WATER RESOURCES RESEARCH, VOL. 37, NO. 5, PAGES 1327–1339, MAY 2001

An analysis of terrestrial water storage variations in Illinois with implications for the Gravity Recovery and Climate Experiment (GRACE)

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Abstract. Variations in terrestrial water storage affect weather, climate, geophysical phenomena, and life on land, yet observation and understanding of terrestrial water storage are deficient. However, estimates of terrestrial water storage changes soon may be derived from observations of Earth's time-dependent gravity field made by NASA's Gravity Recovery and Climate Experiment (GRACE). Previous studies have evaluated that concept using modeled soil moisture and snow data. This investigation builds upon their results by relying on observations rather than modeled results, by analyzing groundwater and surface water variations as well as snow and soil water variations, and by using a longer time series. Expected uncertainty in GRACE-derived water storage changes are compared to monthly, seasonal, and annual terrestrial water storage changes estimated from observations in Illinois (145,800 km^2). Assuming those changes are representative of larger regions, detectability is possible given a 200,000 km^2 or larger area. Changes in soil moisture are typically the largest component of terrestrial water storage variations, followed by changes in groundwater plus intermediate zone storage.

1. Introduction

Groundwater, soil moisture, snow and ice, lakes and rivers, and water contained in biomass are the principal components of terrestrial water storage. Through an array of simple to complex processes and feedback mechanisms, terrestrial water interacts with other terrestrial and meteorological factors to shape climate and control weather. Soil moisture in particular has been shown to exert a significant influence in general circulation models (GCMs) (see *Entekhabi et al.* [1996] for a review) through its capacity for storing and releasing heat and its control of evapotranspiration. Changes in total terrestrial water storage likely cause or balance sea level variations [*Chen et al.*, 1998] and affect the gravity field and rotation of the Earth [*Chao and O'Connor*, 1988; *Kuehne and Wilson*, 1991]; however, the effects on meteorological and climatological phenomena are not well understood because terrestrial water storage is rarely studied as a singular variable.

The importance of terrestrial water storage to modern civilization is immeasurable. Besides supplying water for drinking and other domestic uses, surface and aquifer waters are essential for power generation and irrigation. Plants and animals also depend on soil moisture and surface water. Furthermore, groundwater sustains streams between episodes of surface runoff, and snowmelt recharges the other stocks of water.

Unfortunately, cost and logistics have hindered the development of networks for gathering and distributing terrestrial water storage data. Remote sensing holds promise for surface soil moisture [e.g., *Jackson et al.*, 1999; *Spencer*, 2000] and snow mapping [e.g., *Ferraro et al.*, 1996], but current techniques do not resolve deeper soil moisture or groundwater. Models are an alternative, but they are limited by the science that produced them, the quality and availability of observations for input and validation, and computational capability.

A new source of terrestrial water storage observations is expected to emerge when NASA's Gravity Recovery and Climate Experiment (GRACE) launches in 2001. The goal of GRACE is to measure the Earth's gravity field with unprecedented accuracy for 5 years [*Tapley*, 1997]. The experiment will employ two satellites in a tandem orbit, 170–270 km apart at ~480 km initial altitude, and will use precise measurements of the distance between the two as a basis for producing a new model of the global gravity field every 30 days. Because mass redistributions at the Earth's surface, which would result from atmospheric and oceanic circulations and terrestrial water storage fluxes, are the main contributor to gravitational variations, the satellite observations will be inverted to estimate changes in terrestrial water storage, given modeled or observed atmospheric pressure data [*Dickey et al.*, 1999].

Rodell and Famiglietti [1999] used a modeled, global, 2 year time series of soil moisture and snow to investigate the detectability by GRACE of terrestrial water storage variations in 20 continental-scale river basins. The study concluded that variations would likely be detectable depending on the size of the region (≥200,000 km^2) and the magnitude of the variations themselves (at least a few millimeters). This paper builds upon those conclusions by relying on observations rather than modeled data, by using a longer (13+ years) data set to better understand interannual variability, and by examining the contributions of groundwater and surface water variations as well as snow and soil moisture variations to changes in total terrestrial water storage. Despite being only 145,800 km^2, Illinois was chosen as the study area because it is one of only a few large regions in the world where observations of all of the water storage components are systematically collected and centrally archived. Observational data from Illinois were obtained, qual-

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Paper number 2000WR900306.
0043-1397/01/2000WR900306$09.00

ity checked, and temporally interpolated where necessary to produce monthly time series that are continuous from December 1982 to July 1996. In addition, water storage in the soil zone from 2 m depth down to the top of the water table was estimated because it was not monitored. Terrestrial water storage changes, averaged over Illinois, were then compared to estimated uncertainty in the GRACE technique to determine their detectability by GRACE on monthly, seasonal, and annual time steps. In addition, the detectability of the same changes was considered, given five larger spatial scales.

2. Background

The climate of Illinois is humid continental, with hot summers and cold snowy winters. Annual precipitation averages between 85 and 100 cm in most parts, but interannual variability is high. The mean annual cycle of precipitation is relatively uniform throughout the year with a low in the winter months. Evapotranspiration is radiation-driven, returning close to 70% of precipitation to the atmosphere annually despite being nearly nonexistent in the winter [*Eltahir and Yeh*, 1999].

Illinois has a subdued topography defined by flat, glacial, and fluvial plains and rolling hills. The shallow aquifer is unconfined, and the water table usually exists at depths between 1 and 10 m. Water-bearing units are composed of limestone, dolomite, or a mix of sand and gravel. The texture of the surface soil is typically silty loam or silty clay loam, and porosities generally range from 0.40 to 0.55 [*Hollinger and Isard*, 1994]. Parent soil materials include loess, alluvium, and glacial outwash and till [*Changnon et al.*, 1988].

Several investigations have examined terrestrial water storage in Illinois. *Changnon et al.* [1988] attempted to define statistical relationships between monthly precipitation and shallow groundwater levels from a network of monitoring wells and then applied these relationships with physiographic and soils information to predict groundwater levels in times of drought. They concluded that the lag between precipitation and groundwater response was shorter during wet soil conditions than dry and shorter also when the water table was shallow. In general, they found that the best correlated lag times were 0–2 months, the lags being more closely related to soil type than to physiography.

Hollinger and Isard [1994] analyzed observations from a network of neutron probe monitoring stations. They calibrated the probes to soil moisture, quantified the uncertainty in the observations, and identified statewide patterns and relationships in 10 years of data. They determined that uncertainty in the data was ~5–13% at a moisture content of 0.30 volumetric. They demonstrated a clear annual cycle of soil moisture, soils being wettest in early spring (March 15 on average) and driest in late summer (August 15 on average). A latitudinal gradient of soil moisture was seen in winter and spring which corresponded to higher rates of precipitation in the south, and during summer and autumn a longitudinal gradient existed, with wetter soils in the east corresponding to shallower loess deposits.

Two papers focused on the 1988 drought and 1993 flood in Illinois and the Midwestern United States. *Wendland* [1990] utilized time series from 25 lakes, 39 river gaging stations, and the soil moisture and groundwater networks mentioned above to characterize the 1988 drought. Among other findings he concluded that all of four drought indicators correlated better to multimonth cumulative precipitation than to any single lagged month. Additionally, the best correlated time lag between a drought indicator and a single month of precipitation was 1 month, except in the case of lake level, which was better correlated with no time lag. *Kunkel et al.* [1994] described the hydroclimatic causes of the 1993 flood in the context of historical frequencies of heavy rainfall events. Their investigation utilized two long precipitation data sets, soil moisture conditions assessed by a model, and evapotranspiration estimated using the Penman-Monteith formula. They determined that the flood was caused by seven unusual hydrometeorological conditions that contributed to the wettest summer on record in the region.

