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Sonia I. Seneviratne, Pedro Viterbo, Daniel Lüthi, and Christoph Schär

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ERA-40 Project Report Series No. 9

# Inferring changes in terrestrial water storage using ERA-40 reanalysis data: The Mississippi River basin

Sonia I. Seneviratne[1], Pedro Viterbo, Daniel Lüthi[1], and Christoph Schär[1]

Research Department

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[1]ETH, Zurich

**Abstract**

Terrestrial water storage is an essential part of the hydrological cycle, encompassing crucial elements of the climate system such as soil moisture, groundwater, snow, and land ice. On a regional scale, it is however not a readily measured variable and observations of its individual components are scarce. This study investigates the feasability of estimating monthly terrestrial water-storage variations with the water-balance method, using the following three variables: water vapour flux convergence, atmospheric water content, and river runoff. The two first variables are available with high resolution and good accuracy in present reanalysis datasets, and river runoff is commonly measured in most parts of the world. The applicability of this approach is tested in a 10-year (1987-1996) case study for the Mississippi river basin. Data used include ERA-40 reanalysis data from the European Centre for Medium-Range Weather Forecasts (water vapour flux and atmospheric water content) and runoff observations from the United States Geological Survey.

Results are presented for the whole Mississippi River basin and its subbasins, and for a smaller domain covering Illinois, where direct measurements of the main components of the terrestrial water storage (soil moisture, groundwater level, and snow cover) are available. The water-balance estimates of monthly terrestrial water-storage variations show excellent agreement with observations taken over Illinois. The mean seasonal cycle as well as interannual variations are captured with notable accuracy. Despite this excellent agreement, it is not straightforward to integrate the computed variations over longer time periods, as there are small systematic biases in the monthly changes. These biases likely result from inaccuracies of the atmospheric assimilation system used to estimate the atmospheric water vapour convergence. Nevertheless, the results suggest that the critical domain size for water-balance computations using high resolution reanalysis data such as ERA-40 is much smaller than for raw radiosonde data. The Illinois domain has a size of only $\sim 2\text{x}10^5$ km$^2$ and is shown to be suitable for the computation of the water-balance estimates. A comparison for other regions would be needed in order to confirm this result.

# 1 Introduction

Terrestrial water storage is an essential part of the hydrological cycle, encompassing crucial elements of the climate system such as soil moisture, groundwater, snow and land ice, as well as surface water and biomass water. Beside their key role in the climate system, soil moisture and groundwater are also of essential importance for agriculture and the supply of freshwater. Despite its relevance for both climate and human civilisation, continental and sub-continental terrestrial water storage is not a readily measured quantity and little knowledge is available on its individual components, most of the available observations being of very limited temporal or spatial scope.

In the tropics and the mid-latitudes, soil moisture is generally the main element contributing to seasonal changes in terrestrial water storage. Its key role for the global- and regional-scale climate (through its impact on the partitioning of the sensible and latent heat fluxes) has been recognized in various observational (e.g. Betts et al. 1996, Findell and Eltahir 1997) and modelling studies (e.g. Shukla and Mintz 1982, Milly and Dunne 1994, Schär et al. 1999, Koster et al. 2000). More recently, it has also been shown to significantly impact numerical weather predictions (e.g. Beljaars et al. 1996, Viterbo and Betts 1999, Fukutome et al. 2001, see also Viterbo and Beljaars 2003 for a review). Soil moisture is moreover important for assessing impacts of climate change, but its potential evolution with greenhouse gas warming is still unclear (Wetherald and Manabe 1999, Seneviratne et al. 2002).

In view of its relevance for climate, recent initiatives have tried to palliate the lack of information available on soil moisture. The few observational datasets (mostly for the State of Illinois, the former Soviet Union, Mongolia, and China) have recently been regrouped and are now distributed on the internet by the Soil Moisture Data Bank at Rutgers University (Robock et al. 2000). The Global Soil Wetness Project (GSWP), an ongoing modeling activity of the GEWEX (Global Energy and Water Cycle Experiment) Global Land-

Atmosphere System Study (GLASS) and of the International Satellite Land-Surface Climatology Project (ISLSCP), is another project aiming at obtaining additional information on soil moisture (Dirmeyer et al. 1999). Its goal is to produce large-scale datasets of soil moisture, temperature, runoff, and surface fluxes, by driving uncoupled land-surface schemes using externally specified surface forcings and standardized soil and vegetation distributions. The results of its first initiative (GSWP1) for the years 1987 and 1988 show that all the tested land surface schemes have problems in reproducing the actual soil moisture value for the studied regions (Entin et al. 1999). A second phase of the project (GSWP2) has been recently started to seek an improvement of the produced datasets.

Groundwater is another important component of terrestrial water storage in the tropics and mid-latitudes. There are, however, very few long-term measurements of this variable, and no project comparable to the Soil Moisture Data Bank has been initiated for groundwater as yet. With the exception of water-balance estimates from the Soviet Literature (Zekster and Loaiciga 1993), there is little information available on large-scale seasonal and interannual groundwater variations.

Snow is the only other component of the total terrestrial water storage which can induce similar variations as soil moisture and groundwater in mid- and high-latitude regions. Measurements of snow areal extent and snow depth are available for various areas, but the relevant variable for hydrological modelling is the snow mass per unit area or snow mass (snow water equivalent), which is not measured. The ratio of snow mass to the depth of snow cover depends greatly on the snow type, age, amount of melt and compaction, as well as on the temperature and atmospheric humidity at the time of snowfall (Pomeroy and Gray 1995), and can not be precisely inferred from snow depth without corresponding snow density measurements. Land ice only plays an important part in mountainous areas and in polar regions, while changes in surface water and biospheric water are comparatively negligible.

Although little information is available on terrestrial water storage at the moment, this might change in the future. Soil moisture, for instance, is measured and studied in various new projects such as the Oklahoma mesonet (Basara and Crawford 2000). Moreover, remote sensing based on microwave radiometry is now able to provide datasets of near-surface soil moisture (e.g. Jackson et al. 1999), and data assimilation techniques can be used to infer root zone soil water from these measurements (Calvet and Noilhan 2000). Other techniques might also allow to estimate evaporation over large areas based on satellite remote sensing data (Salvucci 1997), thus providing indirect information on soil moisture depletion. Remote sensing also holds promise for additional observations of snow cover: Mapping of snow areal extent is already performed operationally with visible satellite imagery, and further developments of microwave techniques should enable the monitoring of snow mass (Rango 1996). A currently ongoing satellite project, the Gravity Recovery and Climate Experiment (GRACE) launched in 2001, might even provide estimates of variations in total terrestrial water storage (Wahr et al. 1998). Nevertheless, effective use of remote sensing data to estimate geophysical quantities requires a-priori estimates which, in the case of globally distributed soil water budgets, can only come from model estimates. Another caveat of new remote-sensing techniques is their restriction to the future climate states, while past variations must be assessed from existing observations.

This study investigates the combined atmospheric and terrestrial water balance approach for obtaining estimates of changes in terrestrial water storage for continental and sub-continental areas. The basic concept of using atmospheric data to estimate the terrestrial water balance was first presented in the 1950s, in pioneer studies by Benton and Estoque (1954) and Starr and Peixoto (1958). It was used in various subsequent studies (e.g. Rasmusson 1968, Alestalo 1983) and has received more attention in recent years, thanks to the availability of high-resolution atmospheric data (e.g. Oki et al. 1995, Matsuyama and Masuda 1997, Yeh et al. 1998, Berbery and Rasmusson 1999, Masuda et al. 2001). The advantage of this method is that it is based on variables which have been routinely measured for decades.

Here, we present a 10-year case study for the Mississippi river basin using the new reanalysis product from the European Center for Medium-Range Weather Forecasts (ECMWF), ERA-40 (Simmons and Gibson 2000), and streamflow data from the United States Geological Survey (USGS). Computed estimates of monthly variations in terrestrial water storage are presented for the Mississippi River basin and its major subbasins, as well as for a smaller domain covering the State of Illinois, where observations of various components of the terrestrial water storage are available (soil moisture, groundwater, and snow cover). This study is meant as a pilot study and aims at establishing the validity of the approach, but application to other catchments and continents is currently underway.

The paper is structured as follows. The tested methodology, the investigated region, and the employed datasets are presented in section 2. The observations from Illinois used for the validation are presented in section 3. The computed estimates of terrestrial water-storage variations for the investigated domains are presented in section 4, and the validation for the Illinois domain is discussed in section 5. For comparison purposes, the ERA-40 soil moisture in Illinois is shortly discussed in section 6. A summary of the main results and the conclusions are provided in section 7.

# 2 Methodology

## 2.1 The water balance equations

This section presents the terrestrial, atmospheric, and combined water balance equations. Good reviews on this topic are given in Peixoto and Oort (1992, chapter 12), Yeh et al. (1998) and Oki (1999).

The terrestrial branch of the hydrological cycle is governed by the following equation:

$$\frac{\partial S}{\partial t} = -R_s - R_u + (P - E) \quad , \tag{1}$$

where $S$ represents the terrestrial water storage of the given area, $R_s$ the surface runoff, $R_u$ the subterranean runoff, $P$ the precipittion, and $E$ the evapotranspiration.

Within a given river basin, most of the area groundwater runoff can be considered to be discharged into streams and hence measured together with surface runoff (Rasmusson 1968). For large areas, (1) can thus be simplified to:

$$\left\{ \overline{\frac{\partial S}{\partial t}} \right\} = -\{\overline{R}\} + \{\overline{P} - \overline{E}\} \quad , \tag{2}$$

where the overbar denotes a temporal average (e.g. monthly means), $\{\}$ a spatial average over the region, and $\overline{R} \cong \overline{R}_s + \overline{R}_u$ is the measured runoff.

The water balance for the atmospheric branch of the hydrological cycle is given as follows:

$$\frac{\partial W}{\partial t} + \frac{\partial W_c}{\partial t} = -\nabla_H \cdot \mathbf{Q} - \nabla_H \cdot \mathbf{Q}_c - (P - E) \quad , \tag{3}$$

where $W$ represents the column storage of water vapour (sometimes refered to as precipitable water), $W_c$ the column storage of liquid and solid water, $\mathbf{Q}$ the vertically integrated two-dimensional water vapour flux, and $\mathbf{Q}_c$ the vertically integrated two-dimensional water flux in the liquid and solid phases. The operator $(\nabla_H \cdot)$ represents the horizontal divergence. $\mathbf{Q}$ is the vapour flux vector and is defined as follows:

$$\mathbf{Q} = \int_0^{p_0} q\mathbf{v}\frac{dp}{g} \quad , \tag{4}$$

where $q$, $\mathbf{v}$, $g$, $p$ and $p_0$ represent the specific humidity, wind vector, gravitational acceleration, pressure, and surface pressure at the ground.

Generally, both the time rate of change of the liquid and solid water in clouds and their horizontal transports can be neglected (Peixoto and Oort 1992), and equation (3) simplifies to:

$$\frac{\partial W}{\partial t} = -\nabla_H \cdot \mathbf{Q} - (P - E) \tag{5}$$

Averaging (5) in space and time over a large area (similarly as in (2)) leads to following equation:

$$\left\{\overline{\frac{\partial W}{\partial t}}\right\} = -\left\{\overline{\nabla_H \cdot \mathbf{Q}}\right\} - \{\overline{P} - \overline{E}\} \tag{6}$$

The term $\{\overline{P} - \overline{E}\}$ can be eliminated between equations (6) and (2) to give:

$$\left\{\overline{\frac{\partial S}{\partial t}}\right\} = -\left\{\overline{\nabla_H \cdot \mathbf{Q}}\right\} - \left\{\overline{\frac{\partial W}{\partial t}}\right\} - \{\overline{R}\} \tag{7}$$

In this combined equation, the monthly variations in terrestrial water storage of the studied region can be expressed as the sum of three terms: the water vapour flux convergence, the changes in atmospheric vapour content, and the measured river runoff. The term $\left\{\overline{\frac{\partial W}{\partial t}}\right\}$ is usually negligible for annual means, but not for monthly means, particularly during spring and fall (Rasmusson 1968).

