Vinzani, The 1980–1981 drought in Illinois: Causes, dimensions and impacts, Rep. of Investigation 102, 78 pp., Illinois State Water Survey, Champaign, Ill., 1982.

Changnon, S. A., F. A. Huff, and C. F. Hsu, Relations between precipitation and shallow groundwater in Illinois, *J. Climate, 1*, 1239–1250, 1988.

Chen, T. H., et al., Cabauw experimental results from the Project for Intercomparison of Land-Surface Parameterization Schemes, *J. Climate, 10*, 1194–1215, 1997.

Delworth, T. L., and A. S. Manabe, The influence of potential evaporation on the variability of simulated soil wetness and climate, *J. Climate, 1*, 523–547, 1988.

Domenico, P. A., and F. W. Schwartz, *Physical and Chemical Hydrogeology*, 2nd ed., John Wiley, New York, 1990.

Ernstberger, H., and V. Sokolleck, Effects of land use on the hydrology of small basins in Hessen (Federal Republic of Germany), in the *New Approach in Water Balance Computations, Proceedings of the Hamburg Workshop, IAHS Publ., 148*, 147–157, 1985.

Farnsworth, R. K., and E. S. Thompson, Mean monthly, seasonal, and annual pan evaporation for the United States, *Tech. Rep. NWS 34*, Natl. Oceanic and Atmos. Admin., Silver Spring, Md., 1982.

Findell, K. L., and E. A. B. Eltahir, An analysis of the soil moisture-rainfall feedback, based on direct observations from Illinois, *Water Resour. Res., 33*, 725–735, 1997.

Food and Agriculture Organization (FAO)-UNESCO, *Soil Map of the World*, scale 1:5,000,000, UNESCO, Paris, 1975.

Gangopadhyaya, M., G. E. I. Harbeck, T. J. Nordenson, M. H. Omar, and V. A. Uryvaev, Measurement and estimation of evaporation and evapotranspiration, *Tech. Note No. 83, WMO-No. 201.TP.105*, 121 pp., World Meteorol. Org., Geneva, 1966.

Groisman, P. Y., and D. R. Legates, The accuracy of United States precipitation data, *Bull. Am. Meteorol. Soc., 75*, 215–227, 1994.

Gutowski, W. J. Jr., Y. Chen, and Z. Otles, Atmospheric water vapor transport in NCEP-NCAR reanalyses: Comparison with river discharge in the Central United States, *Bull. Am. Meteorol. Soc., 78*, 1957–1969, 1997.

Higgins, R. W., K. C. Mo, and S. D. Schubert, The moisture budget of the Central United States in spring as evaluated from the NCEP/NCAR and the NASA/DAO reanalyses, *Mon. Weather Rev., 124*, 939–963, 1996.

Hollinger, S. E., and S. A. Isard, A soil moisture climatology of Illinois, *J. Climate, 7*, 822–833, 1994.

Johnson, A. I., Specific yield: Compilation of specific yields for various materials, *U.S. Geol. Surv. Water Supply Pap., 1662-D*, 1967.

Jones, D. M., Variability of evapotranspiration in Illinois, *Circular Rep. 89*, 13 pp., Ill. State Water Surv., Champaign, Ill., 1966.

Kalnay, E., et al., The NCEP/NCAR 40 years reanalysis project, *Bull. Am. Meteorol. Soc.*, *77*, 437–471, 1996.

Kattelmann, R., and K. Elder, Hydrologic characteristics and water balance of an alpine basin in the Sierra Nevada, *Water Resour. Res.*, *27*, 1553–1562, 1991.

Kunkel, K. E., S. A. Changnon, C. G. Lonnquist, and J. R. Angel, A real-time climate information system for the Midwestern United States, *Bull. Am. Meteorol. Soc.*, *71*, 1601–1609, 1990.

Larson, L. W., and E. L. Peck, Accuracy of precipitation measurements for hydrologic modeling, *Water Resour. Res.*, *10*, 857–863, 1974.

Leighton, M. M., G. E. Ekblaw, and L. Horberg, Physiographic divisions of Illinois, *Rep. Of Investigation* 129, 39 pp., Ill. State Water Surv., Champaign, Ill., 1948.

Lesack, L. F. W., Water balance and hydrologic characteristics of a rain forest catchment in the central Amazon basin, *Water Resour. Res.*, *29*, 759–773, 1993.

Liang, X., D. P. Lettenmaier, E. F. Wood, and S. J. Burges, A simple hydrologically model of land surface water and energy fluxes for GCMs, *J. Geophys. Res.*, *99*, 14,415–14,428, 1994.

Malek, E., G. E. Bingham, D. Or, and G. McCurdy, Annual mesoscale study of water balance in a Great Basin heterogeneous desert valley, *J. Hydrol.*, *191*, 223–244, 1997.

Mawdsley, J. A., and W. Brutsaert, Determination of regional evapotranspiration from upper air meteorological data, *Water Resour. Res.*, *13*, 539–548, 1979.

McGowan, M., and J. B. Williams, The water balance of an agricultural catchment, III, The water balance, *J. Soil Sci.*, *31*, 245–262, 1980.

Mo, K. C., and R. W. Higgins, Large-scale atmospheric moisture transport as evaluated in the NCEP/NCAR and the NASA/DAO reanalyses, *J. Climate*, *9*, 1531–1545, 1996.