Eltahir and Yeh [1999] analyzed patterns of hydrological floods and droughts in Illinois, as they propagated from atmosphere through soil to aquifer, by examining atmospheric water vapor flux, precipitation, soil moisture, groundwater level, and river flow. The seasonal cycle of the groundwater level was determined to lag soil moisture by ~1 month. They concluded that seasonal cycles of the hydrological components were forced by the seasonal cycle of solar radiation, while interannual variability in the hydrological cycle was controlled by atmospheric circulation and precipitation. *Yeh et al.* [1998] also used observations of precipitation, runoff, soil moisture, and groundwater in a terrestrial water balance in order to estimate evapotranspiration in Illinois.

In proposing GRACE, *Dickey et al.* [1997] hypothesized that satellite-based gravity measurements obtained by the mission could be inverted to produce estimates of changes in water storage in terrestrial regions. However, the GRACE satellites will be sensitive to gravitational variations caused by the sum of the mass changes in the entire column of fluid and solid material below them. Therefore the contribution of atmospheric mass redistribution will have to be removed from the gravity observations using auxiliary information, such as modeled pressure fields. Furthermore, GRACE will not be able to distinguish changes in the different components of terrestrial water storage. *Wahr et al.* [1998] and *Rodell and Famiglietti* [1999] evaluated the aforementioned hypothesis using 5 years and 2 years, respectively, of modeled soil moisture and snow data with estimates of uncertainty in the inversion technique. The two studies agreed that terrestrial water storage changes would be detectable on monthly and longer intervals given large enough regions and depending on the magnitude of the changes themselves. However, the effects groundwater and surface water storage variations were not considered due to a lack of data. This paper will address that deficiency and examine a longer, observation-based time series.

3. Data

This investigation required four sets of water storage data from Illinois. The groundwater data set consisted of water levels from 18 wells (see Figure 1) monitored by the Illinois State Water Survey (ISWS) [*Changnon et al.*, 1988]. The wells ranged in depth from 3 to 24 m and were in communication with the local unconfined aquifer. None were close enough to be affected by streams or pumping wells. At most of the wells, once-per-month observations were supplemented by monthly high and low water levels read from a continuous recording device. Monitoring continued through March 1997, beginning at one well in 1988, at two in 1984, and at the rest prior to 1981.

Soil moisture measured at 19 sites in Illinois comprised the second data set. *Hollinger and Isard* [1994] provide a thorough description and analysis of these data, which were produced by ISWS and archived in the Global Soil Moisture Data Bank [*Robock et al.*, 2000]. Neutron probes calibrated by the gravimetric technique were used to measure moisture in 11 soil layers from the surface to 2 m depth (nine 20 cm layers bounded by 10 cm top and bottom layers). Aside from a second probe at Dixon Springs that was set in bare soil, the predominant vegetation at all of the sites was grass. Observations were made one, two, or three times per month. Sixteen of the time series began in 1981 or 1982, two began in 1986, and one began in 1991.

Records of daily snow depth, snowfall, precipitation, and temperature observations were downloaded from the Midwestern Climate Information System [*Kunkel et al.*, 1990], which is operated by the Midwestern Climate Center. Twenty-eight stations were selected from a database of 452, and several others were tapped for auxiliary data. Selections were based on the completeness of the time series between 1982 and 1996 and a desire to sample the region evenly.

Observations of water levels at 49 reservoirs were provided by ISWS. Reservoir operators typically measured the height of water at the spillway at the end of each month. Time series ranged in length from 2 to 40 years. Surface areas were determined from an ISWS report [*Singh and McConkey-Broeren*, 1990], data files from ISWS (W. Saylor, ISWS, personal communication, 1999), and various topographic maps. Records from five of the seven lakes with areas >10 km^2 extended from 1983 or 1985 to 1999; the other two extended from 1988 to 1999. The total surface area of the reporting reservoirs ranged from 219 km^2 at the start of the time series to 342 km^2 in the summer of 1993 and back down to 331 km^2 by the end of the time series.

4. Methods

4.1. Interpolation and Averaging

Missing daily observations of snow depth either were taken from neighboring, auxiliary stations or were estimated using existing snowfall, precipitation, and temperature measurements. Following a rigorous examination to remove spurious values, daily time series of groundwater depths and soil moisture in the 11 layers were constructed by linearly interpolating between observation dates. For days when the water table rose above 2 m depth, which happened frequently at certain locations, the depth to groundwater value was reset to 2 m, so that saturated storage in the upper soil zone would not be added twice, once as soil moisture and once as groundwater, when computing total water storage. Whenever two consecutive depth-to-water values were 2 m at a particular location, the change in groundwater level was computed to be zero. Consequently, changes in statewide average groundwater storage, as it is conventionally defined, were attenuated. In short, groundwater storage changes were intentionally underestimated in order to preserve the accuracy of the total water storage changes, and the underestimation was more significant during wet periods when the water table was high. These facts should be weighed when comparing groundwater changes published elsewhere to those herein.

Water stored in the intermediate zone (here defined as the soil zone below the 2 m observation depth and above the water table) was estimated using the daily interpolated time series for eight locations where a soil moisture station and a groundwater well were in close proximity (Figure 1) and had complete

Figure 1. Map of Illinois showing locations of groundwater, soil moisture, and snow measurements and intermediate zone estimates.

records between 1982 and 1996. Moisture content was estimated by assuming a linear increase in wetness with depth between the deepest observed soil layer and the water table, where the soil is saturated, thus

$$\mathrm{IZ} = H_{\mathrm{IZ}}\left(\frac{\theta + n}{2}\right), \tag{1}$$

where IZ is intermediate zone water storage, H_{IZ} is the height of the intermediate zone, θ is the water content of the deepest observed soil layer, and n is the porosity. Though not ideal, this technique was considered adequate for the purposes of this investigation. When the depth to groundwater at a site was 2 m or less, the height of the intermediate zone was 0 m and, accordingly, intermediate zone storage was nil. At each site, porosity values were estimated using information on soil material and aquifer type from *Changnon et al.* [1988], information on soil type and porosity in the upper 1 m from *Hollinger and Isard* [1994], and an evaluation of the maximum recorded water content in the deepest soil layer.

For each station, monthly mean depth to water, soil moisture, intermediate zone moisture, and snow depth were then computed by averaging the daily values. Seasonal (i.e., winter, spring, summer, and autumn) means and annual means were computed similarly. The equivalent depth of water in each soil layer was computed by multiplying the volumetric water content by the height of the soil layer. Well water level (above an arbitrary datum) and snow were converted to equivalent depths of water using site-specific porosity estimates and a snow density of 0.1 g/cm^3 [*Dingman*, 1994]. Time series of statewide average groundwater, soil moisture, intermediate zone, and snow water storages were then calculated using the Thiessen polygon method. The mean reservoir level for a given month was estimated as the average of the observed level at the end of that month and the level at the end of the month previous. Each reservoir's monthly contribution to statewide water storage was computed as the water level multiplied by the area of the reservoir divided by the area of Illinois. Storage in each stock was evaluated relative to its minimum value, which was set to zero. Total terrestrial water storage changes were computed as the sum of the component changes (recall that GRACE will not be able to parse the components). The resulting time series span the period between December 1982 and July 1996.

4.2. Uncertainty Estimation

Uncertainty in the GRACE-derived water storage variations mainly will originate from the instrument's own limitations and the removal of the effect of atmospheric mass redistribution from the observed gravity fields. Because observations of surface pressure (a surrogate for atmospheric mass) are not available globally, modeled fields produced by the National Center for Environmental Prediction and the National Center for Atmospheric Research (NCEP/NCAR) and the European Centre for Medium-Range Weather Forecasts (ECMWF) will be relied upon when removing the effect, and therefore errors in those fields will propagate to the terrestrial water storage change estimates. Uncertainty caused by the removal of the effect of postglacial rebound has been demonstrated to be insignificant in most regions of the world [*Rodell and Famiglietti*, 1999] and was ignored in this investigation.

