Estimating Changes in Terrestrial Water Storage

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Estimating Changes in Terrestrial Water Storage

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Dedication

For Mom, Dad, and most of all Jenna.

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Terrestrial water storage consists of groundwater, soil moisture, snow, ice, surface water, and water stored in vegetation. Its importance within the Earth system is evident, but as a singular component its interactive nature and quantitative characteristics are not well understood. Current techniques for estimating changes in terrestrial water storage include ground-based observation, microwave remote sensing, water balancing, and computer modeling.

The Gravity Recovery and Climate Experiment (GRACE), a satellite-based gravity mapping mission scheduled for a 2001 launch, will provide an alternative approach with distinct advantages. Redistribution of water is a primary source of variation in the gravity field over land. Therefore, it may be possible to estimate changes in terrestrial water storage based on GRACE gravity observations.

The objectives of this research were to characterize terrestrial water storage variability and to explore the methods of estimation while evaluating the potential of the GRACE technique. Three studies were performed. Results demonstrate that terrestrial water storage has a strong seasonal cycle and that year-to-year variations are also significant. Multiple caveats aside, soil moisture and groundwater storage typically exhibit the greatest variability among the components of terrestrial water storage.

Results also indicate that it will be possible to derive water storage variations from GRACE gravity observations on monthly and longer time steps. The primary limiting factors will be the desired spatial scale, which will be related inversely to GRACE instrument errors, and the magnitude of the variations themselves. Generally a region must be larger than 200,000 km^2 for the errors to be small enough to permit a meaningful estimate. Secondary limiting factors will be temporal resolution and uncertainty in the simulated fields used to remove the effect of atmospheric mass redistribution from the observed gravity signal. For a sufficiently large region, total uncertainty may be as little as 3.5 *mm* (equivalent height of water) on a monthly time step. Terrestrial water storage changes typically average 10 to 25 *mm* per month, depending on the region. Given auxiliary information from observations or models, it will be possible to isolate specific components, including changes in groundwater storage, from GRACE-derived total water storage variations.

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Chapter 1: Introduction

Terrestrial water storage is composed of all of the stocks of water that exist within or upon the continental crust. These include groundwater, soil moisture, lakes, streams, snow, ice, and vegetative water storage.

Terrestrial water storage affects weather and climate, enables life on land, and dictates patterns of human civilization. In particular, groundwater provides for irrigation, industry, and domestic usage and maintains streamflow between storms; soil moisture sustains vegetation, influences atmospheric processes through its control of evapotranspiration, and creates thermal inertia at the land surface due to water's high heat capacity; surface water supports aquatic life, provides for human endeavors, and recharges aquifers; snow protects near-surface dwelling flora and fauna from sub-freezing temperatures during winters, fills streams with meltwater in the spring, and limits warming in alpine and polar regions by reflecting sunlight; and glaciers and ice caps store water for long periods as they balance sea level fluctuations and influence the global climate.

Despite its importance, terrestrial water storage is not adequately monitored at regional scales. Consequently, how it varies in time is not well understood, nor are its interactions with other components of the Earth system. In fact, hydrologists often assume that, over the course of one or more years, a net zero change in terrestrial water storage should occur.

Ground-based observation, remote sensing, water balancing, and computer modeling are the contemporary approaches for estimating changes in terrestrial

water storage. Typically, ground-based observation is accomplished by groundwater well monitoring, manual or automated soil moisture sampling by the gravimetric technique, capacitance or heat dissipation sensor, or neutron probe, and reading depths from snow stakes. These methods are labor intensive and susceptible to human error. Furthermore, because terrestrial water, particularly soil moisture, often exhibits a high degree of spatial variability, point measurements are not reliable predictors of regional water storage changes. Satellite-based remote sensing is promising as a future source of water storage observations, but currently mature instruments, such as passive C and L band microwave sensors, can discern only the moisture stored in the upper few centimeters of soil, and the retrieval algorithms are problematic in densely vegetated regions. Another approach is a terrestrial water balance, for example,

$$\Delta S = -R - (E - P), \tag{1.1}$$

where ΔS is the change in terrestrial water storage, R is runoff, E is evapotranspiration, and P is precipitation. This equation is limited by the evapotranspiration term, which is a significant source of uncertainty. A combined atmospheric-terrestrial water balance does not require an estimate of evapotranspiration:

$$\Delta S = -Q_{net} - R, \tag{1.2}$$

where Q_{net} is the net runoff of atmospheric water vapor. However, the atmospheric data must be generated by a model which may or may not resemble the real world. The physical processes that effect changes in terrestrial water storage can be modeled, and as computers continue to become faster and more

powerful and the models more complex, model-based estimates will improve. Nevertheless, at present, more observational data are needed to constrain the models and to verify their accuracy.

The purpose of this body of work was to explore the state of terrestrial water storage change estimation while evaluating the potential of a recently proposed, observation-based technique for estimating changes in terrestrial water storage. The technique will rely on satellite-based measurements of the Earth's time-dependent gravity field, which will be provided by the Gravity Recovery and Climate Experiment (GRACE). GRACE is sponsored jointly by the National Aeronautics and Space Administration (NASA) and the Deutsches Zentrum für Luft-und Raumfahrt. The mission's strategy is to launch two satellites into a tandem, near-polar orbit, 170 to 270 km apart, at an initial altitude of 480 km. For five years a microwave tracking system will measure precisely the distance, or "range", between the two satellites, which will vary because the shape of the Earth's gravity field is not uniform in space or time. These range variations will be the basis for producing a new model of the global gravity field every thirty days. Because water storage redistribution is a significant source of temporal variation in the gravity field over land, it will be possible to identify water storage changes based on two or more GRACE products.

Three separate studies, described in Chapters 2-4, were performed in order to assess the GRACE technique by examining and manipulating time series of terrestrial water storage changes and estimating the uncertainty that would have been involved in deriving those changes from gravity measurements made by

GRACE. The first study (Chapter 2) utilized modeled time series of soil moisture and snow to assess the potential of the GRACE technique in twenty regions around the world, which were chosen to encompass a broad range of climates and spatial scales. In the second study (Chapter 3), the relative importance of the components of terrestrial water storage was examined by comparing observations of groundwater, soil moisture, snow, and reservoir water storage. The detectability by GRACE of variations in the sum of these components was then determined. The third study (Chapter 4) focused on the potential to estimate changes in a single component of terrestrial water storage, groundwater, in the High Plains aquifer region of the central United States, based on future GRACE observations and additional data. Each of Chapters 2-4 is derived from an article that has been published or submitted for publication.

Chapter 5 presents a summary of the research and some conclusions. In general, monthly, seasonal, annual, and multi-annual variations in terrestrial water storage will be detectable by GRACE, mainly depending on the area of the region, to which errors will be inversely related, and the magnitudes of the variations themselves. Soil moisture and groundwater storage variations are typically the greatest contributors to terrestrial water storage variations. Snow is likely to be important in high latitude and alpine regions and surface water may be significant under certain conditions, but these two notions were not tested. Seasonal changes in vegetative water storage were assumed to be negligible, although they were not quantified. Given additional information, isolation of changes in these components will be possible. A combination of auxiliary

monitoring and model assimilation is recommended in order to decompose GRACE-derived terrestrial water storage changes and thereby to maximize the usefulness of the data.

Chapter 2: Detectability of Variations in Continental Water Storage from Satellite Observations of the Time Dependent Gravity Field

SYNOPSIS

Continental water storage is a key variable in the Earth system that has never been adequately monitored globally. Since variations in water storage on land affect the time-dependent component of Earth's gravity field, the NASA GRACE satellite mission, which will accurately map the gravity field at two to four week intervals, may soon provide global data on temporal changes in continental water storage. This study characterizes water storage changes in twenty drainage basins ranging in size from 130,000 km^2 to 5,782,000 km^2, and uses estimates of uncertainty in the GRACE technique to determine in which basins water storage changes may be detectable by GRACE and how this detectability may vary in space and time. Results indicate that GRACE will likely detect changes in water storage in most of the basins on monthly or longer time steps, and that instrument errors, atmospheric modeling errors, and the magnitude of the variations themselves will be the primary controls on the relative accuracy of the GRACE-derived estimates.

Portions of this chapter have been previously published as *Rodell and Famiglietti* [2000].

INTRODUCTION

Soil moisture, groundwater, snow and ice, lake and river water, and vegetative water are the principal components of continental (or terrestrial, total) water storage. Although it constitutes only about 3.5% of the water in the hydrologic cycle, continental water storage has a tremendous influence on climate and weather as well as being fundamental to life on land.

The importance of soil water in the Earth's climate system has been demonstrated using general circulation models (GCMs) and documented by numerous authors [e.g., *Manabe*, 1969; *Shukla and Mintz*, 1982; *Milly and Dunne*, 1994; *Dirmeyer*, 1995; see also *Entekhabi et al.*, 1996 for a recent review]. For example, soil water links the water, energy, and biogeochemical cycles; its high heat capacity provides thermal inertia over multiple time scales; and it fuels evapotranspiration which helps to sustain storms through the process of precipitation recycling [*Brubaker et al.*, 1993; *Eltahir and Bras*, 1996].