Note that for long-term means (multi-year averages) the tendency terms are negligible, and (7) simplifies to:

$$\{\overline{R}\} = -\left\{\overline{\nabla_H \cdot \mathbf{Q}}\right\} \quad , \tag{8}$$

i.e. for any climate equilibrium within a given hydrologic unit, the long-term water input from the atmosphere (vapour flux convergence) has to be balanced by the long-term net water output at the surface (runoff). In studies using combined atmospheric and terrestrial water-balance computations, (8) can be viewed as a criterion for evaluating the agreement between the atmospheric and hydrological data on the spatial scale under consideration (Yeh et al. 1998).

The accuracy of atmospheric water-balance computations is highly dependent on the size of the area investigated. Early studies by Rasmusson (1968, 1971) using raw radiosonde data over North America suggested that for this type of data water-balance computations can give reliable estimates of $\{\overline{P} - \overline{E}\}$ when averaging over areas of $2 \times 10^6$ km$^2$ or larger, but can become quite erratic as the size of the area is reduced to less than $10^6$ km$^2$. The limiting factors for capturing regional-scale features were identified as the density of rawinsonde observations and their sampling frequency (Berbery and Rasmusson 1999).

Although the rawinsonde network over North America has not changed much in the last decades, there is now a large amount of new data available, such as wind and temperature observations from aircraft, data from the wind profiler demonstration network, and various types of surface data (Berbery et al. 1996). Reanalysis data which optimally combine these sources of information with model integrations provide comparatively high resolution atmospheric vapour flux data, which are likely to allow reliable water-balance computations for domains smaller than the critical size of $10^6$ km$^2$. This is confirmed by results of recent studies making use of reanalysis data for atmospheric water-balance computations, which suggest that the critical size limit with such data might be lowered to $5 \times 10^5$ km$^2$ (Berbery and Rasmusson 1999) or even $1 \times 10^5$ km$^2$ (Yeh et al. 1998).

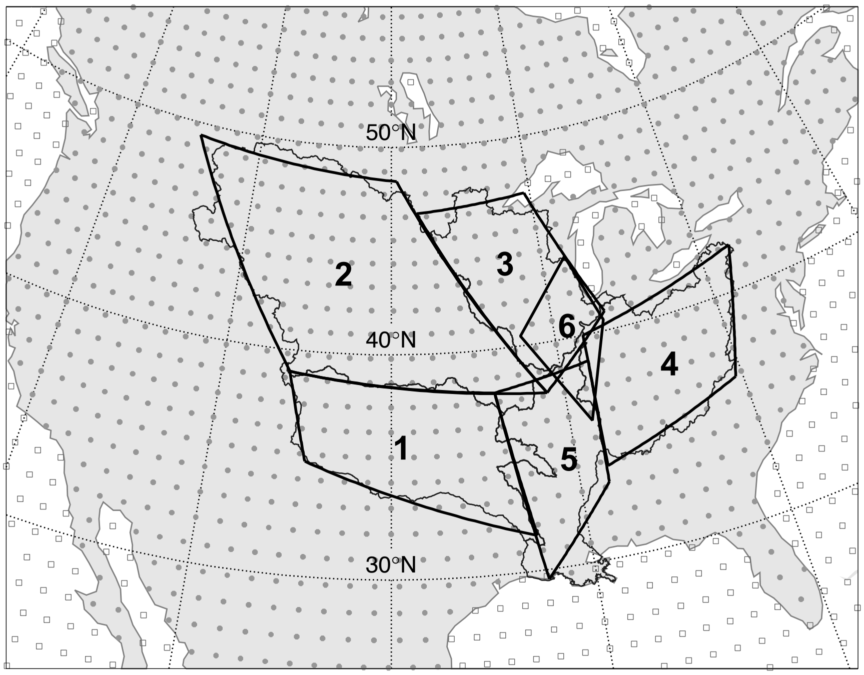


*Figure 1: Major Mississippi subbasins and their approximation in the ECMWF reanalysis model (domains 1-5); domain 6 covers the State of Illinois. Model grid points are represented by closed circles (land) and open squares (sea). (adapted from Betts et al. 2003)*

## 2.2 Investigated region

The Mississippi River basin was chosen for the testing of the proposed methodology, based on various reasons. First, it is a well studied region characterized by abundant meteorological and hydrological datasets: It was for instance the focus area of the GEWEX Continental-Scale International Project (GCIP, now GEWEX Americas Prediction Project or GAPP; see e.g. Coughlan and Avissar 1996, Roads et al. 2003). Second, the existence of comprehensive observations of terrestrial water storage in Illinois (soil moisture, groundwater and snow measurements, see section 3) is another asset of this region. Finally, various studies have performed atmospheric water-balance computations for some of the Mississippi subbasins or Illinois (e.g. Gutowski et al. 1997, Yeh et al. 1998, Ropelewski and Yarosh 1998, Berbery and Rasmusson 1999), which allows some comparisons with our results. Note as well that there are several studies on moisture transport and recycling over the Mississippi basin (e.g. Brubaker et al. 2001, Bosilovich and Schubert 2001), which are also useful for interpreting the present results.

The Mississippi River basin is divided in 5 main subbasins (Fig. 1) The Arkansas and Red River basins, 2) the Missouri basin, 3) the upper Mississippi basin, 4) the Ohio and Tennessee River basins, and 5) the lower Mississippi River basins. Note that there are no actual runoff measurements for domain 5, where runoff can only be computed as the difference between the total runoff of the whole Mississippi River basin (station Vicksburg) and the runoff measured for the other subbasins. Therefore, we only focus on the first four subbasins, on the sum of all 5 subbasins (whole Mississippi River basin), and on a smaller domain approximately covering Illinois (domain 6, Fig. 1). Note that all 6 domains are smaller than the Rasmusson threshold of $2 \times 10^6$ km$^2$ (Table 1), but larger than the critical size of $1 \times 10^5$ km$^2$ identified by Yeh et al. (1998) for their water-balance computations over Illinois. The datasets used for the atmospheric vapour flux, the changes in atmospheric water content, and the areal runoff are described in the following subsection.

| | Domain | Quadrilateral coordinates | Number of Grid points[1] | Quadrilateral Area [km$^2$] |
|---|---|---|---|---|
| 1 | Arkansas-Red | 106.0W, 39.0N; 94.0W, 38.0N; 92.6W, 31.5N; 104.7W, 35.0N | 43 | 604,054 |
| 2 | Missouri | 114.7W, 49.4N; 99.6W, 48.3N; 91.0W, 37.7N; 106.0W, 39.0N | 94 | 1,299,229 |
| 3 | Upper Mississippi | 98.2W, 46.7N; 90.4W, 47.2N; 86.7W, 40.9N; 91.0W, 37.7N | 37 | 512,722 |
| 4 | Ohio-Tennessee | 88.3W, 40.0N; 78.0W, 42.1N; 80.4W, 36.3N; 88.4W, 34.0N | 34 | 462,521 |
| 5 | Lower Mississippi | 94.0W, 38.0N; 88.3W, 38.8N; 88.5W, 33.3N; 92.3W, 29.6N | 27 | 382,664 |
| | Whole Mississippi | – sum of quadrilaterals 1-5 – | 235 | 3,261,190 |
| 6 | Illinois | 92.0W, 40.5N; 88.5W, 43.5N; 87.0W, 40.5N; 88.5W, 36.0N | 15 | 203,549 |

[1]reduced Gaussian grid

*Table 1: Quadrilaterals used for the computation of the ERA-40 domain mean fields*

## 2.3 Datasets

### *2.3.1 ERA-40: Computed water vapour flux divergence and atmospheric water content*

Vertically integrated water vapour fluxes and atmospheric water-content estimates are taken from the latest ECMWF reanalysis data product ERA-40 (Simmons and Gibson 2000). The full ERA-40 reanalysis dataset, recently completed, covers 45 years of data (September 1957 to August 2002). Here we use 10 years of data, covering the period 1987 to 1996.

The ERA-40 model uses a T159 spherical harmonic representation of the atmospheric dynamical and thermodynamical fields, and a grid-point representation of humidity and cloud variables, using the so-called reduced Gaussian grid (Hortal and Simmons 1991). This grid has an almost uniform distribution of grid points on the sphere, with a grid-spacing of 112 km. There are 60 levels in the vertical, with a hybrid sigma-pressure coordinate between the surface and 0.1 hPa. Given its importance for the transport of water vapour, it is worth mentioning the high vertical resolution in the lower troposphere: The lowest model level is at 10 m above the surface, and there are 8, 11, 15, 17, and 22 levels below 500, 1000, 2000, 3000, and 5000 m, respectively. The reanalysis uses a three-dimensional variational assimilation system (Courtier et al. 1998) with a 6-hour analysis cycle. Documentation of the Integrated Forecast System (IFS), cycle 23r4, is available at http://www.ecmwf.int/research/ifsdocs/index.html. A summary and discussion of the observations available at different times during the 40-year reanalysis period is available at http://www.ecmwf.int/research/era/Observations/.

Trenberth (1997) thoroughly discusses issues related to budget computations from analysis fields. In particular, vertically integrated budgets based on pressure level fields have several disadvantages (see also Trenberth and Guillemot 1995): (a) Insufficient detail to resolve model transport in the lower troposphere, particularly detrimental for moisture transport (see also Dirmeyer and Brubaker 1999); (b) the equation of continuity is not fully satisfied by the assimilation cycle, with the consequence that all budget values will see artificial sources/sinks of mass (Trenberth et al. 1995); (c) the lower boundary condition is more complex to formulate than in terrain-following coordinates; and (d) extrapolation below the earth's surface is arbitrary and pollutes the budget values near orography. The obvious alternative is to compute the budgets from the fields defined on the model (hybrid) vertical coordinates. However, model level computations must be performed at maximum resolution. Truncation of model level data is an ill-defined operation, since the vertical coordinate changes with the truncation (Trenberth 1995).

For all the above reasons, the computation of the divergence of the water vapour flux is done as follows. The wind components on terrain-following model levels are transformed from spherical harmonics into the grid-point space (reduced Gaussian grid) and multiplied by the specific humidity at each level. The horizontal fluxes are then integrated over the whole depth of the atmosphere. In order to compute the divergence, and since

| | | River | Station | Basin Area [$km^2$] |
|---|---|---|---|---|
| Mississippi Subbasins (domains 1-5) | (a) | Arkansas | Van Buren, Arkansas | 389,900 |
| | (b) | Red | Index, Arkansas | 124,300 |
| | (c) | Missouri | Kansas City, Missouri | 1,256,100 |
| | (d) | Missouri | St. Louis, Missouri | 1,804,500 |
| | (e) | Ohio | Metropolis, Illinois | 525,500 |
| | (f) | Mississippi | Vicksburg, Mississippi | 2,952,600 |
| Illinois (domain 6) | (g) | Illinois | Valley City, Illinois | 69,200 |
| | (h) | Rock | near Joslin, Illinois | 24,700 |
| | (i) | Kaskaskia | near Venedy Stn, Illinois | 11,400 |

*Table 2: Employed streamflow data (USGS)*

the reduced Gaussian grid is irregular in latitude-longitude, the fluxes are transformed into spectral space for the computation of the divergence. The resulting spectral field is transformed back to obtain the divergence of the integrated water vapour flux on the reduced Gaussian grid. Note that no truncation or interpolation (either vertical or horizontal) is involved. We calculate the vapour flux divergence for each analysis cycle, i.e. at 00, 06, 12 and 18. All figures shown in the paper are based on longer-term averages of these data.

The water-vapour flux divergence and the atmospheric water content are averaged for the chosen domains using latitude-longitude quadrilaterals, following the same procedure as Betts et al. (1998, 1999, 2003). The quadrilaterals are pictured in Fig. 1 and listed in Table 1.

#### *2.3.2 USGS surface streamflow data*

The daily streamflow data was downloaded from the USGS web site (http://water.usgs.gov/nwis). A list of the employed stations is given in Table 2.

For the Mississippi subbasins, measurements from the following 6 stations are used: Arkansas River at Van Buren, Red River at Index, Missouri River at Kansas City, Missouri River at St. Louis, Ohio River at Metropolis, and Mississippi River at Vicksburg. These stations are the same as the ones used by Betts et al. (1999); the streamflow of domains 1-4 and of the whole Mississippi river basin is computed in the same way as in their study (Table 3).