Errors in the orbital parameters, microwave ranging measurements, accelerometer measurements, and error in the ultrastable oscillator will all contribute to instrument uncertainty. Total instrument uncertainty will be inversely related to both the size of the region and the length of the measurement averaging period (defined as the time period during which GRACE observations contributing to a single global gravity field are gathered). *Dickey et al.* [1997] provide a more thorough discussion of the error sources and characteristics. A table of total instrument uncertainty as it varies with spatial and temporal resolution was provided by the GRACE science team (S. Bettadpur, University of Texas, personal communication, 1998). Using this information, the error in a single GRACE measurement was estimated for monthly (30 day), seasonal (90 day), and annual (365 day) averaging periods for Illinois (145,800 km^2) and five larger areas, following the computational technique employed by *Rodell and Famiglietti* [1999]. Two GRACE measurements will be required to identify a change in the gravity field; therefore the following relation was used to calculate instrument errors E_I in an estimate of the change in water storage:

$$E_I = \sqrt{E_{I,1}^2 + E_{I,2}^2}, \qquad (2)$$

where $E_{I,1}$ and $E_{I,2}$ are the instrument errors in GRACE measurements of the global gravity field for averaging periods 1 and 2.

To account for the errors associated with removing the effect of atmospheric mass redistribution from the gravity fields, atmospheric pressure data were obtained from the ECMWF Re-analysis [*ECMWF*, 1996] and the NCEP/NCAR Reanalysis [*Kalnay et al.*, 1996] and used to estimate atmospheric errors $E_{A,i}$. Both data sets are globally gridded at 2.5° resolution. Monthly, seasonal, and annual errors for each region were computed as by *Wahr et al.* [1998]:

$$E_{A,i} = \frac{|\bar{P}_{\mathrm{ECMWF}} - \bar{P}_{\mathrm{NCEP/NCAR}}|}{\sqrt{2}}, \qquad (3)$$

where $\bar{P}$ is mean surface pressure over a particular region and time period i. Dividing by $\sqrt{2}$ accounts for the assumption that the two pressure estimates contribute equally to the variance in ($\bar{P}_{\mathrm{ECMWF}} - \bar{P}_{\mathrm{NCEP/NCAR}}$), which difference is assumed to be comparable to the error in the modeled pressure fields. The atmospheric error estimates were then used to compute the associated uncertainty in changes in water storage:

$$E_A = \sqrt{E_{A,1}^2 + E_{A,2}^2}, \qquad (4)$$

where E_A is the atmospheric error in the storage change and $E_{A,1}$ and $E_{A,2}$ are the atmospheric errors in GRACE measurements for averaging periods 1 and 2, respectively. To produce a conservative estimate, total uncertainty in the change in storage E_T was taken as the sum of the two error components:

$$E_T = E_I + E_A. \qquad (5)$$

5. Results

5.1. Terrestrial Water Storage

Figure 2 depicts the entire 13.5 year time series of terrestrial water storage in Illinois. Statewide average storage in each component is shown relative to its minimum value, which has been set to zero. The five components, groundwater (GW), intermediate zone storage (IZ), soil moisture (SM), snow water (SN), and reservoir water storage (RS), are superposed so that total terrestrial water storage S_T is the resulting uppermost contour.

Because GW was defined as the elevation of the water table (above the minimum elevation) multiplied by the porosity (as opposed to the specific yield), with IZ accounting for the remaining moisture in the unsaturated zone below 2 m, GW changes may appear to be abnormally large at first glance. IZ generally increases (decreases) as GW decreases (increases) because the intermediate zone becomes taller (shorter) with a greater (lesser) storage capacity as the water table declines (rises). In this way, IZ changes buffer GW changes. The control volumes of IZ and GW are not fixed; consequently, it is often simpler and more enlightening to consider the sum of the two (GW + IZ), which is the water stored in a control volume beneath 2 m depth. GW + IZ (the contour at the top of IZ in Figure 2) is less variable than its two components and behaves similarly to groundwater storage estimated as the height of the water table multiplied by the specific yield (not shown). Using a specific yield estimate to study GW alone was considered inferior to studying GW + IZ because the former demands that the volumetric water content of deep soil is always equal

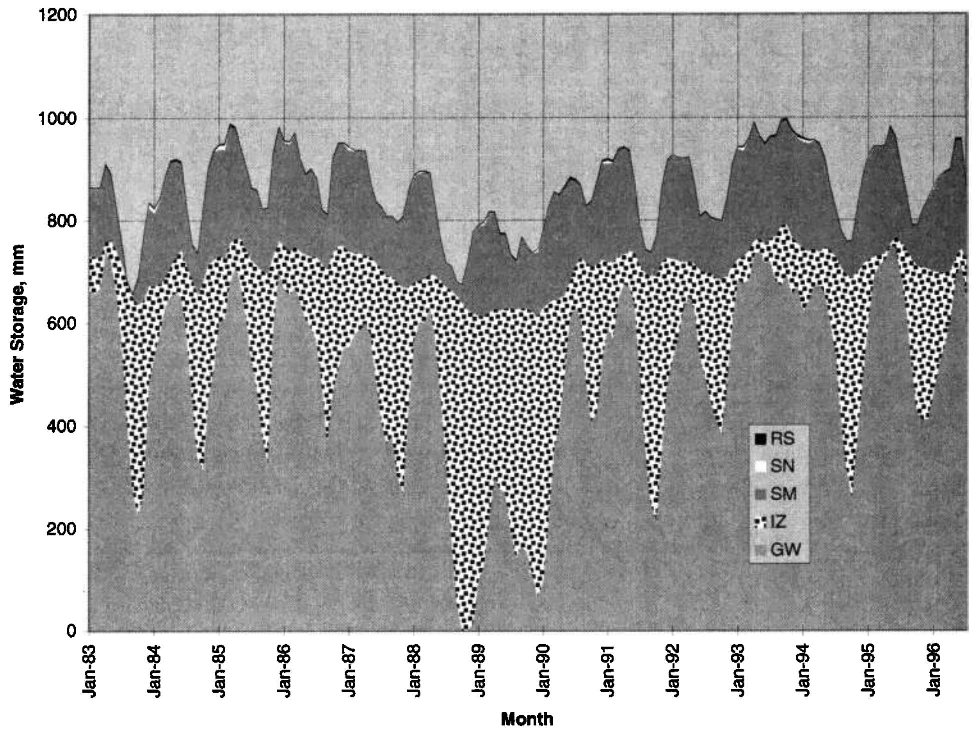

Figure 2. The 13.5 year time series of terrestrial water storage components averaged over Illinois, relative to their minimum values, which have been set to zero.

to either the porosity or the porosity minus specific yield, while the latter allows for a range of water contents.

Figure 2 demonstrates that there is significant seasonal and interannual variability in terrestrial water storage in Illinois. Changes in SM and GW + IZ (ΔSM and ΔGW + IZ) are the dominant contributors to S_T variations (ΔS_T). The state-averaged change in SN (ΔSN) is only occasionally significant, as in February 1985, when it was nearly 10 mm equivalent height of water. Changes in RS (ΔRS) are less substantial, but what effect unmeasured, unregulated bodies of surface water might have is not known. Figure 2 suggests that there may be a cycle of terrestrial water storage in Illinois with a period of ~7 years, possibly linked to the El Niño–Southern Oscillation (ENSO). Both ENSO, as described by the Multivariate ENSO Index [*Wolter and Timlin*, 1998] (not shown), and terrestrial water storage went through about two cycles between 1983 and 1996. However, a longer time series would be needed to test this hypothesis thoroughly.