Frozen water and liquid water stored below soils in aquifers play important roles in the Earth system as well as holding practical significance for society. Seasonal melts replenish soils and streams, while groundwater provides baseflow to streams and sustains deep rooted plants through periods of drought. Predicting the magnitudes of spring melts and the availability of groundwater is critical for natural hazard preparedness and for agricultural and domestic water resources management.

Mass redistribution associated with changes in water storage on land has additional effects on the Earth system beyond those described above. For example, *Chao and O'Connor* [1988] and *Kuehne and Wilson* [1991] showed that changes in terrestrial water storage effect Earth rotation variations; and *Chen et al.* [1998] showed that the redistribution of water from the continents to the oceans is the primary driver of sea level variations.

The benefits that would result from improved monitoring and understanding of variations in continental water storage are evident. However, a system for routinely monitoring changes in terrestrial water storage is not yet in place. Ground-based techniques are labor intensive and provide only point estimates of water storage [e.g., *Famiglietti et al.*, 1998]. Microwave remote sensing of soil moisture by satellite shows promise [e.g., *Jackson and LeVine*, 1996], but it will not provide information on deeper soil moisture or groundwater. Models provide good spatial coverage but their utility is limited by the quantity and quality of data available for input and validation.

The current shortage of large scale terrestrial water storage observations may be remedied in 2001 with the launch of the NASA GRACE (Gravity Recovery and Climate Experiment) satellite mission. The goal of GRACE is to measure the Earth's gravity field with unprecedented accuracy at two to four week intervals [*Tapley*, 1997]. Because mass movements of water at and below the Earth's surface affect the time dependent component of the gravity field, it is hypothesized that satellite-based gravity measurements obtained by GRACE can be manipulated to produce estimates of changes in water storage for given

terrestrial regions [*Dickey et al.*, 1997; *Wahr et al.*, 1998]. Estimates of the absolute magnitude of water storage would not be obtainable, and the technique would integrate changes in soil moisture, snow, groundwater, and the other components of continental water storage to produce an estimate of the net change in total continental water storage. In other words, GRACE will provide the spatial distribution of ΔS, the change in total water storage, during the time period ΔT. Note that in this research, the terms "variations in water storage" and "changes in water storage" are used interchangeably to denote ΔS.

This chapter investigates the potential of GRACE to provide estimates of continental water storage variations by comparing simulated fields of water storage changes to the expected accuracy of GRACE-derived ΔS measurements. The analysis was conducted at various spatial and temporal scales, and over different climatic regions of the world, and its purpose was to determine the limits of detection of the gravity-based technique. Toward that end, twenty continental scale drainage basins were identified for detailed study. Mean changes in water storage were computed on monthly, seasonal, and annual bases for each basin using twelve modeled data sets. Uncertainties in hypothetical GRACE-derived water storage variation measurements were estimated as the sum of three individual error sources. The results presented in this chapter have implications for how the detectability of water storage variations by GRACE and the level of precision of GRACE-derived ΔS estimates change with geographic location and with the time period ΔT.

BACKGROUND

The gravitational force experienced at the surface of the Earth varies in space and time, so that the gravity field of the whole Earth can be visualized as a not-quite-smooth ellipsoid. Spatial variations in the Earth's gravity field arise primarily from irregularities in the mass distribution near the surface of the Earth (e.g., continents, mountains, and depressions in the crust). The Earth's gravity field is said to have both static and time variable components, the static component being orders of magnitude stronger and encompassing all the factors that only vary on geologic time scales, e.g., the total mass of the Earth and the distribution of the continents. *Jeffreys* [1952] was among the first to report the existence of the time variable component, noting that mass movements such as ocean tides could effect temporal changes in the gravity field. At that time, spatial variations in the gravity field were observed primarily with the aid of pendulums.

Mapping spatial irregularities in the Earth's gravity field was facilitated by the first artificial satellites, which began orbiting in the late 1950s. Satellite tracking via optical and Doppler techniques allowed scientists to compute departures from predicted orbits, and these departures were attributed to previously unobserved factors affecting the paths of the satellites, irregularities in the static gravity field in particular. For the past two decades, orbit determination has been accomplished by ground to satellite laser ranging. The increased accuracy afforded by this technique has allowed more detailed assessments of the gravity field.

Yoder et al. [1983] reported that the orbit of the Lageos satellite was sensitive to temporal variations in the gravity field in addition to static, spatial variations. They believed that the primary sources of these temporal variations were redistribution of groundwater and air mass and changes in sea level. *Gutierrez and Wilson* [1987] computed perturbations in the orbits of the Lageos and Starlette satellites caused by seasonal redistribution of air mass and terrestrial water storage. To characterize the changes in water storage, they divided the land surface, neglecting Antarctica, into approximately 600 drainage basins on a $1° \times 1°$ grid, and performed a simple water balance for each basin, based upon archived precipitation data and published coefficients for runoff and evapotranspiration. By comparing their solutions with the observed orbits, they were able to confirm that seasonal variations in terrestrial water storage do influence the time variable gravity field enough to induce predictable satellite orbit perturbations. *Chao and O'Connor* [1988] reached a similar conclusion in their study of the effects of seasonal changes in surface water on the Earth's rotation, length of day, and gravity field. More recent studies have attempted to define the global gravity field and its fluctuations more precisely [e.g., *Ries et al.*, 1989; *Nerem et al.*, 1993], but significant future improvements demand the utilization of advanced technologies.

Recently, *Dickey et al.* [1997] assessed several possible mission scenarios for a dedicated satellite gravity mission and highlighted the potential benefits to research in hydrology (i.e., quantifying changes in continental water storage using gravity field observations), oceanography, and solid earth processes, as well as

geodesy. Mission scenarios were evaluated based on tradeoffs between expected mission lifetime at a particular altitude, measurement resolution, and maturity of the required technologies.

Dickey et al. [1997] helped to crystallize the plans that soon became the Gravity Recovery and Climate Experiment. GRACE is sponsored jointly by NASA, as part of its Earth System Science Pathfinder Project, and two German agencies: GeoForschungsZentrum Potsdam and Deutsche Forschungsanstalt fur Luft und Raumfahrt. Mission design is being led by scientific teams at the University of Texas Center for Space Research and NASA's Jet Propulsion Laboratory. GRACE is scheduled for launch in July, 2001 with a nominal duration of five years. The instrument will consist of two satellites in a tandem orbit 100-400 *km* apart at about 450 *km* altitude. Microwave tracking integrated with Global Positioning System receivers will measure range perturbations between the satellites caused by spatial variations in the gravity field, while nongravitational accelerations will be monitored by onboard accelerometers [*Tapley*, 1997]. The high degree of precision afforded by this technique will enable GRACE to map the gravity field at intervals of two weeks or longer with an accuracy equivalent to a few millimeters of water (n.b. 1 *mm* water depth is equivalent to 1 kg/m^2 water mass, given a constant density of 1 g/cm^2).

In order to use GRACE gravitational measurements to estimate changes in continental water storage, certain operations will be necessary. The following describes these operations in a very general manner (see *Wahr et al.* [1998] for a more detailed explanation). For a particular averaging period (the time period

during which GRACE observations contributing to a single global gravity field are gathered), GRACE observations will be used to estimate thousands of coefficients of a spherical harmonic expansion that describes Earth's total (static plus time variable) gravity field as the shape of a geoid. To estimate changes in water storage, a global field of temporal gravitational variations must first be computed from two such total gravity fields. Postglacial rebound (PGR) and changes in the distribution of atmospheric mass also will influence the time dependent gravity signal over land, so that models of surface pressure changes and PGR must supply the data necessary to remove their effects. On monthly to annual time scales other factors influencing the time variable gravity signal over inland regions (e.g., the solid Earth tide) are assumed to be negligible. For a specific region, the spherical harmonic expansion that describes the corrected global field of temporal gravitational variations will be manipulated mathematically to determine the change in mass (or equivalent height of water) corresponding to the change in terrestrial water storage in the region. It should be noted that redistribution of oceanic mass will contribute to the time dependent component of the gravity signal in coastal regions. *Wahr et al.* [1998] developed a technique for minimizing this "leakage" by first using the GRACE gravity fields to solve for the oceanic mass changes and then removing these from the original fields. Because the basins considered by this study exist largely in the interiors of the continents, it is assumed that oceanic effects will be inconsequential. However, the technique of Wahr et al. [1998] should significantly reduce the effects in basins where substantial oceanic leakage is possible.

Dickey et al. [1997] included a promising forecast of the potential resolution and accuracy with which variations in terrestrial water storage could be derived from a dedicated gravity mission. *Wahr et al.* [1998] continued this investigation from a geophysical perspective. Using modeled global time series of soil moisture, snow, atmospheric pressure, PGR, and sea level, they constructed synthetic total gravity fields. Hypothetical satellite observations, which incorporated estimates of instrument noise and the errors associated with removing the effects of PGR, atmospheric mass variations, and sea level changes from the signal, were produced from the synthetic gravity fields, and time series of water storage were extracted from these "observations". The water storage time series were then compared to the original data. Results were analyzed for circular regions with radii of 100 *km* to 10,000 *km* at locations centered in the Amazon and Mississippi river basins. For a monthly averaging period accuracy was on the order of 26 *mm* at 230 *km* and 2 *mm* at 500 *km*.