For the computation of the mean runoff in Illinois, we use the same procedure as Yeh et al. (1998). Daily discharge measurements are used from the three following hydrological stations: Illinois River at Valley city, Rock River near Joslin, and Kaskaskia River near Venedy Station (Table 2). These three streamflow-gauging stations are located as far downstream as possible, with the largest drainage areas along the three major rivers in Illinois (Yeh et al. 1998). Their integrated monthly discharges in mm/d are weighted by drainage areas in order to obtain an estimation of average streamflow in Illinois (Table 3).

## 3 Observed terrestrial water storage in Illinois

Illinois has a unique and comprehensive collection of hydrological datasets, including measurements of soil moisture, groundwater, and snow cover. Therefore, it is an ideal region for the validation of the methodology

| | Domain | Computation of Domain streamflow[1] | Drainage area [$km^2$] |
|---|---|---|---|
| 1 | Arkansas-Red | (a)+(b) | 514,200 |
| 2 | Missouri | (c) | 1,256,100 |
| 3 | Upper Mississippi | (d)-(c) | 548,400 |
| 4 | Ohio-Tennessee | (e) | 525,500 |
| 5 | Lower Mississippi | - | 108,400 |
| | Whole Mississippi Basin | (f) | 2,952,600 |
| 6 | Illinois | *(g),(h),(i)* [2] | - |

[1]the letters refer to the streamflow at the gauging stations listed in Table 2
[2]for Illinois, areal runoff is computed as the sum of the area-weighted runoff of the three considered catchments

*Table 3: Computation of streamflow for the domains 1-4, for the whole Mississippi River basin (sum of domains 1-5), and for Illinois (domain 6). Note that there are no actual runoff measurements for domain 5 (see section 2.b).*

presented in this study. Here we briefly describe the datasets used for the validation and the observed values of terrestrial water storage computed based on these datasets.

### 3.1 Soil moisture

The soil moisture dataset was collected by the Illinois State Water Survey (ISWS). It is described in Hollinger and Isard (1994) and can be downloaded from the web site of the Global Soil Moisture Data Bank (Robock et al. 2000). The dataset includes 19 (mostly grass-covered) sites scattered throughout the State of Illinois. Nine of the time series begin in 1981, seven in 1982, two in 1986, and one in 1991. The measurements are realised with neutron probes calibrated by the gravimetric technique. They are taken once (November through February) to twice (March through October) a month at 11 different soil layers down to 2 m below the surface.

### 3.2 Groundwater

The ISWS also has an extensive network of groundwater measurements in Illinois (Changnon et al. 1988). Here, we use shallow well data from the ISWS Water and Atmospheric Resource Monitoring (WARM) Program, consisting of monthly measurements of shallow groundwater levels at 17 wells located far away from pumping stations and streams. The wells range in depth from 3 to 24 m and are in communication with the local unconfined aquifer. The average water table levels range between 1 and 10 m below the surface. The measurements are generally conducted at the end of each month (K. Hlinka, ISWS, personal communication, 2002). We therefore compute the monthly changes in groundwater level as the value of the given month minus the value of the preceding month. Missing data are replaced by interpolated values.

The change in groundwater storage $\frac{\partial S_{GW}}{\partial t}$ can be computed from the change in groundwater level with the following equation:

$$\frac{\partial S_{GW}}{\partial t} = S_y \frac{\partial H}{\partial t} \quad , \tag{9}$$

where $H$ is the groundwater level and $S_y$ is the specific yield, i.e. the fraction of water volume that can be drained by gravity in an unconfined aquifer (Domenico and Schwartz 1990).

For regional-scale time averaged values, one can assume (Yeh et al. 1998):

$$\left\{\overline{\frac{\partial S_{GW}}{\partial t}}\right\} = \{S_y\}\left\{\overline{\frac{\partial H}{\partial t}}\right\} \tag{10}$$

Here, we use $\{S_y\}$ = 0.08 as a mean value for Illinois following Yeh et al. (1998). Their choice of $\{S_y\}$ was based on existing measurements of $S_y$ at various small Illinois watersheds and on soil texture considerations (silt loam in most parts of the state).

### 3.3 Snow

The snow dataset employed was provided by the Midwest Regional Climate Center (MRCC, http://mrcc.sws.uiuc.edu/). It includes observations of snowfall, snow depth, precipitation and temperature for 32 stations within Illinois. The choice of the stations was based on the following criteria: 1) the availability of precipitation and temperature measurements for the whole time period 1961-2000, and 2) less than 10% of missing data either in snow depth or snowfall.

A snow density value of 100 kg/m$^3$ is used to convert the monthly changes in snow depth into changes in snow mass. This is a rather low value which is normally more appropriate for fresh snow. Therefore we also made a comparison with a conversion depending on snow depth using the equations described in Pomeroy and Gray (1995). The values obtained with this second method are about twice as large (not shown), but do not induce any significant change in the computed total terrestrial water storage and its monthly variations due to the very small values of snow mass compared to soil moisture and groundwater (see next section). In their study of the hydroclimatology of Illinois, Yeh et al. (1998) mention the state's location upwind of Lake Michigan as the main cause for the comparatively low snow amounts in this region.

### 3.4 Total terrestrial water storage

Figure 2 displays time series of the Illinois observations of soil moisture, groundwater storage, and snow mass, as well as of their total, for the years 1987 to 1996. The groundwater storage is displayed relative to its deepest value within the time series (in October 1988).

The main components contributing to changes in total terrestrial water storage in Illinois are soil moisture and groundwater, while the contribution of the snow reservoir is negligible in comparison. Soil moisture and groundwater have very similar temporal evolutions and are characterized by a lag-correlation of about 1 month. The late spring drought of 1988, the drought of 1991, and the summer flood of 1993, can be easily recognized in both datasets. Interestingly, the computed total terrestrial water storage compares well with the results of Rodell and Famiglietti (2001) for the same region, though they use a different method for the computation of the groundwater storage. They also account for the water stored in the intermediate zone and for occasional overlaps of the soil moisture and groundwater layers when the water table rises above 2 m depth, two aspects which are not considered here. Therefore, these effects are likely to be small relative to the computed total terrestrial water storage.

The monthly mean annual cycle of terrestrial water storage, as well as the standard deviation of the monthly values for the 10 years considered, are displayed in Table 4. On average, terrestrial water storage in Illinois is highest in April and lowest in October. The April maximum is very typical of the northern mid-latitudes (P.A. Dirmeyer, personal communication, 2003), and marks the end of the winter period with surplus of precipitation relative to evapotranspiration; similarly, the October minimum marks the end of the regime with surplus of

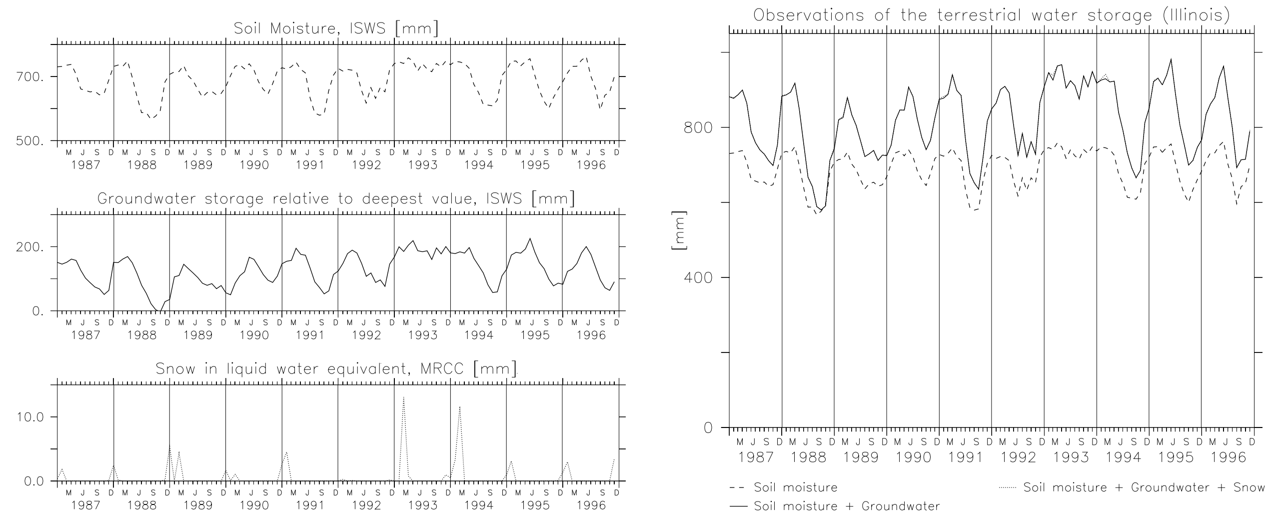


*Figure 2: Observations [mm] of soil moisture, snow mass, and groundwater storage (groundwater level*$\{S_y\}$) in Illinois (left), and their total (terrestrial water storage, right).*

evapotranspiration relative to precipitation. Note that year-to-year variations are lowest in April (saturated soil) and have their highest spread in October, at the end of the drying season. The relatively high spread observed in July corresponds to a maximum in soil moisture variance, probably associated to variability in convective precipitation, while the October maximum is linked to a maximum of groundwater variance (not shown). Inspection of interannual variability shows that the monthly variance figures are strongly affected by the extreme years of 1988 and 1993, and might not be representative of a longer period.

Figure 3 displays the monthly variations of terrestrial water storage and its components in Illinois. The monthly variations in soil moisture and groundwater are of the same order of magnitude, while the contribution of the snow reservoir is negligible in comparison.

As mentioned in section 2.1, variations in terrestrial water storage should cancel out for long-term averages (equation 8). The time scale on which this occurs is generally assumed to be of the order of one (e.g. Oki et al. 1995) to several years (e.g. Gutowski et al. 1997). Table 5 displays annual and 10-year means of the observed variations in terrestrial water storage in Illinois. As expected, the 10-year mean storage variations clearly cancel out, however, annual mean variations can be significant mostly due to the contribution from groundwater storage. In 1988 and 1990, for instance, the yearly variations in terrestrial water storage (groundwater) amount to -0.37 mm/d (-0.31 mm/d) and +0.43 mm/d (+0.26 mm/d), respectively. This represents about 50% of the average streamflow and convergence in Illinois (see section 4.2), confirming that it is generally not feasible to neglect annual changes in basin water storage when inferring mean streamflow from vapour flux convergence for a given year (e.g. Oki et al. 1995).