The drought of 1988–1989 and the wet summer of 1993 are obvious in Figure 2. These episodes help to confirm the validity of the water storage estimation. During the drought, SM became depleted midway through 1988 but recovered somewhat by year's end. GW reached a series low in October of that year and did not recover fully until mid-1990. GW + IZ dropped steadily until February 1989 and never rebounded until after it had reached a series low in December 1989. The minimum S_T between 1983 and 1996 actually occurred in September 1983, but October 1988 was nearly as dry. Furthermore, in 1989 the peak S_T, which typically occurs in the spring, was much lower than normal. Also, 1993 was a wet year from start to finish, and the series maximum S_T occurred in October of that year. SM was high throughout 1993, peaking in April. GW peaked a month later, but GW + IZ reached its series high in October, the same month as the S_T maximum.

Figure 3 plots the annual cycle of mean monthly terrestrial water storage changes. S_T increases from October through April, with a maximum average gain of 46 mm (equivalent height of water) in November, and decreases in the remaining warm months, distributed around a maximum average loss of 41 mm in July. ΔSM has an annual cycle very similar to ΔS_T and tends to be the dominant component in all months except April and September, with maximum average changes of −31 mm in July and +39 mm in November. The cycle of ΔGW + IZ lags 0–2 months behind the cycle of ΔSM. Maximum average changes in GW + IZ are −10 mm in July, August, and September and +7 mm in March and December. Gains to SN are most frequent in December and January, and *SN* losses tend to occur in March, all averaging 1.6–2.6 mm. The largest RS gains tend to occur in April and May, while the largest losses tend to occur in July and August, but all monthly RS changes average <1 mm.

Figure 4 shows the mean absolute changes in RS, SN, SM, GW + IZ, and S_T for each month of the year. As an example, the mean absolute change in total water storage for April was computed as

$$|\overline{\Delta S_{T(\mathrm{April})}}| = \sum_{\mathrm{year}=1983}^{1996} |S_{T(\mathrm{April,year})} - S_{T(\mathrm{March,year})}|/14, \quad (6)$$

where $S_{T(\mathrm{April,year})}$ is the average total water storage in Illinois in April of a given year and 14 is the number of Aprils in the time series. Mean absolute changes were examined because the magnitudes of the changes are what determine their de-

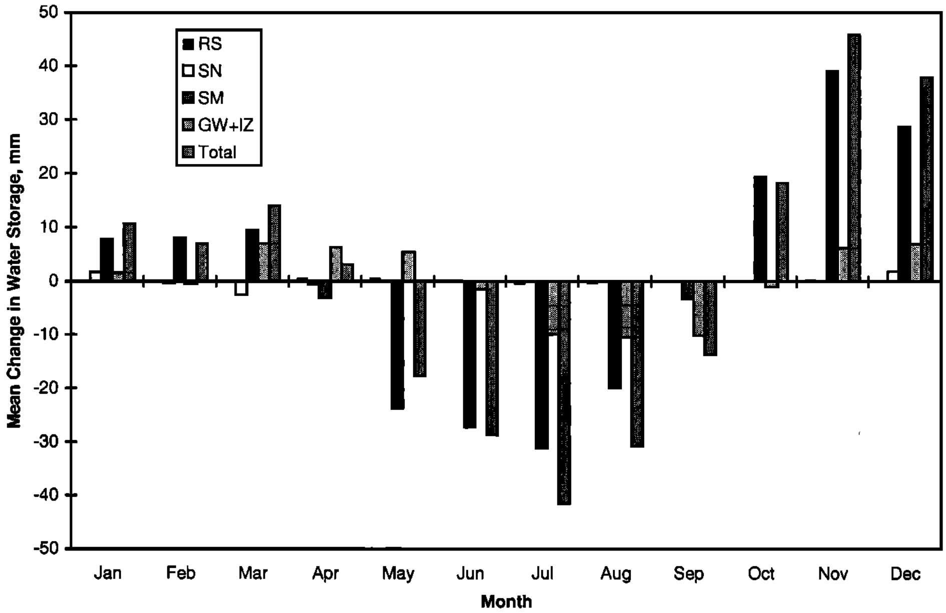

Figure 3. Mean annual cycles of monthly changes in terrestrial water storage and its components.

tectability by GRACE. Simple mean changes misrepresent the average magnitudes.

In Figure 4 it is again apparent that ΔSM and ΔGW + IZ dominate. Mean absolute ΔS_T varies between 10 mm in February and 51 mm in November, averaging 28 mm per month over the course of the time series. ΔSM is typically the largest component, averaging a low of 9 mm in February and April and peaking at nearly 40 mm in November. Mean ΔGW + IZ varies between 4 mm in February and nearly 16 mm in July. Mean ΔSN is smaller but has the potential to be a factor from December through March, when it averages 1.9–2.8 mm. ΔRS averages <1 mm per month throughout the year.

Figure 5 shows mean absolute seasonal (i.e., December–February (DJF), March–May (MAM), June–August (JJA), and September–November (SON)) changes in water storage, calculated similarly to the mean absolute monthly changes (e.g., replace the subscripts April and March with MAM and DJF in (6)). Mean absolute ΔS_T is largest in winter (DJF) and summer (JJA), averaging between 80 and 90 mm in those seasons, and smaller in spring (MAM) and autumn (SON), averaging ~30 mm. ΔSM dominates in winter and summer, averaging over 70 mm, but in spring and autumn it is close in magnitude to ΔGW + IZ, which averages ~20 mm in every season. ΔSN averages close to 3 mm in winter and spring and is insignificant in summer and autumn. ΔRS averages 1 mm or less in all seasons.

Figure 6 shows changes in annual mean terrestrial water storage. For example, the 1984 change in soil moisture storage was calculated as

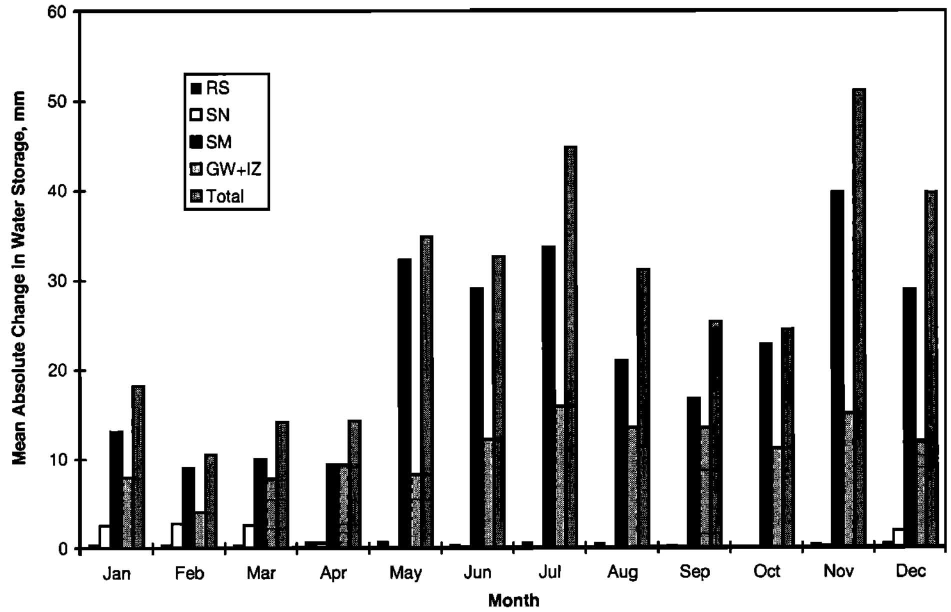

Figure 4. Mean magnitudes of monthly changes in terrestrial water storage and its components.