This study advances the investigation of GRACE's potential to produce accurate estimates of changes in continental water storage in three ways. First, the analysis is global, with 20 regions of varying climate and spatial extent examined individually. Second, to improve understanding of the nature of terrestrial water storage variations themselves, twelve global time series are assessed with a critical eye towards the realism of the estimates, with results from five independent water balance studies included for comparison. Third, GRACE's potential is evaluated for three time scales commonly encountered in hydroclimatology: 30 days, 90 days, and 365 days.

METHODS

Study Areas

Twenty continental scale drainage basins were selected for detailed study. These are shown in Figure 2.1 and listed in Table 2.1 from largest to smallest drainage area (this is the order used in most of the subsequent figures and tables, all of which are located at the end of Chapter 2). Drainage basins, rather than grid squares or circular regions, were chosen as the units over which to analyze GRACE's potential to measure variations in terrestrial water storage because they are fundamental subdivisions of the land surface from a hydrologic perspective. Furthermore, the future use of GRACE data to compute changes in storage over watersheds will facilitate comparisons to runoff measurements. The twenty basins were chosen to encompass an array of climates, from wet equatorial (Amazon) to dry subtropical (Murray-Darling) to boreal/tundra (Ob). Also desirable was the existence of Global Energy and Water Cycle Experiment (GEWEX) projects and other large scale experiments in the regions (e.g., in the Mississippi, Mackenzie, and Amazon basins). The basins range in size from 5,782,000 km^2 to 130,000 km^2 and include four internally draining basins. They were delineated with the aid of a geographic information system using the Terrain Base 5' global digital elevation model [*Row et al.*, 1995].

Water Storage Data

Twelve modeled data series of soil moisture and snow water were acquired. Ten series were modeled by Global Soil Wetness Project (GSWP)

contributing groups on 1° global grids, spanning the 24 month period beginning 1 January, 1987 [*IGPO*, 1998]. GSWP is an ongoing GEWEX project that serves as a pilot study of the feasibility of producing a global data set of soil wetness for use in global climate models. Ten independent groups used a common data set of model parameters, meteorological observations, and analyses to drive state-of-the-art land surface models, which produced fields of soil moisture, snow, and other variables. The ten GSWP contributing groups and their models are listed in Table 2.2. The two other series are products of the European Centre for Medium-Range Weather Forecasts (ECMWF) Re-analysis [*ECMWF*, 1996] and the joint National Center for Environmental Prediction (NCEP) and National Center for Atmospheric Research (NCAR) Reanalysis Project [*Kalnay et al.*, 1996]. Respectively, the two reanalysis products span the years 1979-93 and 1974-96 and have approximately 1.875° and 1.125° spatial resolutions. These twelve modeled data sets were chosen because they contain some of the only terrestrial water storage estimates that are currently available with global coverage. Time series of water storage from each of the twelve models were plotted and compared to each other and to independently derived water storage data, which resulted in the exclusion of the two reanalysis time series from further consideration in this study. Results of the analysis are described in the "Results" section under the heading "Terrestrial Water Storage".

Water storage changes, ΔS, were derived over specific intervals, ΔT (i.e., the time step), for each of the time series of water storage, S, described above. This was necessary because GRACE will not measure the total magnitude of

water storage in the land at any given time; rather, GRACE will allow changes in water storage over specific intervals to be determined by comparing gravity fields measured over different averaging periods. Note that when time series of water storage changes are derived by comparing GRACE measurements from consecutive averaging periods, the measurement averaging period and the series time step, ΔT, will be equivalent in length; therefore, in this research the descriptors "monthly", "seasonal", and "annual" when modifying "change" or "error" will refer to both the averaging period and the time step.

The following approach was used to construct time series of water storage changes from the modeled data. First, because GRACE will not be able to differentiate between soil water and snow water, the two were combined to represent total terrestrial water storage. Second, changes in storage were computed as the backward difference between average terrestrial water storage for each 30-, 90-, or 365-day period corresponding to a GRACE measurement averaging period, and average terrestrial water storage for the immediately preceding period, for the duration of each modeled time series and for each of the basins. This approach was designed to be consistent with the manner in which GRACE will measure water storage changes. The water storage change time series were subsequently compared to the corresponding uncertainty in the GRACE technique, which is described below.

Sources of Uncertainty

Three sources will contribute errors to the GRACE-derived water storage variation fields. The limitations of instrumental and computational precision will

produce one type of error. Two other sources of error will arise from the problems of removing the effects of atmospheric mass redistribution and PGR from the gravity signal. The global fields required for these corrections must be modeled because observations of surface pressure and PGR rates are not continuous in time and space. Therefore errors in the models will impact the water storage change estimates.

Uncertainty Estimation

The GRACE science team provided a table of total "instrument uncertainty" versus the effective radius of a given region for a thirty-day averaging period (*S. Bettadpur*, The University of Texas, 1998, personal communication). Instrument uncertainty accounts for the combined effects of errors in the orbital parameters, microwave ranging measurements, accelerometer measurements, and error in the ultrastable oscillator [*Dickey et al.*, 1997]. It is inversely related to both the length of the measurement averaging period and the length scale of the region (see Figure 2.2), although estimates over long, narrow regions may be less accurate than estimates over more equiproportional regions. This inverse relationship between spatial and temporal resolution and instrument errors is certain, but the error estimates themselves may change before GRACE is launched because the mission specifications are still evolving. From the original table, uncertainty tables for seasonal (90-day) and annual (365-day) averaging periods were produced by using the following:

$$E_{I(N)} = E_{I(30)} \Big/ \sqrt{N/30}\,, \tag{2.1}$$

where $E_{I(N)}$ is the instrument uncertainty for an averaging period of N days. The instrument uncertainty in a single GRACE gravity field then was estimated for each drainage basin based on its area. Recalling from "Water Storage Data" that two GRACE measurements will be required to determine a temporal variation in the gravity field, the following relation was used to calculate instrument errors, E_I, in an estimate of the change in water storage:

$$E_I = \sqrt{E_{I,1}^2 + E_{I,2}^2}\,, \tag{2.2}$$

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, respectively. This relationship is appropriate because GRACE instrument errors are expected to be uncorrelated in time (*S. Bettadpur*, The University of Texas, 1998, personal communication).

Modeled atmospheric surface pressure fields will be used to remove the effect of atmospheric mass changes from the time variable gravity signal, because the atmospheric mass per unit area at a point on the Earth's surface is essentially equal to the surface pressure divided by the acceleration due to gravity. Therefore atmospheric pressure data were obtained from two sources, the ECMWF Re-analysis and the NCEP/NCAR Reanalysis, in order to estimate atmospheric errors, $E_{A,i}$. Both data sets are globally gridded at 2.5° resolution. Monthly, seasonal, and annual errors for each drainage basin were computed as in *Wahr et al.* [1998], as

$$E_{A,i} = \frac{\left|\overline{P}_{ECMWF} - \overline{P}_{NCEP/NCAR}\right|}{\sqrt{2}}, \tag{2.3}$$

where $\overline{P}$ 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 $(\overline{P}_{ECMWF} - \overline{P}_{NCEP/NCAR})$, 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, as

$$E_A = \sqrt{E_{A,1}^2 + E_{A,2}^2}\,, \tag{2.4}$$

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. While temporal correlation in the atmospheric errors may exist, it is assumed to be small (*K. Trenberth*, NCAR, 1998, personal communication). Equation (2.4) results in an estimate of the atmospheric uncertainty that is conservative in that it does not account for future improvements in the models.

Uncertainty resulting from the use of modeled PGR rates, E_{PGR}, was assessed as a uniform 20% error in the rebound rates of a PGR model [*Peltier*, 1994; *Peltier*, 1995], which were converted to equivalent units of water mass change. Total uncertainty in the change in storage, E_T, was taken as the sum of the three error components,

$$E_T = E_I + E_A + E_{PGR}\,. \tag{2.5}$$

RESULTS

Terrestrial Water Storage

Preliminary analysis showed that terrestrial water storage in the ECMWF data series is consistently and significantly smaller than that in the other eleven series because of low soil moisture values. Investigation into this matter revealed

that the ECMWF model does not allow soil moisture in the deeper layers to vary above wilting point values even during episodes of flooding, as in the Midwestern U.S. in 1993. Because of this unrealistic constraint, water storage data from ECMWF were omitted from this analysis.

In comparison, the NCEP/NCAR Reanalysis allows soil moisture to vary over a larger, more realistic range. However, examination of this time series showed that only minor variations in the amplitude and timing of the annual cycle exist, resulting from a dependence on a prescribed climatology. In the NCEP/NCAR model, soil moisture is nudged towards the *Mintz and Serafini* [1992] climatology, which is not governed by the model water and energy balances; and additionally, no observations which directly affect soil moisture are assimilated into the model [*Kalnay et al.*, 1996]. Therefore the NCEP/NCAR water storage data were also eliminated from this analysis.