Munley, W. G., and L. E. Hipps, Estimation of regional evaporation for a tallgrass prairie from measurements of properties of the atmospheric boundary layer, *Water Resour. Res.*, *27*, 225–230, 1991.

Oki, T., K. Musiake, H. Matsuyama, and K. Masuda, Global atmospheric water balance and runoff from large river basins, *Hydrol. Process.*, *9*, 411–434, 1995.

Parlange, M. B., W. E. Eichinger, and J. D. Albertson, Regional scale evaporation and the atmospheric boundary layer, *Rev. Geophys.*, *33*, 99–124, 1995.

Peixoto, J. P., and A. H. Oort, The atmospheric branch of the hydrological cycle and climate, in *Variation in the Global Water Budget*, edited by A. Street-Perrot et al., pp. 5–65, D. Reidel, Norwell, Mass., 1983.

Peixoto, J. P., and A. H. Oort, *Physics of Climate*, Am. Inst. of Phys., Woodbury, N. Y., 1992.

Peixoto, J. P., M. Almeida, R. D. Rosen, and D. A. Salstein, Atmospheric moisture transport and the water balance of the Mediterranean Sea, *Water Resour. Res.*, *18*, 83–90, 1982.

Penman, H. L., The water balance of the Stour catchment area, *J. Inst. Water Eng.*, *4*, 457–469, 1950.

Pruitt, W. O., and D. E. Angus, Large weighing lysimeter for measuring evapotranspiration, *Trans. A.S.A.E.*, *3*, 13–15, 1960.

Rasmusson, E. M., Atmospheric water vapor transport and the water balance of North America, 1, Characteristics of water vapor flux field, *Mon. Weather Rev.*, *95*, 403–426, 1967.

Rasmusson, E. M., Atmospheric water vapor transport and the water balance of North America, 2, Large-scale water balance investigations, *Mon. Weather Rev.*, *96*, 720–734, 1968.

Rasmusson, E. M., A study of the hydrology of Eastern North America using atmospheric vapor flux data, *Mon. Weather Rev.*, *97*, 119–135, 1971.

Rasmusson, E. M., Hydrological application of atmospheric vapor-flux analyses, *Hydrol. Rep.*, *11*, 50 pp., World Meteorol. Org., Geneva, 1977.

Roads, J. O., S.-C. Chen, A. K. Guetter, and K. P. Georgakakos, Large-scale aspects of the United States hydrologic cycle, *Bull. Am. Meteorol. Soc.*, *75*, 1589–1610, 1994.

Roberts, G., and R. J. Harding, The use of simple process-based models in the estimate of water balance for mixed land use catchments in East Africa, *J. Hydrol.*, *180*, 251–266, 1996.

Ropelewski, C. F., and E. S. Yarosh, The observed mean annual cycle of moisture budgets over the Central United States (1973–1992), *J. Climate*, in press, 1998.

Rosen, R. D., and A. S. Omolayo, Exchange of water vapor between land and ocean in the Northern Hemisphere, *J. Geophys. Res.*, *86*, 12,147–12,152, 1981.

Salstein, D. A., R. D. Rosen, and J. P. Peixoto, Modes of variability in annual hemispheric water vapor and transport fields, *J. Atmos. Sci.*, *40*, 788–803, 1983.

Savijarvi, H. I., Global energy and moisture budgets from rawinsonde data, *Mon. Weather Rev.*, *116*, 417–430, 1988.

Schicht, R. G., and W. C. Walton, Hydrologic budgets three small watersheds in Illinois, *Rep. Of Invest. 49*, 40 pp., Ill. State Water Surv., Champaign, Ill., 1961.

Shao, Y., and A. Henderson-Sellers, Validation of soil moisture simulation in landsurface parameterization schemes with HAPEX data, *Global Planet. Change*, *13*, 11–46, 1996.

Shukla, J., and Y. Mintz, Influence of land-surface evapotranspiration on the Earth's climate, *Science*, *215*, 1498–1500, 1982.

Shuttleworth, W. J., *Evaporation*, *Rep. No. 56*, Inst. of Hydrol., Willingford, Oxfordshire, U. K., 1979.

Starr, V. P., and J. Peixoto, On the global balance of water vapor and the hydrology of deserts, *Tellus*, *10*, 189–194, 1958.

Street-Perrott, A., M. Beran, and R. Ratcliffe, *Variations in the Global Water Budget*, 518 pp., D. Reidel, Norwell, Mass., 1983.

Sugita, M., and W. H. Brustaert, Daily evaporation over a region from lower boundary-layer profiles measured with radiosondes, *Water Resour. Res.*, *27*, 747–752, 1991.

Thronthwaite, C. W., An approach toward a rational classification of climate, *Geogr. Rev.*, *38*, 55–94, 1948.

Van der Beken, A., and A. Herrmann (Eds.), *New Approaches in Water Balance Computations*, *IAHS Publ. 148*, 1–167, 1985.

Vandewiele, G. L., C.-Y. Xu, and N. L. Win, Methodology and comparative study of monthly water balance models in Belgium, China and Burma, *J. Hydrol.*, *134*, 315–347, 1992.

Xu, C.-Y., and J. S. S. Halldin, Regional water balance modelling in the NOPEX area: Development and application of monthly water balance, *J. Hydrol.*, *180*, 211–236, 1996.

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