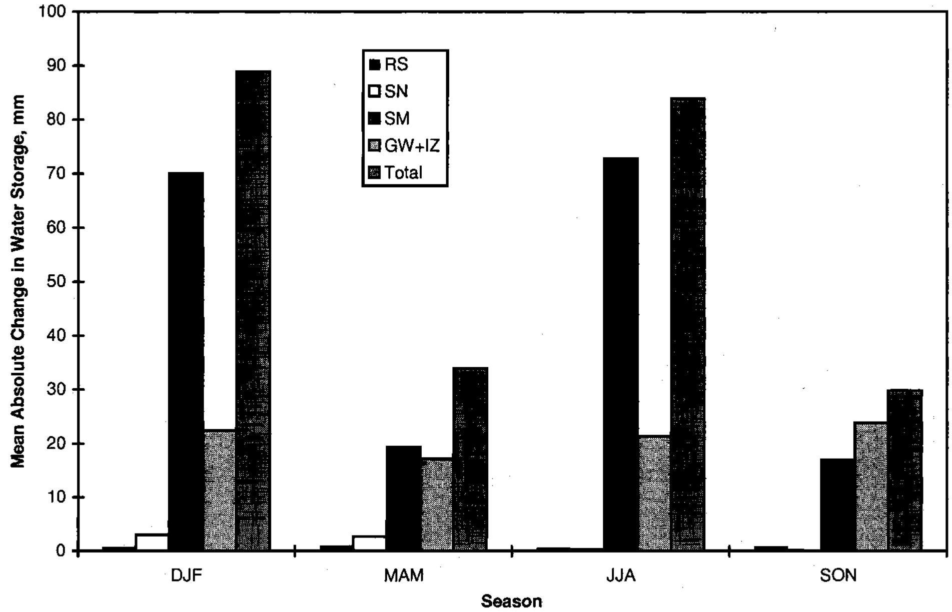

Figure 5. Mean magnitudes of seasonal changes in terrestrial water storage and its components.

$$\Delta SM_{1984} = SM_{1984} - SM_{1983}, \tag{7}$$

where SM_{1984} is the average soil moisture storage in 1984 and SM_{1983} is the average in 1983. The largest change in S_T occurred in 1993, a gain of 105 mm that coincided with massive flooding. A large gain (86 mm) also occurred in 1990, which ended the drought of the previous 2 years. The largest loss to S_T, nearly 100 mm, occurred in 1994, compensating for the previous year's gain. Consecutive S_T losses from 1986 to 1989, including a loss that exceeded 75 mm in 1988, contributed to the drought of 1988–1989. The average magnitudes of annual changes in SM and GW + IZ were nearly identical, ~30 mm, but individual changes were often dissimilar. In 1989, 1991, and 1992, significant but opposite changes in SM and GW + IZ neutralized each other, resulting in small S_T changes. That circumstance resulted from the apparent lagging relationship of GW + IZ to SM: from 1985 to 1993, ΔGW + IZ had the same sign as the previous year's ΔSM, and the trend might have continued if not for the unusually large storage gains in 1993 and subsequent recovery the following year. Thus annual soil moisture changes may prove to be a good predictor of annual changes in deeper water storage. Figure 6 also demonstrates that annual ΔRS was insubstantial, peaking at 1.6 mm in 1993. The magnitude of ΔSN also was small on an annual basis, only exceeding 1 mm once in 12 years.

5.2. Potential Accuracy of GRACE-Derived Water Storage Change Estimates

Figure 7 plots mean monthly absolute changes in S_T for Illinois. The error bars represent $\pm E_T$, the total uncertainty in

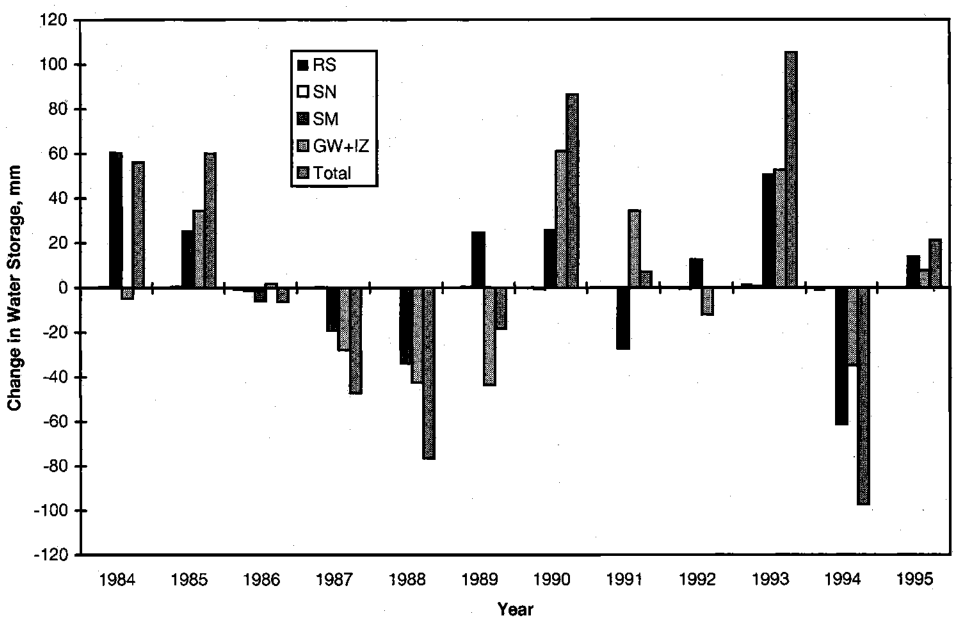

Figure 6. Annual changes in terrestrial water storage and its components.