Terrestrial water storage data produced by the ten GSWP models appear to be more realistic than the reanalysis data. Figure 2.3 shows the two year time series of water storage, *S*, for each of the GSWP models and each of the twenty drainage basins. Note that in each time series *S* is shown as a departure from its two-year mean, which was arbitrarily set to zero. The amplitudes of the variations and the differences between the two years are seemingly reasonable. All ten GSWP model data series display similar characteristics because all ten models used the *ISLSCP Initiative I Global Data Sets for Land-Atmosphere Models, 1987-88* [*Meeson et al.*, 1995, and *Sellers et al.*, 1995] for input, and all but the CCSR model used the same pixel-specific soil depths, which are

summarized in Table 2.3. The soil depths averaged about two meters, which indicates that variations in deeper soil water storage are not reflected in the GSWP data. Contrasting the time series in Figure 2.3, the amplitude of the strongest modeled annual cycle of water storage is typically not more than twice the amplitude of the weakest annual cycle. Lag time between the water storage peaks of different GSWP models is rarely more than a month and in general the phase differences are small. The similarity of the ten data sets is an encouraging validation of the GSWP results.

Additional information on water storage variations was derived from five terrestrial and combined atmospheric-terrestrial water balance studies [*Matsuyama*, 1995; *Oki et al.*, 1995; *Rasmusson*, 1968; *Roads*, 1994; *Ropelewski and Yarosh*, 1997, 1988] in order to compare the GSWP time series to independent, observation-based results and thereby justify their use in this study. Estimates of the amplitude of the mean annual cycle of water storage from these sources are summarized and compared to mean GSWP amplitudes in Table 2.4. Amplitude was computed as half the difference between the maximum and minimum storage values in the mean annual cycle. A limitation of these comparisons is that the drainage basins were not delineated in exactly the same way by the various authors. Furthermore, the mean annual cycles are often based on data from incongruent time periods (Table 4). Nevertheless, the amplitudes of the storage cycles determined by these water balances should be, for our purposes, reasonable approximations of the actual amplitudes. Note that only *Oki et al.* [1995] extended their study beyond North America.

The comparisons seen in Table 2.4 indicate that the GSWP time series are similar to the water balance estimates. In the Mississippi basin the GSWP amplitude of the annual cycle is 36 *mm*, which is larger than the smallest water balance estimate (29 *mm*) and just less than half of the largest estimate (78 *mm*), with the median estimate being 52 *mm*. That the GSWP amplitude is at the low end of the range may be explained by two factors: (1) it only reflects surface soil moisture and snow variations, while results from the independent water balance studies implicitly include deeper soil water, groundwater, vegetative water, and lake and river channel storage variations; and (2) the GSWP years, 1987 and 1988, were dry in terms of precipitation for the Mississippi basin, which may have caused storage variations to be subdued.

Comparing the estimated amplitudes of *Oki et al.* [1995] to those from GSWP in 17 other basins shows a similar result. GSWP amplitudes are generally smaller, but rarely by more than 50%. From this analysis it seems reasonable to conclude that actual variations in total terrestrial water storage may be underestimated to some extent by GSWP modeled soil moisture and snow water variations, but these estimates would be of the correct order of magnitude. Further, the use of GSWP estimates in this study is well justified: when the estimated errors derived in "Uncertainty Estimation" are imposed upon GSWP estimates of changes in water storage, use of these lower amplitude storage estimates may provide a conservative, upper bound on relative error.

Table 4 also compares the average GSWP estimate of the annual change in continental water storage over the Mississippi River basin to estimates of the

annual change from the water balance studies. Six other estimates are shown. An annual change is defined here as the difference between two consecutive values of mean annual water storage. The GSWP annual variation, 18 *mm*, is smaller than all of the water balance estimates, which range from 23 *mm* to 58 *mm*, but it is within the same order of magnitude. As before it is speculated that the GSWP value is smaller because it only includes variations in surface soil and snow water storage and because it was computed as the change between two drier-than-normal years. A longer time series would be more appropriate for determining a representative value of the annual variation. Lacking other estimates for comparison, it is difficult to draw conclusions about the GSWP annual variations seen in the other basins. Considering that these variations do not include the effects of deeper soil moisture and groundwater, it is hypothesized that they are low, order-of-magnitude estimates of typical annual variations, although some may be substantially smaller than the true mean variations.

Potential Accuracy of GRACE-Derived ΔS Estimates

As described in "Sources of Uncertainty", uncertainty in the GRACE-derived estimates of water storage changes will arise from errors inherent to the instrument and errors in the modeled fields of atmospheric mass variations and postglacial rebound. The average contributions from these sources are shown in Table 2.5 for the twenty basins on monthly, seasonal, and annual time scales. As smaller and smaller basins are considered, between the Columbia and Tibetan Plateau regions monthly total error increases abruptly from 7.50 *mm* to 22.96 *mm*, with similarly abrupt increases in seasonal and annual total errors. Total errors in

the largest fifteen basins are considerably smaller than in the Tibetan Plateau, Chao Phraya, Wisla, Great Salt Lake, and Odra basins. This is attributed to increased instrument errors in the latter four basins and unusually large atmospheric errors in the Tibetan Plateau region.

Figure 2.4 plots monthly changes in water storage for each of the twenty drainage basins, computed as described in "Water Storage Data". Each change in storage value shown is the median from the ten GSWP models. Each error bar represents $\pm E_T$, the total uncertainty in a hypothetical GRACE-derived estimate of the change in water storage for the particular basin and month to month period, computed using (2.5). The changes and uncertainty values plotted in Figure 2.4 are summarized in Table 2.6. Mean uncertainty is less than 25% of the mean change in storage in the Amazon, Ob, Volga, and Ganges drainage basins; mean uncertainty is 25-50% of the mean change in storage in the Parana-Uruguay, Mississippi, Niger, Lena, Mackenzie, Murray-Darling, and Columbia basins; mean uncertainty is 50-100% of the mean change in storage in the Zaire, Lake Chad, Huang He, Oranje, and Chao Phraya basins; and water storage variations in the Tibetan Plateau, Wisla, Great Salt Lake, and Odra basins are likely to be smaller, on average, than mean uncertainty (i.e., undetectable). These results will be evaluated further in the "Discussion" section.

Seasonal changes in terrestrial water storage in the twenty basins are plotted in Figure 2.5, with an error bar depicting the uncertainty in the GRACE estimate for each season-to-season period. These data are also summarized in Table 2.6. Thirteen of the twenty basins have a mean uncertainty that is less than

25% of the mean change in storage; of these, mean uncertainty in the Amazon, Parana-Uruguay, Ob, Mackenzie, Volga, and Ganges basins is less than 10% of the mean water storage change. Mean uncertainty in the Zaire, Huang He, Chao Phraya, and Wisla is 25-60% of the mean water storage change. Only in the Odra, Great Salt Lake, and Tibetan Plateau regions are water storage variations expected to be undetectable.

The median GSWP annual changes in terrestrial water storage from 1987 to 1988 are shown in Figure 2.6, with error bars representing uncertainty in the GRACE estimates. These data are also summarized in Table 2.6. Owing to the relatively small errors for annual averaging periods, the mean annual uncertainty is less than 50% of the mean water storage variation for more than half of the twenty basins. The data also imply that annual variations in water storage will be undetectable by GRACE in seven of the basins including the Amazon. However, these water storage change estimates may not be reliable, because in addition to neglecting deeper groundwater they result from a single year-to-year cycle.

DISCUSSION

Table 7 lists the number of time intervals during which water storage changes are detectable by GRACE (relative error, $E_T/\Delta S < 1$) within each basin for monthly, seasonal, and annual averaging periods, given the data in Figures 2.4, 2.5, and 2.6. The table shows that monthly water storage variations are detectable between 50% and 91% of the time in 15 of the 17 basins of area 201,000 km^2 and greater; that seasonal variations are detectable between 50% and

100% of the time in 17 of the 18 basins of area 184,000 km^2 and greater; and that annual variations are detectable in 13 of the 17 basins of area 201,000 km^2 and greater. Close inspection of Table 2.6 reveals that when mean error is compared to mean change in storage, a similar pattern emerges: water storage variations are detectable in 16, 17, and 13 basins on monthly, seasonal, and annual time scales, respectively.

Relative error (uncertainty), and therefore detectability, in the GRACE-derived estimates will depend mainly on the size of the region (which is inversely related to the instrument uncertainty), the atmospheric modeling errors, and the magnitude of the variations themselves. The temporal resolution of the change in storage estimates will also affect relative uncertainty. Monthly total uncertainty is always larger than seasonal uncertainty, which in turn is larger than annual uncertainty, because the instrument and atmospheric errors decrease with increasing averaging period (long-term average modeled pressure estimates should approach measured mean values). PGR errors show the opposite trend, increasing linearly with time scale, but as seen in Table 2.5, uncertainty due to PGR is typically two orders of magnitude smaller than total uncertainty in the regions of this study.

In the Wisla, Great Salt Lake, and Odra basins, where water storage changes are expected to be undetectable (i.e., $\Delta S \leq E_T$), instrument errors will be large and will tend to dominate E_T because of the small size of the basins. Instrument errors also will dominate in the fourth smallest basin, the Chao Phraya, and will nearly obscure the water storage changes. In contrast, atmospheric errors

will dominate E_T in the other sixteen basins, which are more spatially extensive, but total uncertainty will not be large enough to obscure the water storage variations. Exceptions occur in the Amazon, the Huang He, and Columbia basins, where the estimated annual variations are smaller than the uncertainty (but recall that these estimates are derived from a single year-to-year cycle), and also in the Tibetan Plateau region, where the variations themselves appear to be small on all three time scales. From an examination of the atmospheric errors in Table 2.5, which bear no relationship to basin size, and the instrument error versus spatial scale relation shown in Figure 2.2, it can be concluded that, in general, instrument errors are small and atmospheric errors, though small themselves, will dominate E_T in regions larger than about 400,000 km^2. Instrument errors increase as smaller regions are considered and dominate E_T in regions smaller than 200,000 km^2. These factors combined with the magnitude of ΔS will determine the detectability and relative error of water storage variations by GRACE. Estimates of seasonal water storage changes will have the lowest relative error primarily because seasonal changes in terrestrial water storage are significantly larger on average than monthly or annual changes.