## 4 Results: Water-balance estimates of terrestrial water-storage variations

The outline of this section is as follows. First, the ERA-40 water vapour flux and its divergence are briefly analysed (subsection 4.1). The computed monthly terrestrial water-storage variations for the 10-year climatology and the individual years are then presented in subsections 4.2 and 4.3, respectively. Finally, subsection 4.4 discusses the long-term imbalances between the computed vapour flux convergence and the

| | mean [mm] | standard deviation |
|---|---|---|
| January | 841.4 | 69.0 |
| February | 872.2 | 57.8 |
| March | 889.2 | 42.3 |
| April | 907.0 | 33.3 |
| May | 894.3 | 45.5 |
| June | 863.4 | 76.3 |
| July | 804.6 | 83.9 |
| August | 763.4 | 77.5 |
| September | 720.6 | 75.8 |
| October | 718.2 | 94.2 |
| November | 723.3 | 78.8 |
| December | 800.1 | 71.7 |

*Table 4: Mean annual cycle of terrestrial water storage in Illinois (1987-1996)*

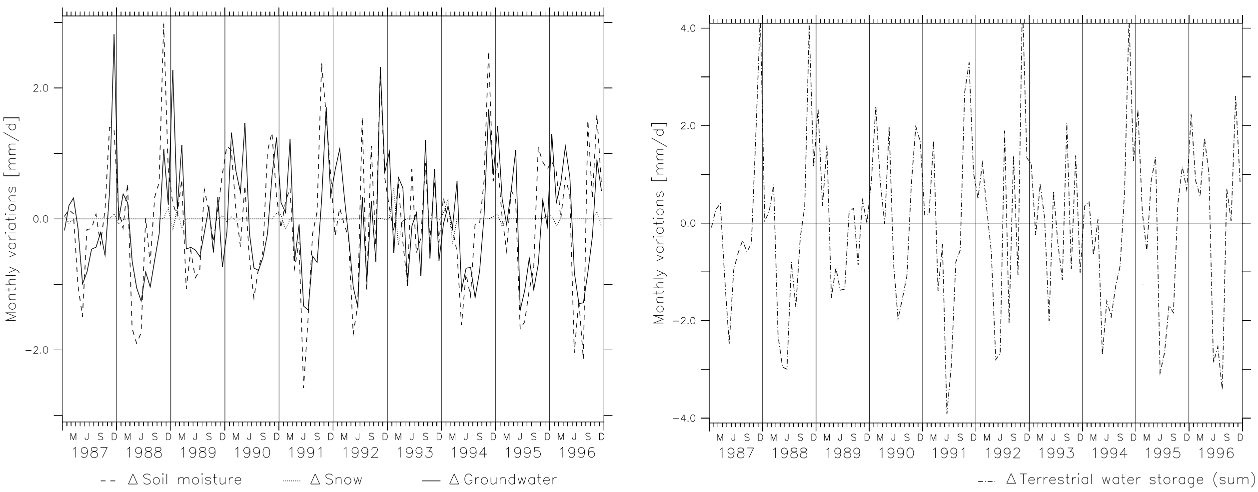


*Figure 3: Observed monthly variations [mm/d] of soil moisture, snow mass, and groundwater storage (groundwater level*$\{S_y\}$) in Illinois (left), and the associated total terrestrial water storage (right).*

| | annual mean storage variations [mm/d] | | | | | | | | | | mean absolute | 10-year |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | 87 | 88 | 89 | 90 | 91 | 92 | 93 | 94 | 95 | 96 | annual variations | mean |
| Soil moisture | 0.00 | -0.07 | -0.11 | 0.17 | -0.01 | 0.04 | -0.01 | -0.04 | -0.11 | 0.09 | 0.06 | 0.00 |
| Groundwater | 0.00 | -0.31 | 0.06 | 0.26 | -0.06 | 0.13 | 0.04 | -0.13 | -0.13 | 0.06 | 0.12 | -0.01 |
| Snow depth | 0.01 | 0.01 | -0.01 | 0.00 | -0.01 | 0.00 | 0.01 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 |
| Total | 0.01 | -0.37 | -0.06 | 0.43 | -0.08 | 0.17 | 0.03 | -0.17 | -0.23 | 0.15 | 0.17 | -0.01 |

*Table 5: Long-term variations of terrestrial water storage in Illinois (1987-1996).*

measured streamflow, comparing the present results with other water-balance studies investigating some of the domains considered here.

### 4.1 ERA-40 water vapour flux and flux divergence over the United States

This section gives a short description of the ERA-40 water vapour flux and flux divergence over the United States in annual, winter (DJF), and summer (JJA) means for the years 1987-1996 (see Figure 4). The maps represent means of analysis values sampled 4 times a day.

In the annual mean, the dominant vapour flux occurs over the Gulf Coast states, a region of comparatively large moisture convergence. There is also a substantial vapour flux from the Pacific Ocean to the Northwest. Convergence values dominate over land, while there is significant vapour flux divergence over sea. Note that mean annual divergence values over land are likely to be an artefact of the reanalysis. As one expects, divergence over sea is largest in tropical and subtropical regions.

The winter mean vapour flux presents regions of large convergence, mainly in the Pacific Northwest as well as in the Gulf Coast states. There is a sharp land-sea gradient in vapour flux divergence along the East Coast during this season. Over land, the flux is almost always convergent, while it is divergent over the sea areas, except in the Northwestern Pacific.

The picture is very different in summer, where there is divergence in many regions over land and convergence over sea off the East Coast. Some convergence also appears west of the Mexican Sierra Madres. One of the main features of the summer moisture circulation over the United States is the large vapour flux from the Gulf of Mexico into the central United States associated with the North American Monsoon system. Despite this circulation, most of the central United States exhibits a net export of moisture during the summer season.

On the whole, these features are consistent with what is known about the climate and circulation characteristics of the United States (e.g. Brubaker et al. 2001) and are similar to the patterns of vapour flux divergence described by Roads et al. (1994) in their analysis the NCEP/NCAR (National Center for Environmental Prediction/National Center for Atmospheric Research) analysis data for the years 1984 to 1990.

### 4.2 Computed estimates: 10-year climatology

Figure 5 displays the 10-year mean annual cycle of vapour flux convergence, observed runoff, monthly variations in atmospheric water content, and computed monthly variations in terrestrial water storage for the regions investigated: the Mississippi subbasins Arkansas-Red, Missouri, upper Mississippi and Ohio-Tennessee (domains 1-4), the whole Mississippi river basin (sum of domains 1-5), and Illinois (domain 6).

In most domains, the vapour flux convergence is positive from the fall until the middle of spring. It generally presents two distinctive peaks, one in November-December and the other in spring (in March in the Ohio-Tennessee River basin, in April in the upper Mississippi River basin, and in May in the Missouri River basin). In summer, there is divergence in most domains. The Ohio-Tennessee River basin (situated in the large moisture convergence region of the Gulf Coast, see section 4.1) has unique characteristics: It exhibits large water vapour convergence in most months, and divergence in one month (August) only. This is due to the important orographic effects of Appalachian Mountains which are responsible for the heavy precipitation typical of this region (e.g. Pardé 1930, Berbery and Rasmusson 1999).

Streamflow strongly varies in magnitude between the considered domains. It is comparatively negligible in the Arkansas-Red and Missouri River basins, while it attains very large values in the Ohio-Tennessee River basin

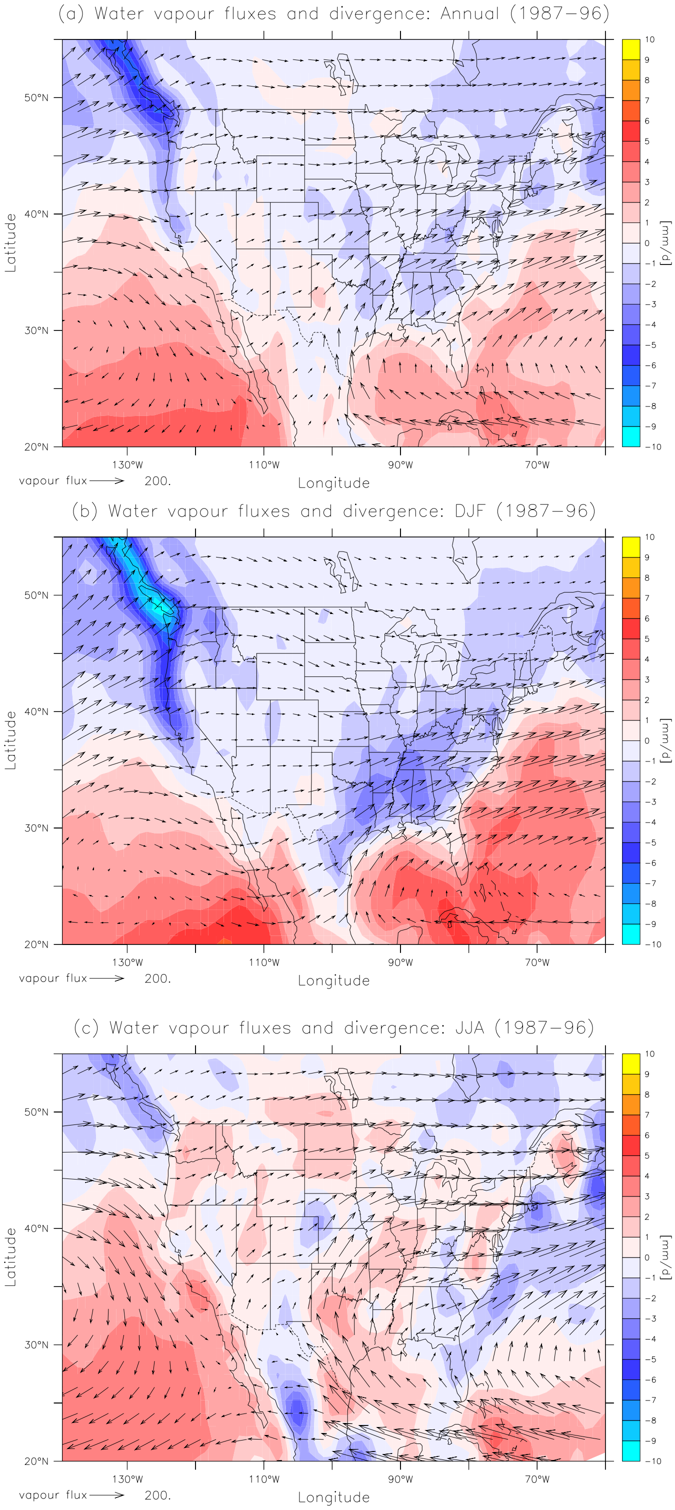


*Figure 4: Water vapour flux [kg m$^{-1}$ s$^{-1}$] and flux divergence [mm/d] over North America in ERA-40: (a) annual mean, (b) winter mean (DJF), and (c) summer mean (JJA) fields for the years 1987-1996. A light smoothing has been applied for display purposes. The arrow scale at the bottom of the plots represents 200 kg m$^{-1}$ s$^{-1}$.*

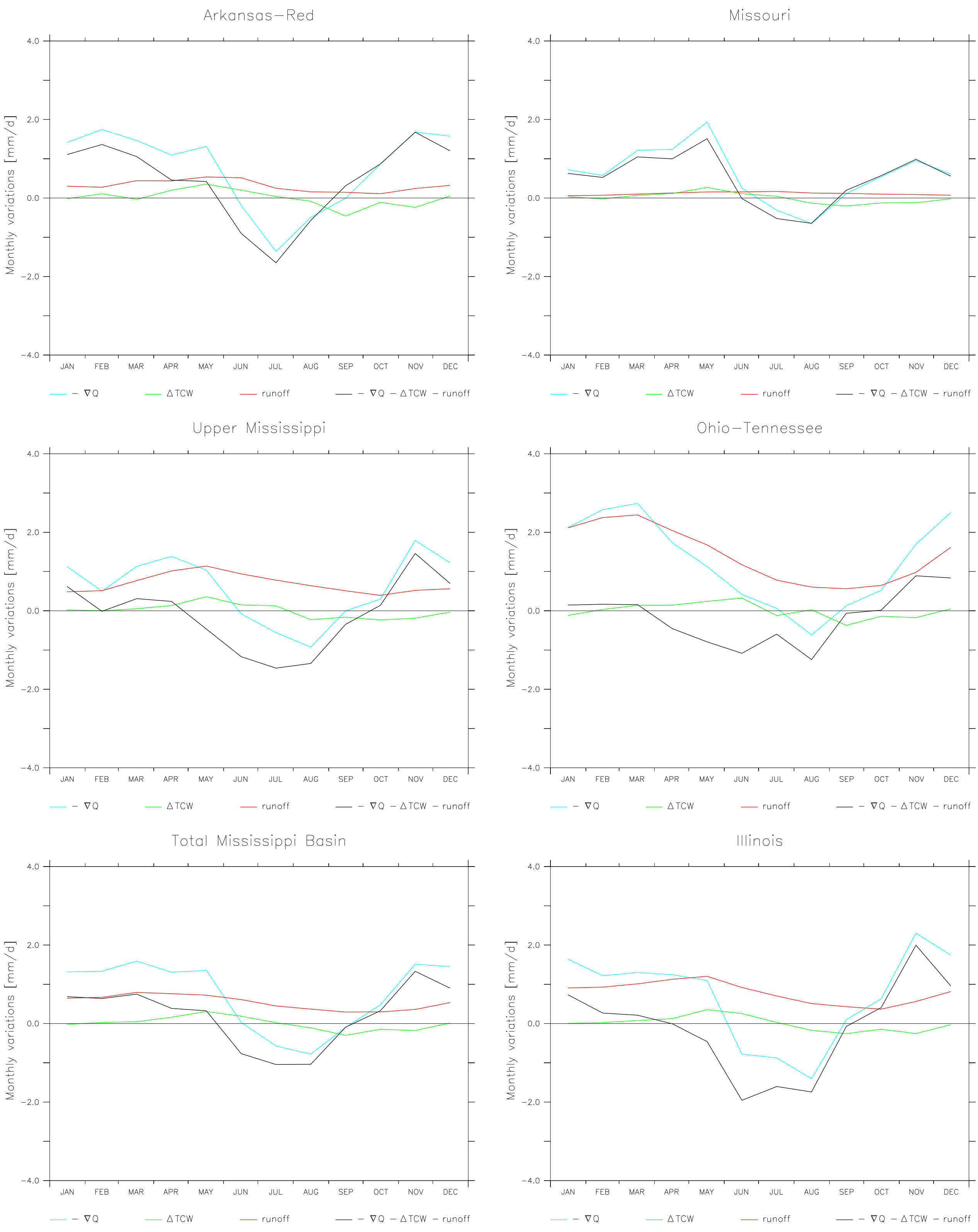


*Figure 5: Ten-year (1987-1996) mean annual cycle [mm/d] of vapour flux convergence (denoted $-\nabla Q$; blue line), of the changes in atmospheric water content (denoted $\Delta TCW$; green line), of runoff (red line), and of the computed changes in terrestrial water storage (black line), for the Arkansas-Red, Missouri, upper Mississippi, and Ohio-Tennessee River basins, the whole Mississippi River basin, and Illinois (see Fig. 1 and Table 1 for the definition of the domains).*

(in relation with the heavy precipitation in this region). It is generally largest in spring, at the time of the peak in vapour flux convergence.