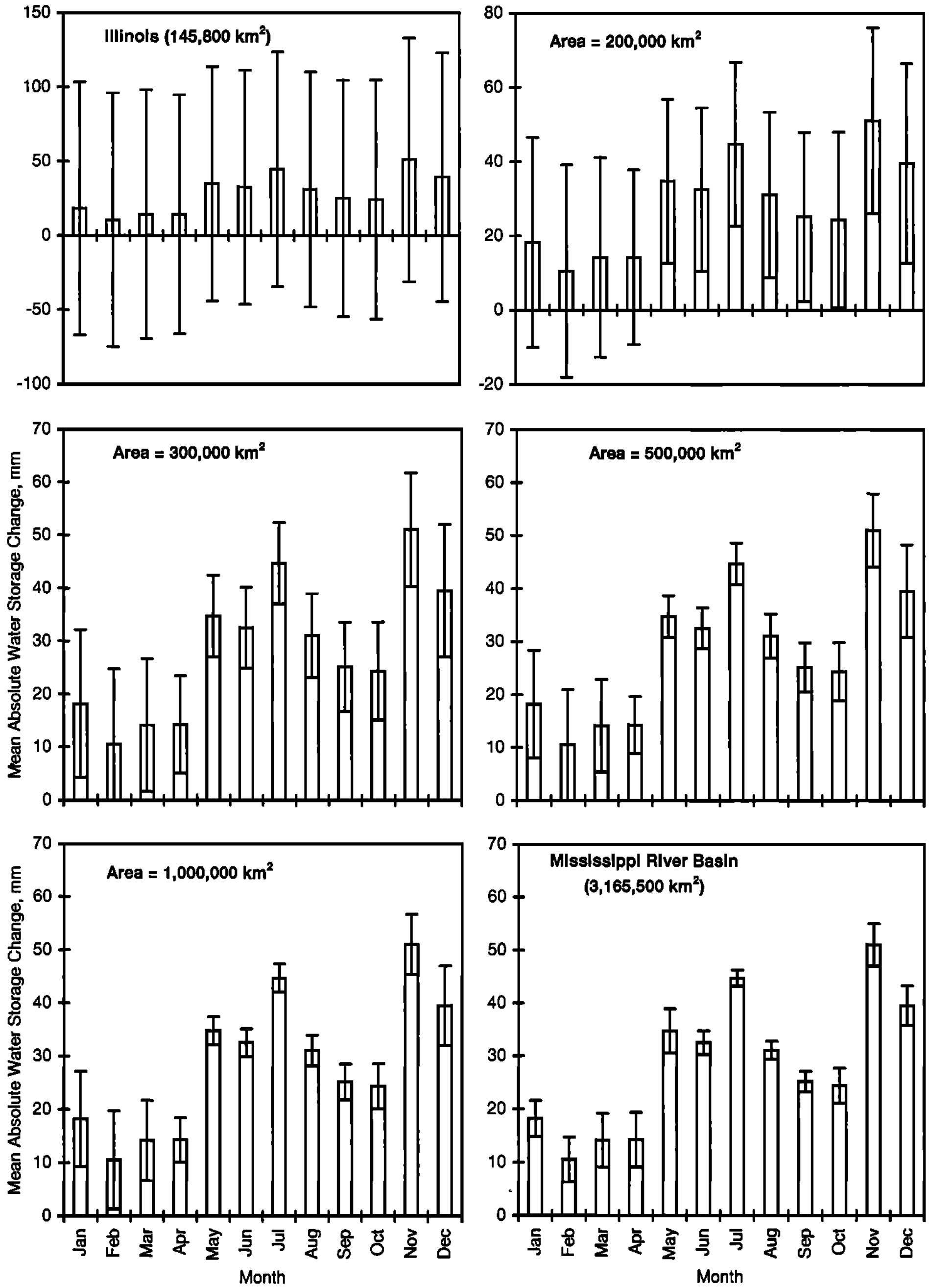

Figure 7. Mean magnitudes of monthly changes in terrestrial water storage, with error bars that represent the total uncertainty in GRACE-derived estimates, for six spatial scales.

a hypothetical GRACE-derived estimate of ΔS_T, averaged for each month of the year. As mentioned before, GRACE instrument errors increase as the area of the observed region decreases. Because the area of Illinois is only 145,800 km^2 (not including its share of Lake Michigan), uncertainty is large enough that monthly changes in terrestrial water storage in Illinois normally will not be detectable by GRACE ($E_T/\Delta S_T \geq 1$), as seen in the top left panel of Figure 7. This is not surprising considering the conclusions of previous studies; however, the use of observations was prioritized in this investigation. The data from Illinois become more practical if the assumption is made that larger surrounding regions have water storage changes that are comparable in magnitude, so that the monthly changes may be detectable by GRACE depending on the size of the region. The other five panels in Figure 7 show the same ΔS_T values as the first, with E_T for regions with areas 200,000, 300,000, 500,000, 1,000,000, and 3,165,500 km^2, the last area being equal to that of the Mississippi River basin. For a 200,000 km^2 region the total water storage changes typically would be detectable ($E_T/\Delta S_T < 1$) from May through December and undetectable the other four months of the year. For a 300,000 km^2 region, ΔS_T would typically be detectable in all months except February. For 500,000 km^2 and larger regions, ΔS_T would typically be detectable in all months of the year with relative uncertainty ($E_T/\Delta S_T$) decreasing as the area of the region increased.

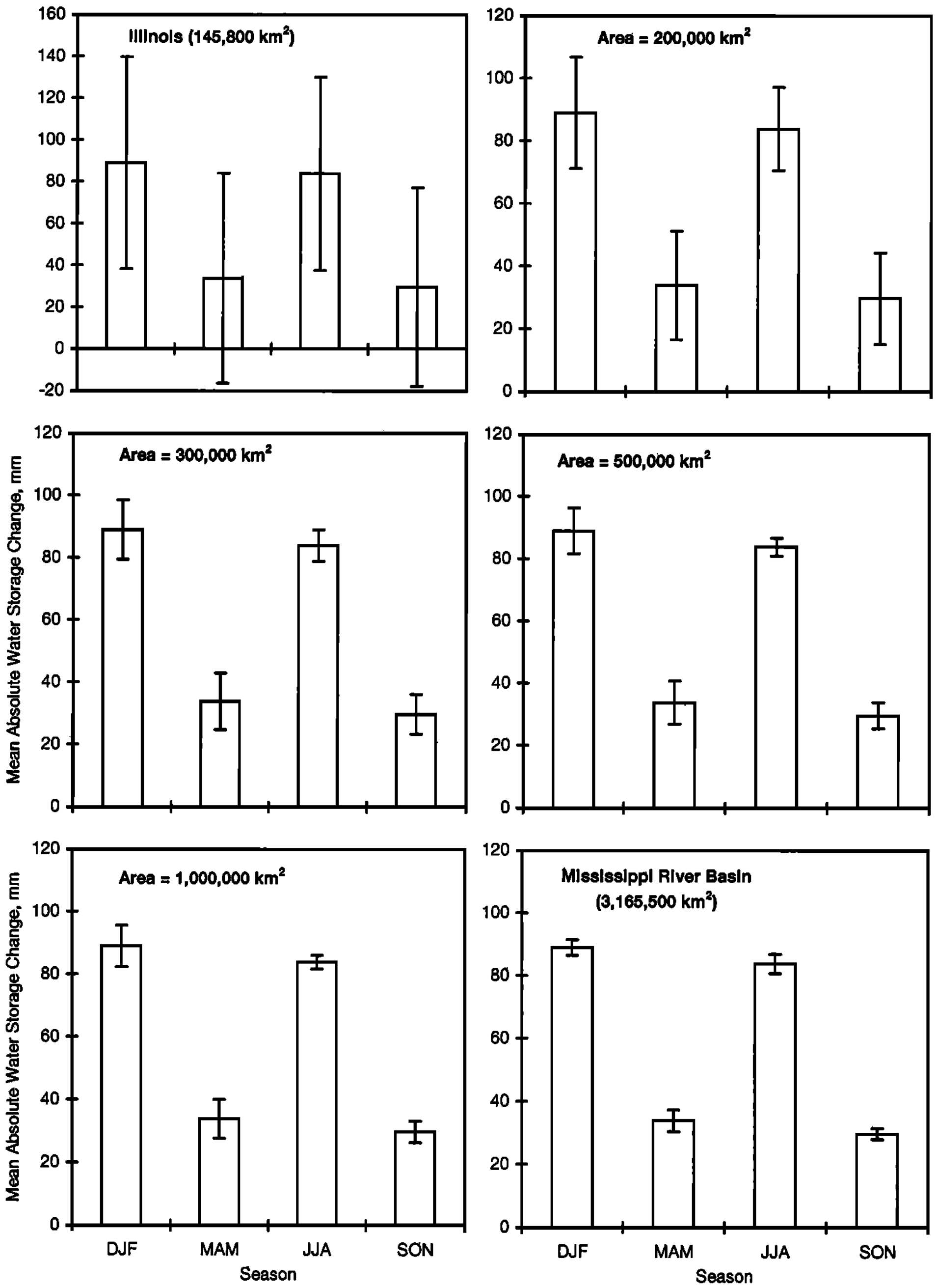

Figure 8. Mean magnitudes of seasonal changes in terrestrial water storage, with error bars that represent the total uncertainty in GRACE-derived estimates, for six spatial scales.

Figure 8 plots mean seasonal absolute changes in S_T for Illinois with error bars to represent the mean uncertainty in a GRACE estimate for each season. Recall that GRACE errors decrease as the averaging period increases. ΔS_T will typically be detectable in Illinois in winter and summer and undetectable in spring and autumn. ΔS_T will often be detectable in all seasons in 200,000 km^2 and larger regions, with decreasing relative uncertainty as the area increases.

Figure 9 plots the annual changes in S_T from 1984 to 1995 with error bars to depict the uncertainty in the GRACE estimate for each year. In Illinois, ΔS_T was detectable in 7 of 12 years. ΔS_T was detectable in 9 of 12 years for a 200,000 km^2 region, 10 of 12 years for a 300,000 km^2 region, and 11 of 12 years for 500,000 km^2 and larger regions. In 1992, ΔS_T was undetectable at all scales because it was only -0.11 mm, but this is effectively a nonchange that would have been identified by GRACE to within ± 5 mm for a 300,000 km^2 region.