Since atmospheric errors dominate E_T and, hence, significantly impact relative error in most of the basins considered here, it is worth further exploring their characteristics. The accuracy of atmospheric models is not spatially or temporally constant, so that the atmospheric component of total error will vary with time and location. The average of the monthly atmospheric errors, $E_{A,i}$, as computed by (2.3), is shown in Figure 2.7 for each of the twenty basins. Note that

the basins are ordered so that average uncertainty increases from front to back. Recall that $E_{A,i}$ is the error in a single GRACE measurement of the gravity field; given two such values, (2.4) would be used to compute the atmospheric error, E_A, in an estimate of water storage change. The monthly errors range from a low of 0.5 *mm* in July in the Mississippi basin to a high of nearly 38 *mm* in November in the Tibetan Plateau region. Larger errors stem from a deficiency of surface, upper air, and satellite observations, so that fewer measurements are available for model assimilation and developing an understanding of regional climatology, although the way the two models handle topography is different in some regions, which also affects the atmospheric error estimates. Figure 2.7 also demonstrates that the level of uncertainty in the GRACE estimates due to atmospheric errors will not necessarily be constant for a particular region throughout the course of a year. There exists a seasonality that is stronger in some regions than in others. This seasonality can clearly be seen in Figure 2.7 in more than half of the basins, including the Ob, Niger, and Chao Phraya.

Since the magnitude of ΔS is another important aspect of relative error, Figure 2.8 provides a more spatially continuous perspective of the range of water storage variations across the continents by mapping the amplitude of the mean annual cycle of water storage at 1° grid resolution. The map was generated using data from one GSWP model, that of the Japan Meteorological Agency (JMA), which was consistently close to the median of the GSWP models in the data series. Comparison of Figure 2.8 with the instrument error curve in Figure 2.2 and the range of atmospheric errors in Figure 2.7 and Table 2.5 gives a rough

indication of potential detectability in regions of the world not considered in this study.

Future circumstances, such as the evolution of the mission specifications, advances in technology, and improvements in atmospheric modeling capability, may influence the uncertainty in GRACE-derived water storage change estimates. Furthermore, the effects that hourly to daily water storage and atmospheric mass variations might have on the gravity measurements remains to be explored. It should be recalled that total terrestrial water storage variations may be significantly larger than those represented in the GSWP data, so that water storage variations may be detectable in more of the basins and at a higher level of relative accuracy than indicated by this study. The lack of a global scale groundwater fluctuation data set has precluded a more thorough analysis of variations in total terrestrial water storage and the potential for monitoring these variations. The water storage changes listed in Table 2.6 may have been quite different if groundwater fluctuations were taken into account. This inadequacy implicates the need for further study in this area, but it also exemplifies the shortcomings in our basic understanding of large scale hydrologic processes, some of which may be reconciled by GRACE.

SUMMARY

Global time series of continental water storage were obtained from twelve modeled data sets. Two sources were eliminated because they were not believed to be representative of actual conditions. The ten remaining data sets compared

favorably with independent water balance studies and were shown to provide conservative estimates of actual water storage variations. Based on these data sets, time series of changes in continental water storage were produced for twenty drainage basins ranging in size from 130,000 km^2 to 5,782,000 km^2 on monthly, seasonal, and annual time steps. The modeled changes in water storage were compared to the expected total uncertainty in GRACE-derived estimates of the changes. Estimated errors from the GRACE instruments, atmospheric modeling, and postglacial rebound modeling all contributed to the total uncertainty estimates. The primary controls on the relative accuracy of GRACE-derived water storage change estimates were determined to be instrument errors and therefore the area of the region, atmospheric modeling errors in the region, and the magnitude of the variations themselves. Monthly changes in continental water storage were predicted to be detectable on average in 16 of the 17 basins of area 201,000 km^2 and greater; seasonal changes were predicted to be detectable on average in 17 of the 18 basins of area 184,000 km^2 and greater; and annual changes were predicted to be detectable in 13 of the 17 basins of area 201,000 km^2 and greater.

Map Number	Drainage Basin	Area, km^2
1	Amazon	5,782,000
2	Zaire	3,788,000
3	Parana-Uruguay	3,375,000
4	Mississippi	3,166,000
5	Ob	3,143,000
6	Lake Chad	2,416,000
7	Niger	2,306,000
8	Lena	2,273,000
9	Mackenzie	1,933,000
10	Volga	1,290,000
11	Huang He	1,154,000
12	Murray-Darling	1,009,000
13	Ganges	997,000
14	Oranje	932,000
15	Columbia	817,000
16	Tibetan Plateau	416,000
17	Chao Phraya	201,000
18	Wisla	184,000
19	Great Salt Lake	142,000
20	Odra	130,000

Table 2.1: Twenty continental scale drainage basins.

Abbreviation	Model	Group/Agency
NCEP	ETA	NOAA/NCEP
CNRM	ISBA	Météo-France Centre National de Recherches Météorologiques
JMA	Modified SiB	Japan Meteorological Agency
MOS	Mosaic LSM	NASA/GSFC Hydrological Sciences Branch
MAPB	PLACE	NASA/GSFC Mesoscale Atmospheric Processes Branch
GSMW	Simplified SiB	NASA/GSMW Climate and Radiation Branch
COLA	Simplified SiB	Center for Ocean-Land-Atmosphere Studies
CCSR	Bucket (200 mm depth)	Center for Climate System Research, Univ. of Tokyo / National Institute for Environmental Studies
BATS	BATS	University of Arizona, Dept. of Hydrology and Water Resources
CSU	SiB2	Colorado State University Dept. of Atmospheric Science

Table 2.2: Global Soil Wetness Project models.

	Soil Depth, mm		
Drainage Basin	Minimum	Mean	Maximum
Amazon	1000	2339	6000
Zaire	1000	3104	8000
Parana-Uruguay	1000	2377	6000
Mississippi	1000	2205	3600
Ob	1000	2202	3600
Lake Chad	1000	1715	8000
Niger	1000	2261	8000
Lena	1000	1079	3600
Mackenzie	1000	1898	3600
Volga	1000	2337	3600
Huang He	1000	2181	3600
Murray-Darling	1000	1665	3600
Ganges	1000	1518	3600
Oranje	1000	1673	3600
Columbia	1000	2332	3600
Tibetan Plateau	1000	1676	5000
Chao Phraya	1000	2641	3600
Wisla	1000	1922	3600
Great Salt Lake	1040	2098	3600
Odra	1000	1670	3600

Table 2.3: Range of soil depths used in the GSWP models.

Drainage Basin	Annual Amplitude, mm								Annual Change, mm						
	A	**B**	**C**	**D**	**F**	**G**	**H**	**I**	**A**	**C**	**E**	**F**	**J**	**K**	**L**
Amazon	82	191							2						
Zaire	37	62							12						
Parana-Uruguay	45	30							21						
Mississippi	36	52	55	37	78	29	41	65	18	29	25	58	56	23	38
Ob	68	70							18						
Lake Chad	40	90							14						
Niger	62	156							19						
Lena	50	70							12						
Mackenzie	57	57							9						
Volga	80	91							22						
Huang He	19	43							1						
Murray-Darling	31	34							28						
Ganges	76	162							9						
Oranje	22	89							30						
Columbia	46	146							2						
Tibetan Plateau	5								5						
Chao Phraya	99	93							27						
Wisla	61	61							8						
Great Salt Lake	26								15						
Odra	61	75							28						

	Source	Averaging Period	Technique
A	GSWP (ten-model median)	1987-88	model
B	Oki et al. [1995]	1989-92	combined water balance
C	Roads [1994]	1984-88	combined water balance
D	Roads [1994]	1960-88	terrestrial water balance
E	Roads [1994]	1984-88	terrestrial water balance
F	Ropelewski and Yarosh [1997]	1973-88	combined water balance
G	Rasmusson [1968]	1958-63	combined water balance
H	Rasmusson [1968]	1958-63	terrestrial water balance
I	Rasmusson [1968]	1958-63	terrestrial water balance
J	Matsuyama et al. [1995]	1985-89	combined water balance
K	Matsuyama et al. [1995]	1989-92	combined water balance
L	Matsuyama et al. [1995]	1985-88	terrestrial water balance

Table 2.4: Comparison of estimates of water storage changes.

Two characteristics of the water storage cycle are compared: amplitude of the annual cycle; and annual change in storage.