The variations in atmospheric water content are generally smaller than the other water-balance components. Their mean annual cycle is very similar for all 6 domains, and presents a net increase in spring and a decrease in the fall.

The mean annual cycle of the computed monthly variations in terrestrial water storage (the residual of the other three variables) is relatively similar for the 6 domains considered. Storage depletion occurs in spring and summer (generally from April-May to September), with a peak between June and August. In most domains, there is a recharge peak in November, associated with the peak in vapour flux convergence occurring during this month. The Missouri River basin also presents an additional recharge peak in spring, which again closely follows the peak in vapour flux convergence. This is probably a realistic feature as it is known that a large part of the recharge occurs as snow accumulation in this region (e.g. Pardé 1930). In the other domains, the spring peak in vapour convergence only partly goes into terrestrial water recharge, the remaining going into runoff.

### 4.3 Computed estimates: Interannual variability

Figure 6 displays the temporal evolution of the computed monthly variations in terrestrial water storage for the investigated regions. As for the climatological means, the estimates generally follow the vapour flux convergence, while runoff and the changes in atmospheric water content are usually comparatively small. Runoff, however, can be significant in years with large vapour flux convergence (for instance in 1993 in the upper Mississippi River basin and Illinois), and is especially large in the Ohio-Tennessee River basin, where it amounts to similar values as the vapour flux convergence (see also previous section).

Note that, contrary to what is seen in the other domains, there is almost no lag-correlation between vapour flux convergence and the observed river runoff in the Ohio-Tennessee River basin. This subbasin has, in general, much shallower soils and much more terrain variation than the other Mississippi subbasins; these two characteristics are responsible for the shorter residence time of soil water, and for the comparatively large surface (Hortonian) flow observed in this region (P.A. Dirmeyer, personal communication, 2003). As most of the vapour flux convergence goes into river runoff, the computed changes in terrestrial water storage are a mere residual of two large values and may be rather inaccurate for this subbasin.

In the other domains, the temporal evolution of the estimates is generally characterized by a clear seasonal cycle, with storage depletion in spring and summer, and recharge during the rest of the year. Note that individual events such as the late spring 1988 drought and the summer 1993 flood in the Midwest are clearly captured in the regions concerned (Upper Mississippi, Illinois).

### 4.4 Computed estimates: Long-term imbalances

Though the mean vapour flux convergence should equate areal runoff for long-term averages (see equation 8, section 2.1), this is not the case for the investigated domains (Table 6), and correspondingly the 10-year average of the computed variations in terrestrial water storage do not cancel out (contrary to the variations in atmospheric water content). Such imbalances are a common problem of studies investigating atmospheric water balances, both with raw radiosondes data (e.g. Rasmusson 1968, Ropelewski and Yarosh 1998, Yarosh et al. 1999) and modelling products (e.g. Roads et al. 1994, Oki et al. 1995, Gutowski et al. 1997). They are likely due to errors in the determination of the regional vapour flux divergence, as streamflow measurements are known to be accurate within a few percent (e.g. Gutowski et al. 1997). An open issue is the extent to which

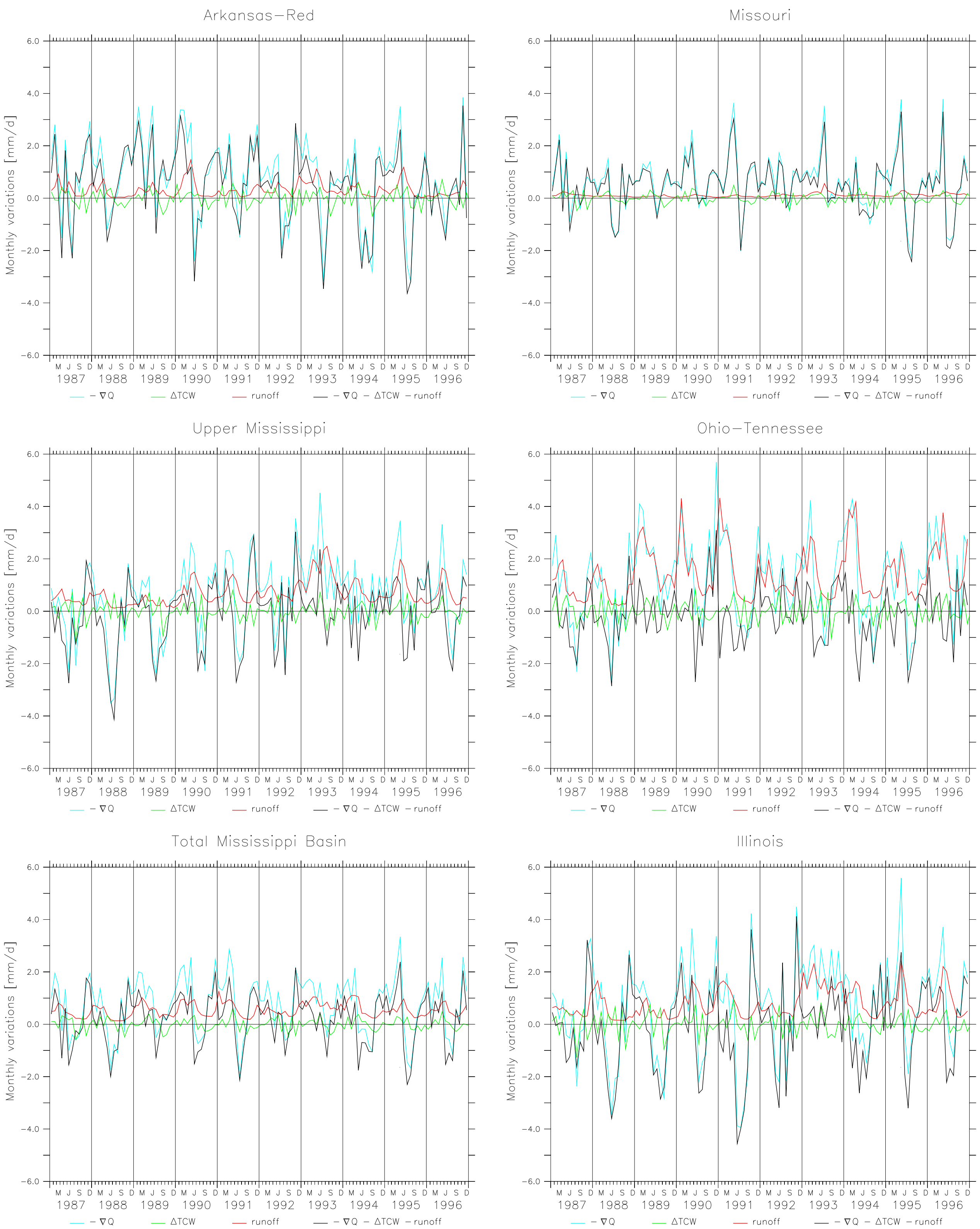


*Figure 6: Monthly water-balance components for the Arkansas-Red, Missouri, upper Mississippi, Ohio-Tennessee, and whole Mississippi River basins, and for Illinois [mm/d]: vapour flux convergence (denoted $-\nabla Q$; blue line), changes in atmospheric water content (denoted $\Delta TCW$; green line), runoff (red line), and computed estimates of changes in terrestrial water storage (black line).*

| | ERA-40 present study (87-96) | | | | NCEP G97(84-93)*, Y98(83-94)** | | | EDAS BR99 (05/95-04/97) | | |
|---|---|---|---|---|---|---|---|---|---|---|
| | $-\overline{divQ}$ | $\overline{R}$ | $\overline{\frac{\partial W}{\partial t}}$ | $\overline{\frac{\partial S}{\partial t}}$ | $-\overline{divQ}$ | $\overline{R}$ | $Imb^1$ | $-\overline{divQ}$ | $\overline{R}$ | $Imb^1$ |
| Arkansas-Red | 0.76 | 0.31 | 0.002 | 0.44 | - | - | - | -1.07 | 0.30 | -1.37 |
| Missouri | 0.60 | 0.11 | 0.002 | 0.49 | - | - | - | -0.18 | 0.16 | -0.34 |
| Upper Mississippi | 0.58 | 0.69 | 0.000 | -0.11 | 1.0 | 0.72 | 0.28 * | 0.41 | 0.76 | -0.35 |
| Ohio-Tennessee | 1.30 | 1.42 | 0.003 | -0.17 | 0.89 | 1.33 | -0.44* | 2.11 | 1.74 | 0.37 |
| Whole Mississippi | 0.74 | 0.54 | 0.002 | 0.20 | - | - | - | - | - | - |
| Illinois | 0.69 | 0.79 | 0.001 | -0.11 | 0.80 | 0.86 | -0.06** | - | - | - |

[1] Imbalance ($-\overline{divQ} - \overline{R}$): This term corresponds to $\overline{\frac{\partial S}{\partial t}} + \overline{\frac{\partial W}{\partial t}}$ in the first column.

*Table 6: Ten-year (1987-1996) average values of vapour flux convergence, runoff, changes in atmospheric water content, and changes in terrestrial water storage of the present study (left column), and long-term averages of vapour flux convergence, runoff, and corresponding imbalances in the studies of Gutowski et al. 1997 (G97), Yeh et al. 1998 (Y98), and Berbery and Rasmusson 1999 (BR99). All values in [mm/d].*

the non-accounting of the groundwater fluxes at the domains' lateral boundaries might be responsible for some of the imbalance, although this effect is likely to be small in comparison (see also Yarosh et al. 1999).

Results from other studies using analyses for the computation of vapour flux convergence in some of the domains investigated are presented in Table 6. Gutowski et al. (1997) and Yeh et al. (1998) use NCEP reanalysis data, Berbery and Rasmusson (1999) Eta Data Assimilation System (EDAS) analyses. In the case of Berbery and Rasmusson (1999), we computed the mean streamflow for the relevant time period from the USGS dataset, as comparisons with streamflow data were not provided in their study. Note, moreover, that our domain 2 "Missouri" corresponds to the sum of their domains "Missouri" and "Lower Missouri", and that we averaged the data of these two domains for the present comparison.

From the comparison with these three studies, it is apparent that the long-term imbalances between the ERA-40 vapour flux convergence and the measured streamflow is comparable as (and in general smaller than) for other analysis products. Note, however, that the results of Berbery and Rasmusson (1999) are available for only two years of data, which might explain the large imbalances between the vapour flux convergence and streamflow in their study.