Table 1 lists the means and ranges of terrestrial water storage changes in Illinois between January 1983 and July 1996. On average, seasonal ΔS_T is largest (58.7 mm), followed by annual (48.5 mm) and monthly (28.2 mm) ΔS_T. ΔSM is often the principal component, and it is also greatest on a seasonal basis. Mean ΔGW + IZ becomes larger for longer averaging periods; annual ΔGW + IZ is about the same magnitude as annual ΔSM. ΔSN is occasionally significant on a monthly or seasonal basis, while ΔRS is never >2–3 mm, being largest seasonally. All types of water storage changes are occasionally as small as 6 mm or less.

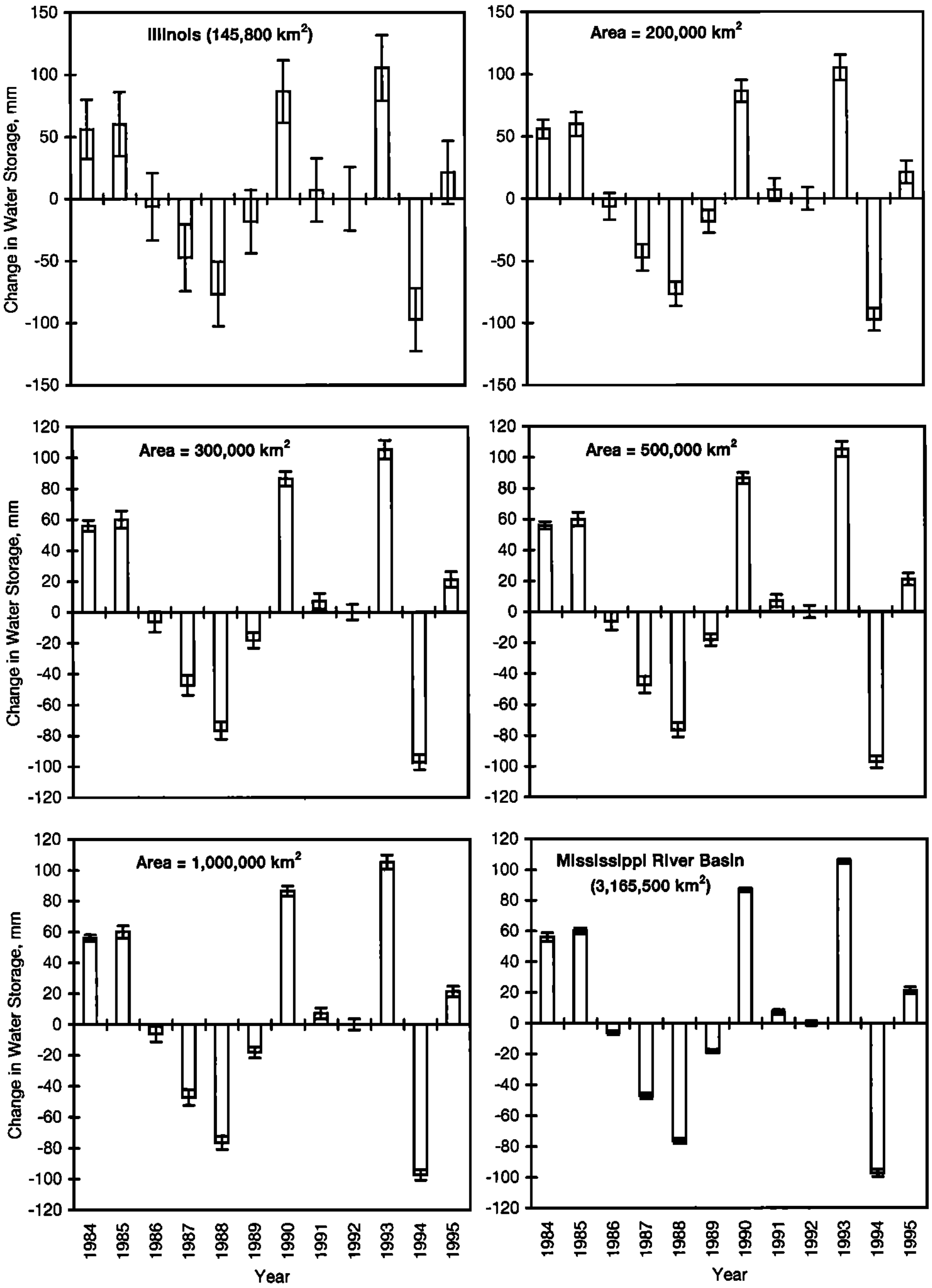

Figure 9. Annual changes in terrestrial water storage, with error bars that represent the total uncertainty in GRACE-derived estimates, for six spatial scales.

Table 2 lists the means and ranges of uncertainty in GRACE-derived estimates of ΔS_T for the six previously defined areas. Variations in E_T for a particular area and averaging period are due to changes in atmospheric uncertainty, as agreement between the NCEP/NCAR and ECMWF models varies. Comparison of Tables 1 and 2 reveals what areas and averaging periods should allow GRACE to produce workable estimates of ΔS_T. Monthly ΔS_T will only be detectable over Illinois when it is very large, >81.4 mm on average. However, seasonal and annual ΔS_T may be detectable over Illinois more often than not. Mean ΔS_T is larger than mean E_T for all regions 200,000 km^2 and greater, for all three timescales, but for any specific situation, ΔS_T must be greater than the minimum for it to be detectable.

6. Discussion

Table 3 lists the number of intervals when ΔS_T was detectable by GRACE (i.e., it was large enough that it would, in the future, be detectable by GRACE) during the period of the time series for the six spatial and three temporal scales. Monthly ΔS_T was detectable in Illinois (145,800 km^2) only 5% of the time, but seasonal and annual ΔS_T were detectable about half the time. The rate of monthly water storage change detectability jumped to 44% for a 200,000 km^2 region and to 67% for a 300,000 km^2 region, then increased more gradually up to 82% for the Mississippi River basin (3,165,500 km^2). It appears that 82% is an approximate upper limit to the monthly change detection rate because ~18% of the ΔS_T values are

Table 1. Range of Absolute Changes in Terrestrial Water Storage for Monthly, Seasonal, and Annual Averaging Periods[a]

Component	Monthly ΔS, mm			Seasonal ΔS, mm			Annual ΔS, mm		
	Min	Mean	Max	Min	Mean	Max	Min	Mean	Max
RS	0.0	0.5	2.3	0.0	0.8	2.7	0.0	0.5	1.6
SN	0.0	0.9	9.8	0.0	1.5	6.0	0.0	0.4	1.1
SM	0.2	22.0	76.5	0.2	44.3	121.5	5.8	29.9	61.3
GW + IZ	0.1	10.8	35.2	0.8	21.1	68.7	1.8	29.8	61.0
S_T	0.0	28.2	109.8	4.8	58.7	141.8	0.1	48.5	105.2

[a]Water storage values are equivalent heights of water. ΔS is change in water storage. RS is reservoir storage, SN is snow water storage, SM is moisture storage in the top 2 m of soil, GW + IZ is groundwater and intermediate zone storage, and S_T is total terrestrial water storage.

smaller than the typical uncertainty at the largest spatial scales. Seasonal ΔS_T was detectable 85% of the time for a 200,000 km^2 region, and the rate increased rapidly to 100% for a 500,000 km^2 region; recall from Tables 1 and 2 that the minimum seasonal ΔS_T and the mean seasonal E_T are both ~5 mm at that spatial scale. Annual ΔS_T was detectable 9 times out of 12 for a 200,000 km^2 region and 11 times out of 12 for 500,000 km^2 and larger regions. As mentioned in section 5, the 1992 annual change (−0.11 mm) was not realistically detectable at any scale.