Drainage Basin	Monthly Uncertainty, mm				Seasonal Uncertainty, mm				Annual Uncertainty, mm			
	E_I	E_A	E_{PGR}	E_T	E_I	E_A	E_{PGR}	E_T	E_I	E_A	E_{PGR}	E_T
Amazon	0.39	3.97	0.01	4.38	0.23	3.53	0.03	3.79	0.11	2.70	0.14	2.95
Zaire	0.49	9.33	0.01	9.83	0.28	9.26	0.04	9.58	0.14	9.26	0.14	9.55
Parana-Uruguay	0.52	2.96	0.01	3.49	0.30	2.70	0.03	3.03	0.15	2.38	0.14	2.66
Mississippi	0.54	3.16	0.00	3.70	0.31	2.78	0.00	3.09	0.15	2.08	0.02	2.25
Ob	0.54	3.24	0.01	3.79	0.31	2.70	0.02	3.04	0.15	2.12	0.09	2.36
Lake Chad	0.63	6.44	0.01	7.08	0.36	6.55	0.03	6.94	0.18	5.74	0.14	6.06
Niger	0.65	6.59	0.01	7.25	0.37	6.83	0.04	7.24	0.19	6.24	0.14	6.57
Lena	0.65	5.62	0.01	6.28	0.38	5.17	0.02	5.57	0.19	2.80	0.10	3.09
Mackenzie	0.72	4.38	0.02	5.12	0.42	3.44	0.05	3.91	0.21	1.83	0.20	2.23
Volga	0.95	4.79	0.01	5.74	0.55	4.80	0.02	5.36	0.27	4.42	0.07	4.76
Huang He	1.03	6.24	0.01	7.29	0.59	6.41	0.04	7.04	0.29	5.59	0.16	6.05
Murray-Darling	1.14	3.06	0.01	4.21	0.66	2.96	0.03	3.65	0.33	2.76	0.12	3.21
Ganges	1.15	4.82	0.01	5.99	0.67	3.87	0.04	4.58	0.33	2.06	0.17	2.56
Oranje	1.22	5.21	0.01	6.44	0.70	4.89	0.03	5.63	0.35	4.39	0.13	4.87
Columbia	1.37	6.13	0.00	7.50	0.79	5.68	0.00	6.47	0.39	3.45	0.01	3.85
Tibetan Plateau	3.17	19.77	0.01	22.96	1.83	18.77	0.04	20.64	0.91	13.41	0.14	14.46
Chao Phraya	20.08	9.43	0.02	29.53	11.60	9.51	0.05	21.15	5.76	9.19	0.20	15.16
Wisla	28.23	2.50	0.00	30.74	16.30	1.98	0.01	18.29	8.09	1.43	0.06	9.58
Great Salt Lake	87.70	12.46	0.01	100.16	50.63	12.20	0.02	62.85	25.14	11.82	0.07	37.03
Odra	131.26	2.27	0.00	133.54	75.79	1.73	0.01	77.53	37.63	1.14	0.05	38.83

Table 2.5: Mean Uncertainty from instrument, atmospheric, and PGR errors.

	Monthly						Seasonal						Annual			
	ΔS, mm			E_T, mm			ΔS, mm			E_T, mm			ΔS, mm	E_T, mm		
Drainage Basin	Min	Avg	Max	Min	Avg	Max	Min	Avg	Max	Min	Avg	Max		Min	Avg	Max
Amazon	0.9	26.1	53.8	0.8	4.4	11.7	54.9	74.5	93.4	0.5	3.8	9.5	1.9	0.8	2.9	6.3
Zaire	0.7	13.1	27.5	1.4	9.8	19.3	6.0	25.4	56.0	3.8	9.6	17.1	12.2	5.0	9.5	14.9
Parana-Uruguay	0.0	12.1	28.8	0.8	3.5	12.2	18.3	36.3	55.2	0.7	3.0	8.1	21.4	0.9	2.7	6.0
Mississippi	0.2	11.5	30.1	0.6	3.7	9.5	0.8	27.7	58.0	0.5	3.1	6.2	18.4	1.1	2.2	4.2
Ob	1.3	18.9	49.1	0.8	3.8	14.4	0.8	47.3	97.7	0.8	3.0	9.1	18.3	0.7	2.4	4.0
Lake Chad	1.1	12.1	69.7	0.9	7.1	19.8	13.0	28.3	44.6	1.8	6.9	14.2	13.9	3.6	6.1	8.4
Niger	3.8	19.6	62.2	0.9	7.3	16.0	16.0	44.4	60.0	2.0	7.2	12.8	19.0	4.9	6.6	8.0
Lena	0.9	16.4	61.1	1.3	6.3	19.0	0.7	38.9	73.8	1.2	5.6	13.2	12.0	0.8	3.1	5.1
Mackenzie	2.2	18.8	62.4	1.0	5.1	16.3	6.0	43.9	97.6	0.9	3.9	10.2	8.8	0.9	2.2	3.5
Volga	3.9	23.9	69.9	1.2	5.7	20.2	7.8	57.7	128.7	1.4	5.4	12.7	22.4	2.0	4.8	6.7
Huang He	0.5	7.8	30.0	1.1	7.3	19.7	0.9	11.6	19.8	1.1	7.0	13.2	0.7	3.5	6.0	8.2
Murray-Darling	2.0	12.6	47.4	1.3	4.2	11.3	8.6	26.7	49.4	1.0	3.6	7.2	28.4	1.9	3.2	4.0
Ganges	1.5	27.1	85.7	1.7	6.0	15.2	22.8	53.4	92.7	0.9	4.6	9.7	9.4	1.0	2.6	4.4
Oranje	0.4	9.6	49.7	1.4	6.4	14.2	8.1	22.8	52.0	1.1	5.6	11.7	29.5	0.9	4.9	9.2
Columbia	0.9	16.3	42.7	1.6	7.5	16.6	15.5	38.7	66.2	2.4	6.5	12.3	2.0	0.9	3.8	6.2
Tibetan Plateau	0.0	1.6	6.0	3.6	23.0	62.8	1.5	2.5	3.7	5.6	20.6	41.9	4.6	11.1	14.5	21.4
Chao Phraya	1.5	34.9	116.6	21.0	29.5	38.2	33.4	82.8	145.1	16.7	21.2	27.0	26.5	13.5	15.2	16.9
Wisla	0.8	18.3	44.4	28.4	30.7	35.8	1.4	37.0	69.8	16.5	18.3	21.4	8.3	8.3	9.6	11.5
Great Salt Lake	0.0	8.2	19.3	88.1	100.2	121.4	6.0	18.3	33.5	54.0	62.8	72.4	15.4	31.8	37.0	39.7
Odra	3.1	19.1	46.1	131.4	133.5	138.0	8.0	32.4	58.7	76.0	77.5	80.0	27.9	37.8	38.8	40.0

Table 2.6: Comparison of the ranges of water storage changes and uncertainty.

Drainage Basin	Number of Time Intervals		
	Monthly (23 Total)	Seasonal (4 Total)	Annual (1 Total)
Amazon	20	4	0
Zaire	12	2	1
Parana-Uruguay	15	4	1
Mississippi	18	4	1
Ob	21	4	1
Lake Chad	14	4	1
Niger	18	4	1
Lena	20	3	1
Mackenzie	19	4	1
Volga	19	3	1
Huang He	10	3	0
Murray-Darling	20	4	1
Ganges	20	4	1
Oranje	14	3	1
Columbia	17	4	0
Tibetan Plateau	0	0	0
Chao Phraya	13	4	1
Wisla	5	3	0
Great Salt Lake	0	0	0
Odra	0	0	0

Table 2.7: Rates of water storage change detectability.

The number of time intervals, ΔT, during which ΔS is detectable ($E_T/\Delta S < 1$) by GRACE, given the data in Figures 2.4, 2.5, and 2.6.

Figure 2.1: Twenty continental scale drainage basins.

Robinson projection. The names of the basins are listed in Table 2.1.

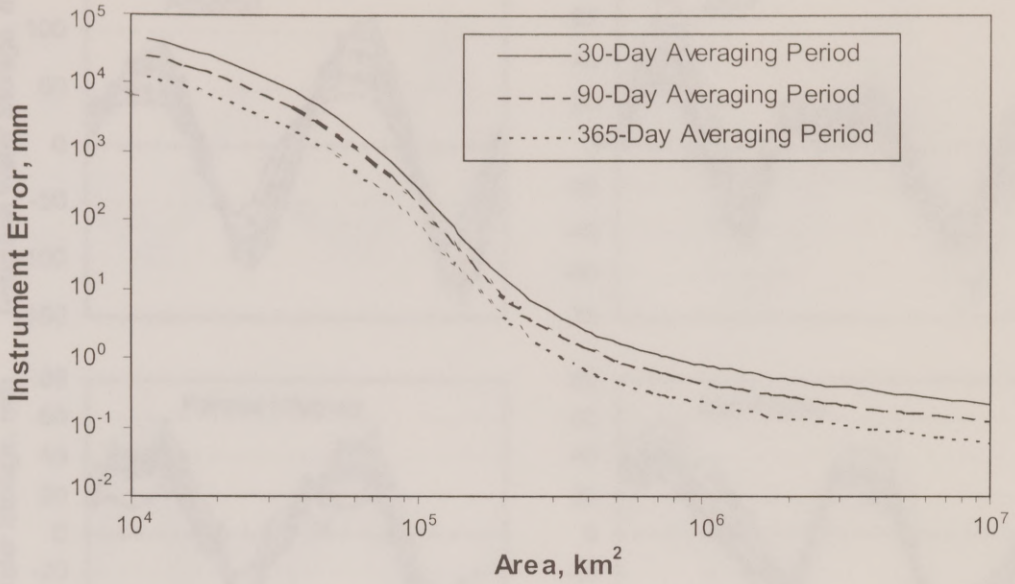

Figure 2.2: Instrument errors versus spatial scale.