It is difficult to identify the exact causes for the imbalances between the reanalysis data and the measured streamflow. Possible explanations for biases in the computed monthly ERA-40 vapour flux convergence could be related to: 1) Insufficient spatial and temporal sampling of the radiosonde measurements used in the assimilation; 2) possible errors in the radiosonde measurements themselves; 3) lack of water conservation due to the analysis increments; 4) general deficiencies in the data assimilation procedure, especially with regards to humidity and a balance between the humidity and the vertical motion (Holm et al. 2002); 5) systematic biases in the model representation of the atmospheric and surface branches of the hydrological cycle (Holm et al. 2002); and 6) insufficient temporal resolution of the reanalysis (6-hour data). For atmospheric water budgets based on raw radiosonde data, biases appear mainly linked to the temporal resolution of the employed measurements (e.g. Yarosh et al. 1999). In the case of reanalysis data, this aspect could possibly also be of relevance, as the 6-hour data might not be sufficient to resolve the diurnal cycle of the moisture fluxes.

For the whole Mississippi basin, the total imbalance is relatively small due to a compensation between the positive and negative imbalances of the various subbasins. Such compensations between regional biases could explain why atmospheric water-balance computations for larger domains tend to be more accurate as discussed

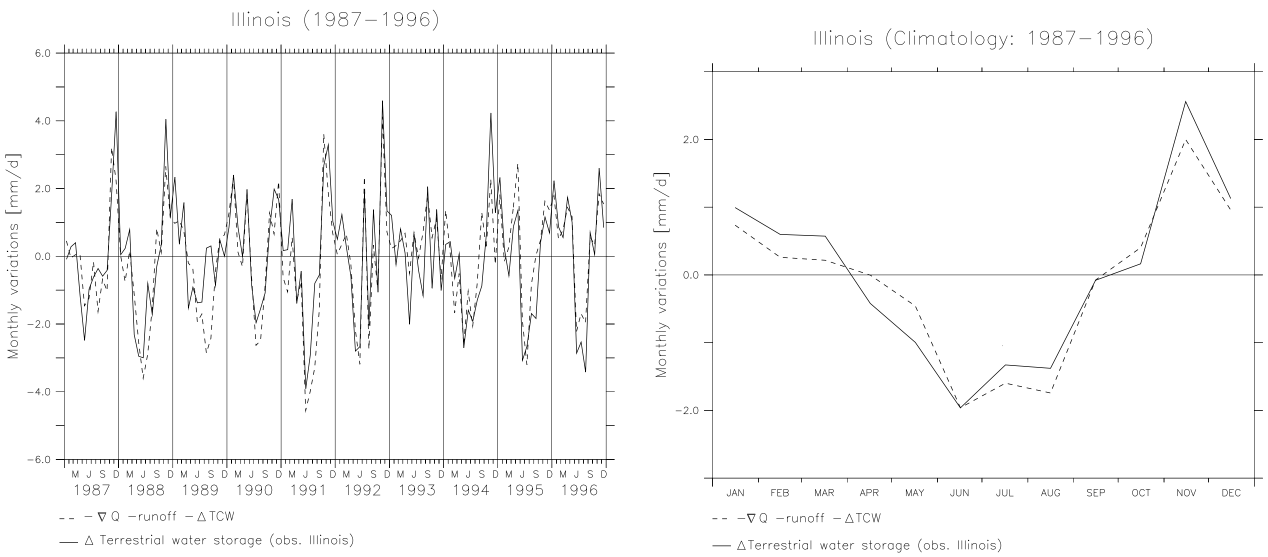


*Figure 7: Computed and observed variations in terrestrial water storage in Illinois [mm/d].*

in Rasmusson (1968, 1971; see also section 2.1). Note, however, that the imbalance is small for Illinois as well, despite the very small size of this domain. In the NCEP reanalysis, the imbalance in Illinois is even lower, possibly due to a similar compensation between negative and positive imbalances, as the NCEP reanalysis imbalances switch sign in this region (W.J. Gutowski, personal communication, 2002).

Preliminary results for other river basins suggest that the magnitude of the water budget imbalances vary geographically, possibly linked with regional climate characteristics (see Seneviratne 2003, chapter 4). In the case of the Mississippi River basin, the vapour flux convergence appears to be overestimated in the western domains (Arkansas-Red, Missouri), and underestimated in the eastern domains (Upper Mississippi, Ohio-Tennessee, Illinois). Interestingly, Betts et al. (2003) found a similar subdivision of the Mississippi subbasins relative to precipitation forecasts in ERA-40, with a tendency for overestimation of summer precipitation in the east (and slight underestimation in the west). It is possible that these biases are related, but a more detailed analysis would be needed in order to investigate such links.

# 5 Results: Validation against observations in Illinois

## 5.1 Monthly estimates

Figure 7 compares the water-balance estimates of the terrestrial water-storage variations in Illinois with the available observations, for the ten years investigated (top) and their climatology (bottom). Overall, the estimates show an excellent agreement with the observations, with the only exception of 1989. Important features such as the late spring drought of 1988 and the summer flood of 1993 are well captured (as is the drought of 1991), a sign that the interannual variability of the vapour flux convergence is well represented in ERA-40 (as seen in section 4.3). Note that the mean climatology of the terrestrial water-storage variations for the ten years investigated is also well captured.

Figure 8 shows a scatter diagram of the computed estimates of monthly terrestrial water-storage variations versus the observational values in Illinois. The correlation between both time series is equal to 0.84, with a

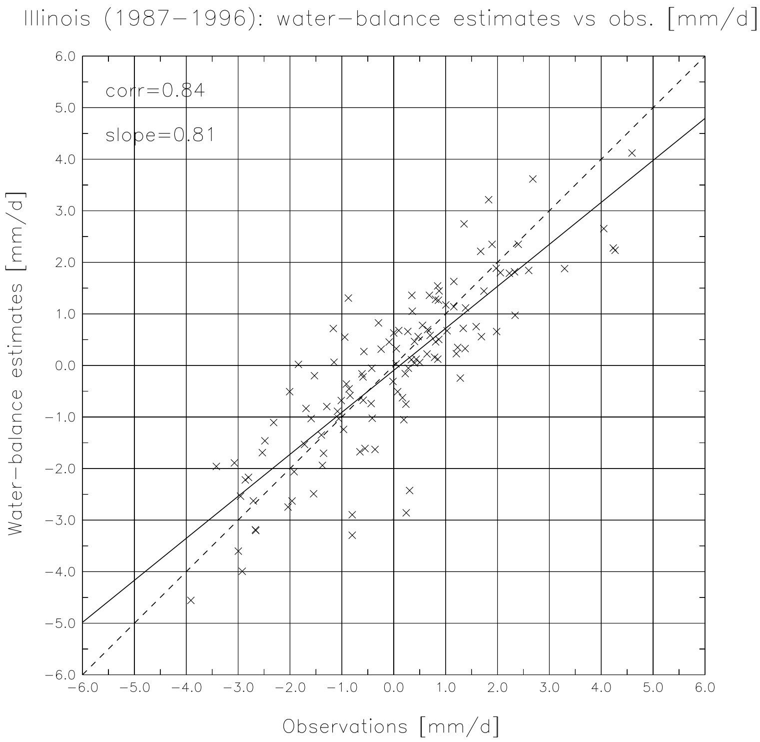


*Figure 8: Scatter diagram of the computed and observed monthly variations in terrestrial water storage in Illinois [mm/d].*

slope of 0.81, i.e. a slight tendency to underestimate observed changes. Note that this can be explained in part by the fact that the observations are discrete points which experience greater extremes than the grid areas of the model (heavier local downpours, longer stretches with no rain). Although the agreement between the monthly water-balance estimates and the observations is not perfect, the clear correlation between them suggests that relatively accurate estimates of monthly terrestrial water-storage variations could be obtained using the tested methodology with the ERA-40 reanalysis data.

Note that the year displaying the poorest agreement between the estimates and the observations is 1989, which was identified by Bosilovich and Schubert (2001) as having the largest ratio of precipitation recycling in a 15-year sample (1981-1995) due to a combination of low moisture transport and high moisture convergence. This might be an indication that processes related to such climatic conditions are not accurately captured by the ERA-40 reanalysis.

## 5.2 Seasonal changes in terrestrial water storage

An important issue is the possibility to obtain estimates of the yearly amplitude of terrestrial water storage, as there is little consistent information on this critical hydrological quantity. Heck et al. (2001) found for instance that the amplitude of the soil moisture cycle between April and September can be as disparate as $\sim$100 mm and $\sim$300 mm in Europe (average for Spain, years 1987-1992), depending on the dataset considered (ERA-15 and NCEP reanalysis, respectively). Similar inconsistencies are also found in the United States (e.g. Seneviratne 2003, chapter 4).

Table 7 displays the mean observed and estimated seasonal changes in terrestrial water storage in Illinois for

| | observations | estimates | | | |
|---|---|---|---|---|---|
| | change [mm] | change [mm] | bias [mm] | bias [%] | corr |
| Aug1-Apr1 | -144 | -123 | +21 | -14.4 | 0.94 |
| Sep1-Apr1 | -186 | -177 | +9 | -5.0 | 0.83 |
| Oct1-Apr1 | -189 | -179 | +10 | -5.1 | 0.74 |
| Nov1-Apr1 | -184 | -167 | +17 | -9.3 | 0.69 |

*Table 7: Computed mean (1987-1996) seasonal change in terrestrial water storage vs. observations for various time ranges [mm] in Illinois.*

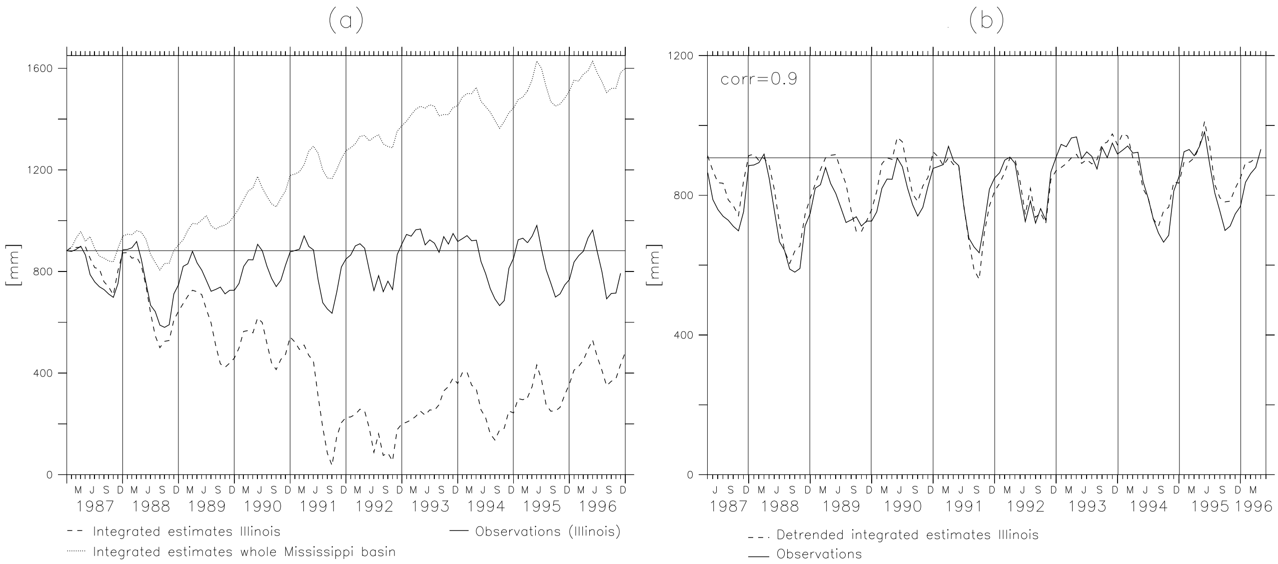


*Figure 9: (a) Integrated estimates of terrestrial water-storage variations for Illinois and the whole Mississippi River basin compared against observations in Illinois [mm]; (b) Detrended integrated estimates for Illinois compared against observations [mm].*

various time ranges. Interestingly, the water-balance estimates show a good correlation with the observations for periods of 4-5 months (from April 1 to August 1/September 1). For longer time ranges, the correlation is lower, although the 10-year mean estimated storage depletion is generally close to the observations (but always slightly underestimated).