Because the lifetime of the GRACE mission will be 5 years, it is worth examining the range of variability of ΔS_T for periods of 5 consecutive years. Mean E_T does not vary appreciably among 5 year periods, so the series means in Table 2 are appropriate for comparison. For monthly changes the least variable 5 year period of the time series was May 1989 through April 1994, when absolute ΔS_T averaged 22.6 mm, or 5.6 mm less than the series mean, 28.2 mm (Table 1). That difference was enough to reduce the detectability rate to 36% for a 200,000 km^2 region, down from a series mean of 44%. Rate changes at other spatial scales were less significant. The 5 year period with the largest mean monthly ΔS_T (33.4 mm) was March 1983 to February 1988. Detectability rates at the six spatial scales were 2–7% greater than the series means for that period. For seasonal ΔS_T the least and most variable 5 year periods were June 1989 through May 1994 and March 1983 through February 1988, when the means were 47.1 mm and 64.5 mm. The series mean seasonal ΔS_T was 58.7 mm. During the first period the detectability rate for Illinois was 32%, compared to a series mean of 49%, while the rate for the second period was 58%. The rate did not range >6% from the mean at the other spatial scales. For annual ΔS_T the least and most variable periods were 1988–1992 and 1986–1990, when the means were 28.0 mm and 57.1 mm, compared to a series mean of 48.5 mm. For Illinois, 1 of 4 annual water storage changes were detectable during the first period, and 3 of 4 were detectable during the second period, 7 of 12 changes being detectable over the course of the time series. For the other spatial scales, 2 or 3 out of 4 changes were detectable for the first period, and 4 of 4 were detectable for the second, compared with 9–11 out of 12 for the time series. Thus the degree of water storage variability is not constant among 5 year periods and will influence detectability during the 5 year GRACE mission, especially at smaller spatial scales and on longer averaging periods.

Additional information, such as auxiliary observations, concurrent model runs, or at least a knowledge of the soil moisture climatology, will be required to isolate changes in the component stocks (e.g., ΔSM) from GRACE-derived ΔS_T estimates. Given that caveat, comparison of the ranges of terrestrial water storage changes in Table 1 to the uncertainty information in Table 2 provides some insight into the potential for decomposition of ΔS_T. Presumably, the uncertainty in ΔS_T would have to be smaller than the magnitude of a component change for an estimate of that component change to be meaningful. If this is true, then ΔRS would rarely, if ever, be resolvable. In the Midwestern United States, ΔSN would be resolvable only on monthly or seasonal timescales in 500,000 km^2 or larger regions in winters when a deep snow cover persisted. ΔSN would be more easily resolvable in higher latitudes where the annual cycle of SN is more prominent. Monthly ΔSM is likely to be isolated from ΔS_T in regions 300,000 km^2 and larger, while seasonal and annual changes have the potential to be distinguished in regions as small as 200,000 km^2 or possibly the size of Illinois. Monthly ΔGW + IZ might be isolated in 300,000 km^2 regions, seasonal ΔGW + IZ might be isolated in 200,000 km^2 regions, and annual ΔGW + IZ might be isolated in regions as small as Illinois.

An assumption of this investigation was that the record of water storage observations in Illinois could be used as a proxy for larger regions. However, in studying progressively larger regions from the point scale to the global scale, one might expect terrestrial water storage changes of all sizes and signs to begin to be encompassed, causing the magnitudes of the mean changes to approach zero. The authors deemed the assump-

Table 2. Range of Total Uncertainty in GRACE-Derived Water Storage Change Estimates[a]

Region (Area)	Monthly E_T, mm			Seasonal E_T, mm			Annual E_T, mm		
	Min	Mean	Max	Min	Mean	Max	Min	Mean	Max
Illinois (145,800 km^2)	77.8	81.4	89.9	44.8	48.6	54.3	23.9	25.4	27.1
Illinois (200,000 km^2)	21.0	24.5	33.1	12.0	15.8	21.5	7.6	9.1	10.8
Illinois (300,000 km^2)	6.6	10.1	18.7	3.7	7.5	13.2	3.5	5.0	6.6
Illinois (500,000 km^2)	2.8	6.4	14.9	1.5	5.3	11.0	2.4	3.9	5.6
Illinois (1,000,000 km^2)	1.6	5.1	13.7	0.8	4.6	10.3	2.1	3.5	5.2
Mississippi basin (3,165,500 km^2)	0.7	3.4	7.3	0.5	2.7	5.0	1.1	2.2	4.2

[a]Uncertainty values are equivalent heights of water. E_T is total uncertainty.

Table 3. Number of Time Intervals During Which ΔS_T Is Detectable ($E_T/\Delta S_T < 1$) by GRACE

Region (Area)	Number of Time Intervals: Monthly (163 Total)	Seasonal (53 Total)	Annual (12 Total)
Illinois (145,800 km^2)	8	26	7
Illinois (200,000 km^2)	72	45	9
Illinois (300,000 km^2)	109	51	10
Illinois (500,000 km^2)	121	53	11
Illinois (1,000,000 km^2)	131	53	11
Mississippi basin (3,165,500 km^2)	133	53	11

tion necessary in order to achieve the objective of evaluating the GRACE technique based on observations rather than modeled data, but the reader is advised to use caution in interpreting the results. Furthermore, the water storage data should be viewed as specific to the Midwestern United States, Illinois in particular. For a global assessment of the potential to derive water storage change estimates from GRACE the reader is advised to consult *Rodell and Famiglietti* [1999].

The surface water storage changes presented herein might have been larger if unregulated lake, river channel, and floodplain storage had been included with reservoir storage. In the summer of 1993, when the Mississippi River and its tributaries overflowed their banks and inundated their floodplains, there was certainly more surface water and total water storage than described by these time series. Unfortunately, observations that quantified these events were not found. Observations of confined aquifer storage changes also would have made this study more complete. It is recommended that future studies make a global assessment of water storage change detectability that includes groundwater and intermediate zone storage variations, which have been shown to be significant. Also, a technique should be developed for decomposing GRACE-derived total water storage changes, after sources of auxiliary observations have been identified around the world. Employing all GRACE-derived and auxiliary terrestrial water storage data as constraints in a hydrological model is suggested. These advancements would expedite the delivery of useful GRACE-derived products to the hydrological community.

7. Summary

Time series of groundwater, soil moisture, snow depth, and reservoir level observations from Illinois were collected, quality checked, and temporally interpolated where necessary. Intermediate zone water storage was estimated. Expected uncertainty in GRACE-derived water storage changes, which decreases with increasing spatial and temporal scales, was calculated. Terrestrial water storage changes, averaged over Illinois, were compared to uncertainty estimates to determine their detectability by GRACE on monthly, seasonal, and annual time steps on six spatial scales. Uncertainty was typically too large to allow detection of monthly water storage changes over Illinois, while seasonal and annual changes were detectable about half the time. However, assuming that the estimated water storage changes were representative of progressively larger regions, the same monthly, seasonal, and annual changes were often detectable given a 200,000 km^2 or larger area. The rate of detectability increased and relative uncertainty decreased as spatial scale increased. Changes in soil moisture were typically the largest component of total water storage changes, followed by groundwater plus intermediate zone, snow water, and reservoir storage changes. Given additional information, soil moisture and groundwater plus intermediate zone storage changes have the best potential to be isolated from GRACE-derived total water storage changes in 300,000 km^2 or larger regions for monthly changes and 200,000 km^2 and larger regions for seasonal and annual changes.

Acknowledgments. This research was funded by NASA grant NGT5-30162. Additional support was provided by the Geology Foundation at the University of Texas. We would like to thank Srinivas Bettadpur at the University of Texas for providing estimates of GRACE instrument errors; Pat Yeh at MIT and Bryan Coulson, Ken Hlinka, and Jake Woolfson at ISWS for groundwater level data; Bill Saylor at ISWS for reservoir level data; the Illinois State Water Survey and the Global Soil Moisture Data Bank for soil moisture data; and the Midwestern Climate Information System for snow data.

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(Received April 3, 2000; revised September 11, 2000; accepted September 26, 2000.)