Errors are equivalent heights of water. Monthly, seasonal, and annual averaging periods are shown.

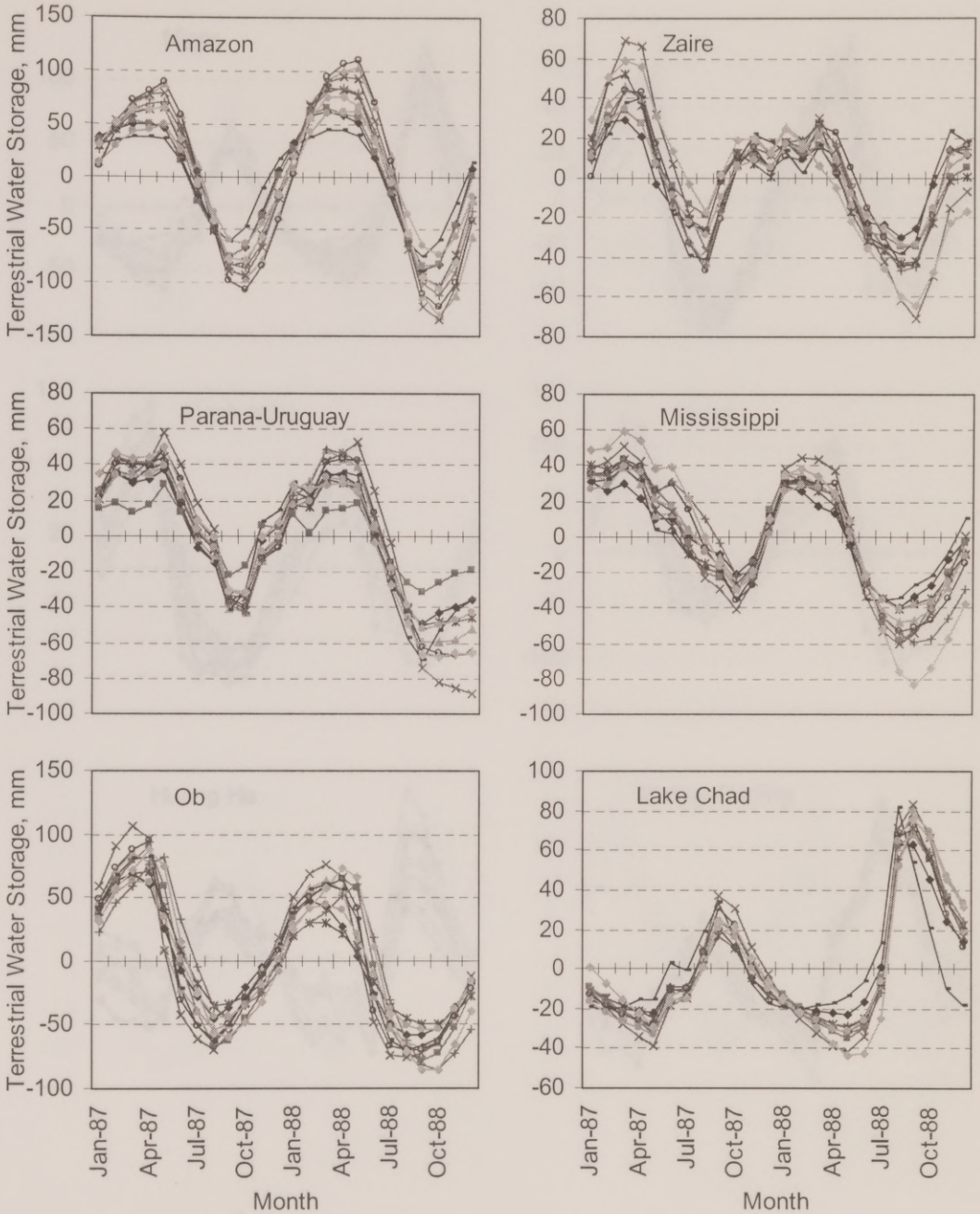

Figure 2.3: GSWP time series of terrestrial water storage.

Terrestrial water storage is the equivalent height of water, relative to the mean.

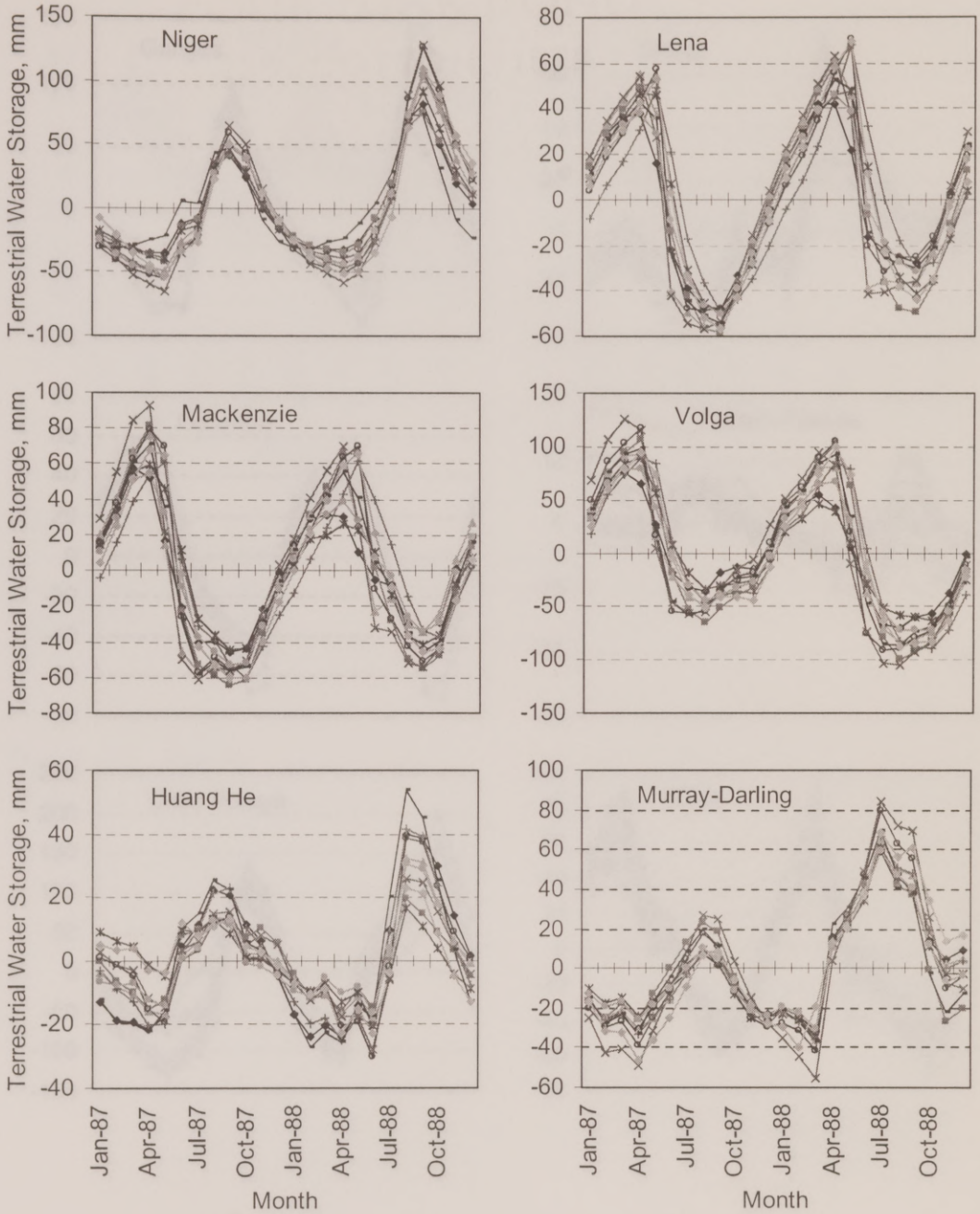

Figure 2.3: Part 2.