## 5.3 Long-term integration of the estimates

Figure 9a displays the integrated monthly estimates of terrestrial water-storage variations for Illinois and the whole Mississippi River basin, compared with the absolute values of the observations, for the whole 10-year period. A striking feature is the drift of the integrated estimates that starts around 1989 for both regions (with opposite sign), and which can be directly related to the water-balance imbalances discussed in section 4.4.

In order to correct for such drifts, the detrending procedure that is usually applied consists of a uniform monthly correction factor (e.g. Rasmusson 1968, Oki et al. 1995, Ropelewski and Yarosh 1998). As is apparent from Figure 9a, such a correction would not correctly suppress the drift, as some years present more drift than others, and most of all due to the fact that the drift changes sign during the considered period over Illinois.

Therefore, we correct the annual drift separately for each year, under the assumption of unchanged terrestrial water storage for annual means. In order to minimize the error induced by this approximation (see also section 3.4 and Table 5), we chose April as reference month, since it exhibits the smallest spread in yearly values of observed terrestrial water storage (Table 4). The constant of integration (mean terrestrial water storage in April) is taken from the 10-year climatology of the observations. Note that this method could also be applied for inferring storage changes in years where no observations are available, under the assumption that the mean climatological conditions of the observational period is representative for the investigated time frame.

Figure 9b displays the integrated values of the so-obtained detrended estimates for Illinois compared with the observations. In general, the detrended integrated estimates exhibit a very good agreement with the observations (correlation of 0.9), despite the relative simplicity of the applied detrending.

## 6 Soil moisture simulation in ERA-40

For comparison purposes, this section briefly describes the ERA-40 analysis fields of soil moisture in Illinois. The model used in the reanalysis includes the land-surface scheme described in Van den Hurk et al. (2000). An optimal interpolation procedure links analysis increments of screen-level temperature and humidity to increments of soil moisture in the top 3-layers, spanning the top 1-m of soil (Douville et al. 2000). Physically, the initial soil moisture values will change the latent and surface heat fluxes in order to minimize the mismatch between observed and background screen-level parameters. The ERA-40 soil moisture is available at 4 levels: from 0 to 7 cm, from 7 to 28 cm, from 28 to 100 cm, and from 100 to 289 cm. As the soil moisture observations are available for a 2-m soil layer only, we scaled the soil moisture content of the lowermost layer to a depth of 1 m and summed this value together with the moisture content of the other three layers for the present comparisons. In the remaining part of the section, it is important to keep in mind that the quantities that should be compared are changes in soil moisture, rather than absolute values (see also Koster and Milly 1997).

The monthly variations in soil moisture show good agreement with the observations (Figures 10a and 10b). The interannual variability is in general well captured, however, the reanalysis underestimates both soil moisture depletion in summer and soil moisture recharge in the fall (Fig. 10c), leading to a damping of the annual cycle (see hereafter). The temporal evolution of the absolute soil moisture values (Fig. 10d) shows again a good agreement with the observations (with a relatively high correlation, Fig. 10e), while no real physical meaning can be attributed to the mismatch between the mean values of soil moisture (e.g. Koster and Milly 1997, see above comment). Note, however, that the reanalysis underestimates the mean amplitude of the annual soil moisture changes: Between April and October the net decrease in soil moisture amounts to $\sim$50 mm in ERA-40, compared to $\sim$100 mm for the observations (Fig. 10f).

In summary, the ERA-40 soil moisture appears to be consistent with the observations, presents relatively accurate monthly variations, but has a damped yearly cycle. The damping of the yearly cycle is likely attributable to the increments in soil water, which systematically supply water in summer and remove water in winter and early spring (Betts et al. 2003).

Note that the ERA-40 monthly variations in soil moisture do not correlate as well with the observations as the water-balance estimates of terrestrial water-storage variations discussed in the preceding sections. This illustrates the advantage of using vapour flux divergence for the computation of water budgets at the surface, as this quantity is likely to be less dependent on the reanalysis model than other variables, thanks to the assimilation of the radiosonde data. In Illinois, groundwater and soil moisture are clearly correlated and snow amounts are small (see section 3.4, and Figs. 2 and 3), so that it should be possible to scale the estimated changes in terrestrial water storage to obtain information on soil moisture for this region. This might apply to other regions with similar climatic conditions.

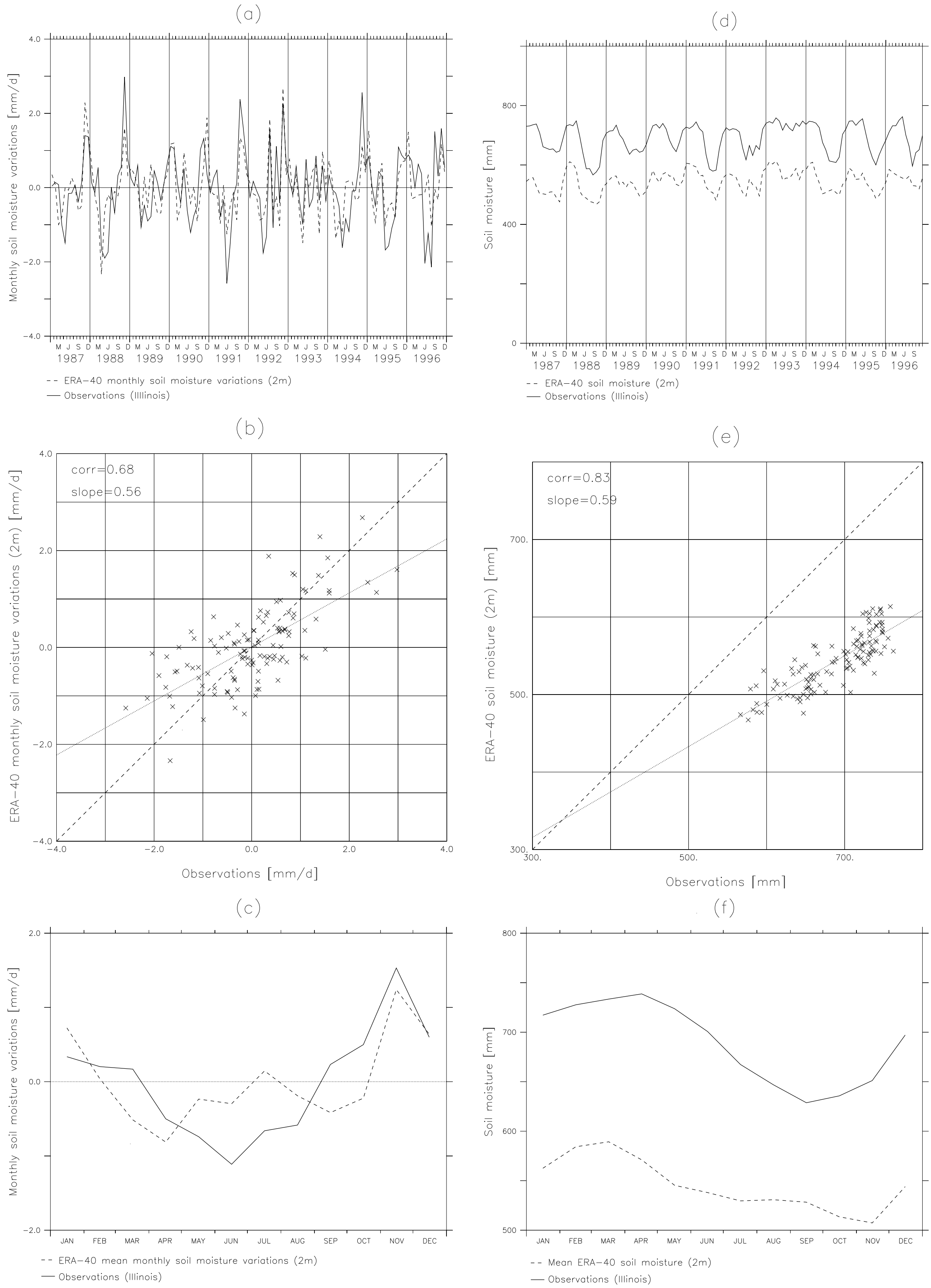


*Figure 10: Monthly variations (left-hand panels, [mm/d]) and values (right-hand panels, [mm]) of ERA-40 soil moisture (scaled to 2 m) over Illinois compared against observations: (a,d) temporal evolution, (b,e) scatter plot, and (c,f) mean climatology.*

## 7 Summary and conclusions

This study investigates the feasibility of estimating monthly variations in terrestrial water storage from water-balance computations, using atmospheric water vapour convergence from the ERA-40 reanalysis data and conventional runoff data. The results are very promising, as the computed estimates of monthly terrestrial water-storage variations appear realistic for the investigated domains within the Mississippi River basin and show very good agreement with observations in Illinois. The mean seasonal cycle is well represented for the studied period and the interannual variability is in general well captured.

In the long-term average, the computed variations in terrestrial water storage do not cancel out due to imbalances between the ERA-40 vapour flux convergence and the measured streamflow, which are likely due to systematic biases in the reanalysis data. Because of these biases, an estimation of the temporal evolution of terrestrial water storage over time periods longer than 4-6 months is possible only with an appropriate detrending. The simple detrending procedure applied here allows a good estimation of absolute terrestrial water storage in most years.

An important result is that the critical domain size for water-balance computations using high resolution reanalysis data appears to be much smaller than for raw radiosonde data (e.g. Rasmusson 1968, 1971). The Illinois domain has a size of only $\sim 2\text{x}10^5$ km$^2$ and is shown to be suitable for the computation of the water-balance estimates. Yeh et al. (1998) come to similar conclusions in their study of the hydroclimatology of Illinois.

One should note that Illinois is a region with flat and homogeneous terrain, and that this methodology might not be as accurate for areas presenting more horizontal and vertical heterogeneities, or other climatic characteristics. We are not aware of other regions with such comprehensive observational datasets of terrestrial water storage as Illinois, and it might thus not be possible to perform such a detailed validation for other river basins, but further validation studies for regions with observations of at least some components of the terrestrial water storage would be desirable. Moreover, it would also be useful to perform cross-comparisons with estimates obtained with other methodologies, such as the model-computed soil moisture from the phase 2 of the GSWP project (see Dirmeyer et al. 1999), or remote sensing measurements of the GRACE mission (Wahr et al. 1998), which will provide estimates of changes in terrestrial water storage for the entire globe.

Finally, the ERA-40 reanalysis project has recently been completed, hence allowing the computation of 45-year timeseries of the proposed estimates for various regions of the world. A study for several river basins of the northern mid-latitudes is currently underway (Hirschi 2002; Seneviratne 2003, chapter 4) and could possibly be extended to other regions. The creation of such a dataset would constitute a useful contribution in view of the scarcity of observations of terrestrial water storage and its components.

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## References

Alestalo, M., 1983: The atmospheric water vapour budget over Europe. In: *Variations in the Global Water Budget*, A. Street-Perrot et al. (eds.), pp. 67-79. D. Reidel Publishing Company.

Basara J.B., and T.M. Crawford, 2000: Improved installation procedures for deep-layer soil moisture measurements. *J. Atm. Ocean. Tech.*, **17 (6)**, 879-884.

Beljaars, A.C.M, P. Viterbo, M. J. Miller, and A.K. Betts, 1996: The anomalous rainfall over the United States during July 1993: Sensitivity to land surface parameterization and soil moisture anomalies. *Mon. Wea. Rev.*, **124**, 362-383.

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

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

Berbery, E.H., and E.M. Rasmusson, 1999: Mississippi moisture budgets on regional scale. *Mon. Wea. Rev.*, **127**, 2654-2673.

Betts, A.K., J.H. Ball, A.C.M. Beljaars, M.J. Miller, and P.A. Viterbo, 1996: The land-surface atmosphere interaction: A review based on observational and global modeling perspectives. *J. Geophys. Res.*, **101D**, 7209-7225.

Betts, A.K., P. Viterbo, and E. Wood, 1998: Surface energy and water balance for the Arkansas-Red River basin from the ECMWF reanalysis. *J. Climate*, **11**, 2881-2897.

Betts, A.K., J.H. Ball, and P. Viterbo, 1999: Basin-scale surface water and energy budgets for the Mississippi from the ECMWF reanalysis. *J. Geophys. Res.*, **104D**, 19,293-19,306.

Betts, A.K., J.H. Ball, M. Bosilovich, P. Viterbo, Y.-C. Zhang, and W.B. Rossow, 2003: Intercomparison of water and energy budgets for five Mississippi Sub-basins between ECMWF reanalysis (ERA-40) and NASA-DAOfvGCM for 1990-1999. *J. Geophys. Res.*, **108D**, 8618, doi:10.1029/2002JD003127.

Bosilovich, M.G., and S.D. Schubert, 2001: Precipitation recycling over the central United States diagnosed from the GEOS-1 data assimilation system. *J. Hydrometeor.*, **2**, 26-35.

Brubaker, K.L., P.A. Dirmeyer, A. Sudradjat, B. Levy, and F. Bernal, 2001: A 36-year climatology of the evaporative sources of warm-season precipitation in the Mississippi River Basin. *J. Hydrometeor.*, **2**, 537-557.

Calvet, J.C., and J. Noilhan, 2000: From near-surface to root-zone soil moisture using year-round data. *J. Hydrometeor.*, **1**, 393-411.

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

Coughlan, M., and R. Avissar, 1996: The Global Energy and Water Cycle Experiment (GEWEX) Continental-Scale International Project (GCIP): An overview. *J. Geophys. Res.*, **101D**, 7139-7147.

Courtier, P., E. Andersson, W. Heckley, J. Pailleux, D. Vasiljevic, M. Hamrud, A. Hollingsworth, F. Rabier, and M. Fisher, 1998: The ECMWF implementation of three dimensional variational assimilation (3D-Var). Part I: Formulation. *Quart. J. Roy. Meteor. Soc.*, **124**, 1783-1808.

Dirmeyer, P.A., and K.L. Brubaker, 1999: Contrasting evaporative moisture sources during the drought of 1988 and the flood of 1993. *J. Geophys. Res.*, **104**, 19,383-19,397.

Dirmeyer, P.A., A.J. Dolman, and N. Sato, 1999: The pilot phase of the Global Soil Wetness Project. *Bull. Am. Met. Soc.*, **80**, 851-875.

Domenico, P.A., and F.W. Schwartz, 1990: *Physical and chemical hydrogeology*, 2nd ed. John Wiley and Sons, Inc., New York, 1990.

Douville, H., P. Viterbo, J.-F. Mahfouf, and A.C.M. Beljaars, 2000: Evaluation of optimal interpolation and nudging techniques for soil moisture analysis using FIFE data. *Mon. Wea. Rev.*, **128**, 1733-1756.

Entin, J.K., A. Robock, K.Y. Vinnikov, V. Zabelin, S. Liu, A. Namkhai, and Ts. Adyasuren, 1999: Evaluation of Global Soil Wetness Project soil moisture simulations. *J. Met. Soc. of Japan*, **77**, 183-198.

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

Fukutome, S., C. Prim, and C. Schär, 2001: The role of the soil in medium-range weather predictability. *Nonlinear Proc. Geoph.*, **8**, 373-386.

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

Heck P., Lüthi D., Wernli H., Schär C., 2001: Climate impacts of European-scale anthropogenic vegetation changes: A sensitivity study using a regional climate model. *J. Geophys. Res.*, **106D**, 7817-7835.

Hirschi, M., 2002: Abschätzung der terrestrischen Wasserspeicheränderung mit Wasserbilanzen und ERA-40 Reanalysedaten in Einzugsgebieten von Flüssen in Asien und Europa ("Estimation of changes in terrestrial water storage with water balance computations and ERA-40 reanalysis data in river basins of Asia and Europe"). Diploma thesis, ETH Zurich, Switzerland, 65 pp. [Available from the Institute for Atmospheric and Climate Science, ETH Zurich, Winterthurerstrasse 190, 8057 Zurich, Switzerland]

Hollinger, S.E. and S.A. Isard, 1994: A soil moisture climatology of Illinois. *J. Climate*, **4**, 822-833.

Holm, E., E. Andersson, A. Beljaars, P. Lopez, J.-F. Mahfouf, A.J. Simmons, and J.-N. Thepaut, 2002: Assimilation and modelling of the hydrological cycle: ECMWF's status and plans. ECMWF Tech. Mem. 383, 55 pp.

Hortal, M., and A.J. Simmons, 1991: Use of reduced Gaussian grids in spectral models. *Mon. Wea. Rev.*, **119**, 1057-1074.

Jackson, T.J., D.M. Le Vine, A.Y. Hsu, A. Oldak, P.J. Starks, C.T. Swift, J.D. Isham, and M. Haken, 1999: Soil moisture mapping at regional scale using microwave radiometry: The southern great plains hydrology experiment. *IEEE transactions on geoscience and remote sensing*, **37**, 2136-2151.

Koster, R.D., and P.C.D. Milly, 1997: The interplay between transpiration and runoff formulations in land surface schemes used with atmospheric models. *J. Climate*, **10**, 1578-1591.

Koster, R.D., M.J. Suarez, and M. Heiser, 2000: Variance and predictability of precipitation at seasonal-to-interannual timescales. *J. Hydrometeor.*, **1**, 26-46.

Masuda, K., Y. Hashimoto, H. Matsuyama, and T. Oki, 2001: Seasonal cycle of water storage in major river basins of the world. *Geophys. Res. Lett.*, **28**, 3215-3218.

Matsuyama, H., and K. Masuda, 1997: Estimates of continental-scale soil wetness and comparison with the soil moisture data of Mintz and Serafini. *Clim. Dyn.*, **13**, 681-689.

Milly, P.C.D., and K.A. Dunne, 1994: Sensitivity of the global water cycle to the water-holding capacity of the land. *J. Climate*, **7**, 506-526.

Oki, T., K. Musiake, H. Matsuyama, and K. Masuda, 1995: Global atmospheric water balance and runoff from large river basins. *Hydrological Processes*, **9**, 655-678.

Oki, T, 1999: The global water cycle. In: *Global Energy and Water Cycle*. K. Browning and R. Gurney (eds.), pp. 10-27. Cambridge University Press.

Pardé, M., 1930: Le régime du Mississippi (“The regime of the Mississippi River”). *Revue de Géographie Alpine*, **18**.

Peixoto, J.P., and A.H. Oort, 1992: *Physics of climate*. Am. Inst. of Phys. Press, Woodbury, New York.

Pomeroy, J.W., and D.M. Gray, 1995: *Snow cover - Accumulation, relocation and management*. National Hydrology Research Institute Science Report No. 7, Saskatoon, Canada, 144 pp.

Rango, A., 1996: Spaceborne remote sensing for snow hydrology applications. *Hydrological Sciences Journal*, **41**, 477-494.

Rasmusson, E.M., 1968: Atmospheric water vapor tranport and the water balance of North America. II. Large-scale water balance investigations. *Mon. Wea. Rev.*, **96**, 720-734.

Rasmusson, E.M., 1971: A study of the hydrology of eastern North America using atmospheric vapor flux data. *Mon. Wea. Rev.*, **99**, 119-135.

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

Roads, J., R. Lawford, E. Bainto, E. Berbery, S. Chen, B. Fekete, K. Gallo, A. Grundstein, W. Higgins, M. Kanamitsu, W. Krajewski, V. Lakshmi, D. Leathers, D. Lettenmaier, L. Luo, E. Maurer, T. Meyers, D. Miller, K. Mitchell, T. Mote, R. Pinker, T. Reichler, D. Robinson, A. Robock, J. Smith, G. Srinivasan, K. Verdin, K. Vinnikov, T. Vonder Haar, C. Vrsmarty, S. Williams, and E. Yarosh, 2003: GCIP water and energy budget synthesis (WEBS). *J. Geophys. Res.*, **108D**, 8609, doi:10.1029/2002JD002583.

Robock, A., K.Y. Vinnikov, G. Srinivasan, J.K. Entin, S.E. Hollinger, N.A. Speranskaya, S. Liu, and A. Namkhai, 2000: The Global Soil Moisture Data Bank. *Bull. Am. Meteorol. Soc.*, **81**, 1281-1299.
Web site: http://climate.envsci.rutgers.edu/soil_moisture/ .

Rodell, M., and J.S. Famiglietti, 2001: An analysis of terrestrial water storage variations in Illinois with implications for the Gravity Recovery and Climate Experiment (GRACE). *Wat. Res. Res.*, **37**, 1327-1339.

Ropelewski, C.F., and E.S. Yarosh, 1998: The observed mean annual cycle of moisture budgets over the central United States (1973-92). *J. Climate*, **11**, 2180-2190.

Salvucci, G.D., 1997: Soil and moisture independent estimation of stage-two evaporation from potential evaporation and albedo or surface temperature. *Wat. Res. Res.*, **33**, 111-122.

Schär, C., D. Lüthi, U. Beyerle, and E. Heise, 1999: The soil-precipitation feedback: A process study with a regional climate model. *J. Climate*, **12**, 722-741.

Seneviratne, S.I., J.S. Pal, E.A.B. Eltahir, and C. Schär, 2002: Summer dryness in a warmer climate: A process study with a regional climate model. *Clim. Dyn.*, **20**, 69-85.

Seneviratne, S.I., 2003: Terrestrial water storage: A critical variable for mid-latitude climate and climate change. PhD dissertation, ETH Zurich, Switzerland, 155 pp. [Diss. ETH No. 14944. Available from the Institute for Atmospheric and Climate Science, ETH Zurich, Winterthurerstrasse 190, 8057 Zurich, Switzerland]

Shukla, J., and Y. Mintz, 1982: Influence of land-surface evaporation on the Earth's climate. *Science*, **215**, 1498-1501.

Simmons, A.J., and J.K. Gibson, 2000: *The ERA-40 Project Plan*. ERA-40 Project Report Series, No. 1, 62 pp. Available from ECMWF.

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

Trenberth, K.E., 1995: Truncation and use of model coordinate data. *Tellus*, **47A**, 287-303.

Trenberth, K.E., and C.J. Guillemot, 1995: Evaluation of the global atmospheric moisture budget as seen from analyses. *J. Climate*, **8**, 2255-2272.

Trenberth, K.E., J.W. Hurrell, and A. Solomon, 1995: Conservation of mass in three dimensions in global analysis. *J. Climate*, **8**, 692-708.

Trenberth, K.E., 1997: Using atmospheric budgets as a constraint on surface fluxes. *J. Climate*, **10**, 2796-2809.

Van den Hurk, B.J.J.M., P. Viterbo, A.C.M. Beljaars, and A.K. Betts, 2000: Offline validation of the ERA-40 surface scheme. *ECMWF Tech Memo*, **295**, 43 pp. European Centre for Medium-Range Weather Forecasts, Shinfield Park, Reading RG2 9AX, England, UK.

Viterbo, P., and A.K. Betts, 1999: Impact of the ECMWF reanalysis soil water forecasts of the July 1993 Mississippi flood. *J. Geophys. Res.*, **104D**, 19361-19366.

Viterbo, P., and A.C.M. Beljaars, 2003: Impact of land surface on weather. Chapter A3.5, In: *Vegetation, Water, Humans and the Climate: A New Perspective on an Interactive System*, M. Claussen et al. (eds.), Springer-Verlag, New York, in press.

Wahr, J., M. Molenaar, and F. Bryan, 1998: Time-variability of the Earth's gravity field: Hydrological and oceanic effects and their possible detection using GRACE. *J. Geophys. Res.*, **103B**, 30,205-30,229.

Wetherald R.T., and S. Manabe, 1999: Detectability of summer dryness caused by greenhouse warming. *Clim. Change*, **43**, 495-511.

Yarosh, E.S., C.F. Ropelewski, and E.H. Berbery, 1999: Biases of the observed atmospheric water budgets over the central United States. *J. Geophys. Res.*, **104D**, 19,349-19,360.

Yeh P.J.F., M. Irizarry, and E.A.B. Eltahir, 1998: Hydroclimatology of Illinois: A comparison of monthly evaporation estimates based on atmospheric water balance and soil water balance. *J. Geophys. Res.*, **103D**, 19,823-19,837.

Zekster, I.S., and H.A. Loaiciga, 1993: Groundwater fluxes in the global hydrological cycle: past, present and future. *J. Hydrology*, **144**, 405-427.