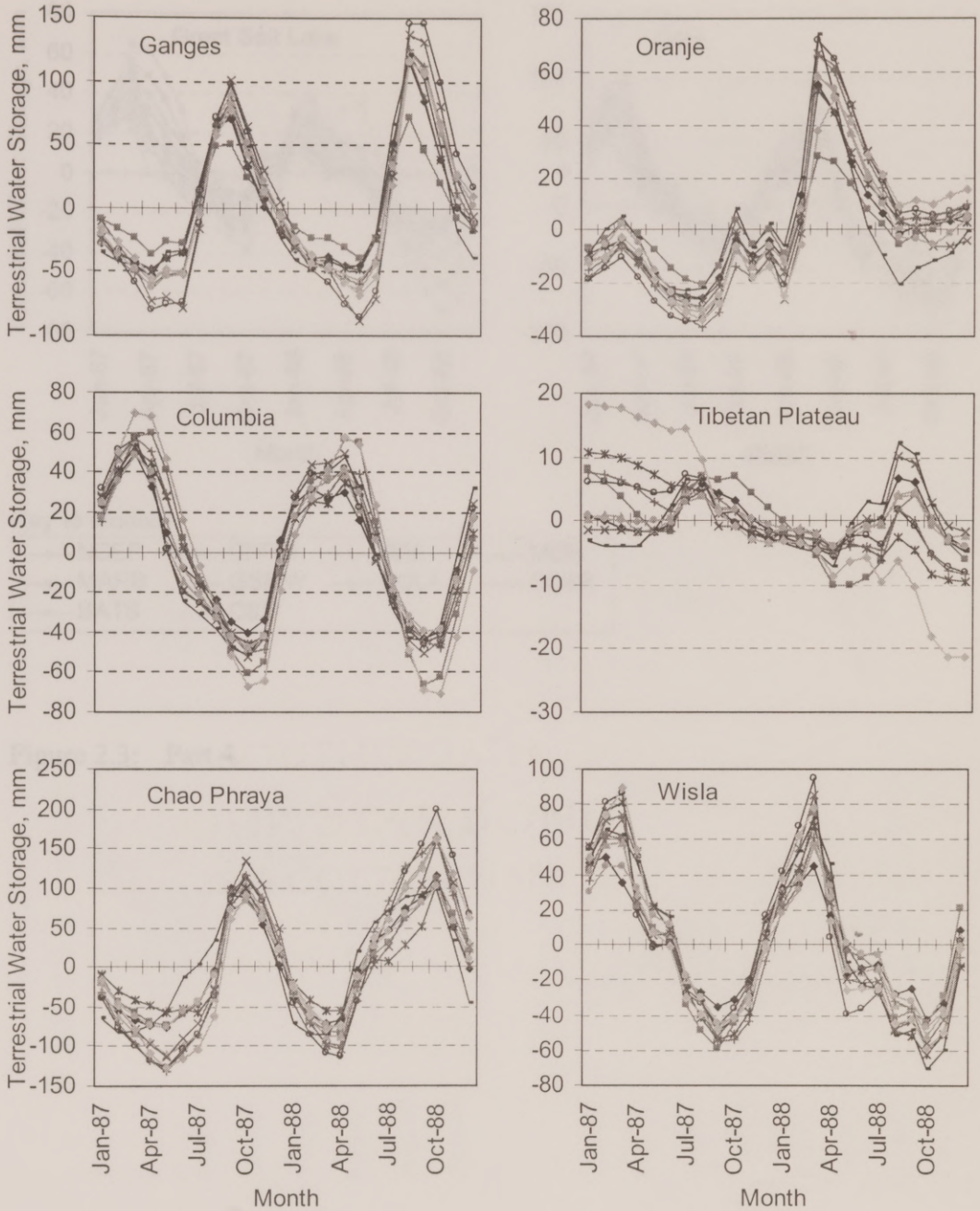

Figure 2.3: Part 3.

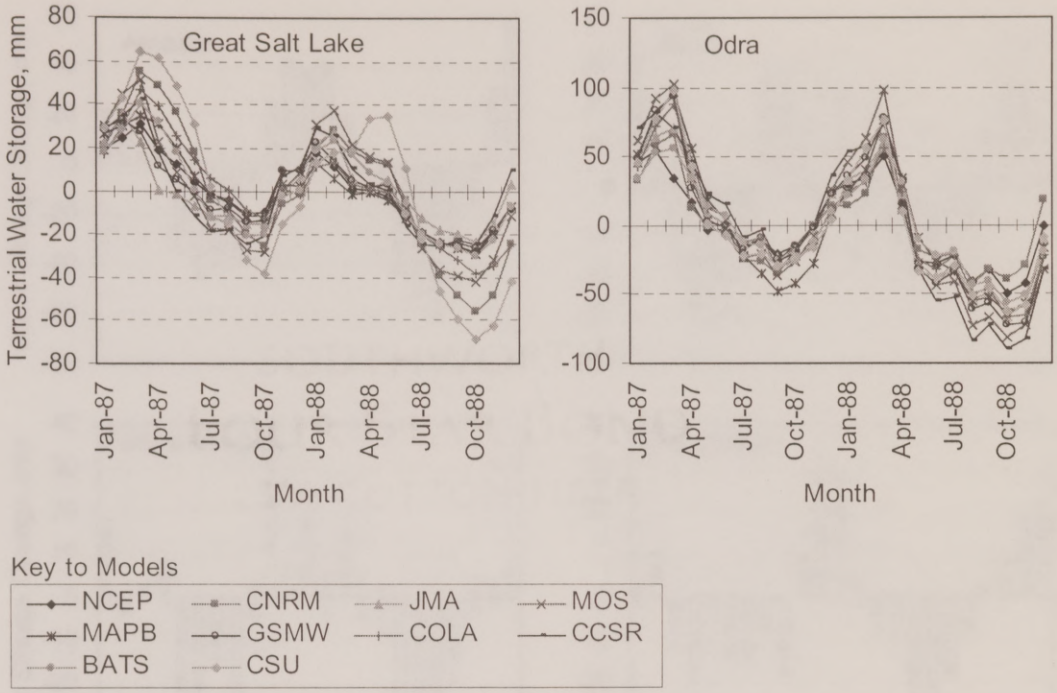

Figure 2.3: Part 4.

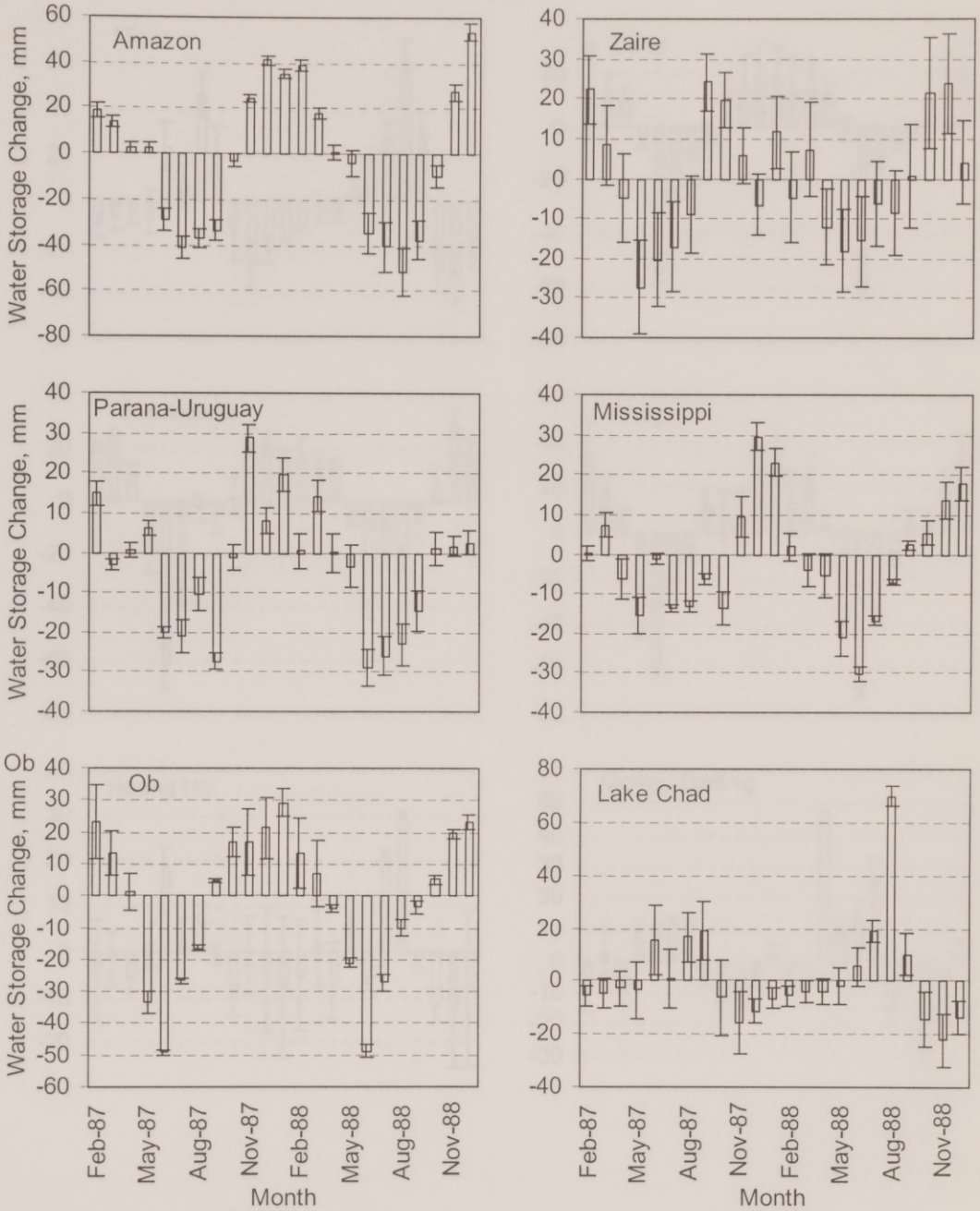

Figure 2.4: Monthly terrestrial water storage changes and uncertainty. GSWP median changes. Error bars represent $\pm E_T$.

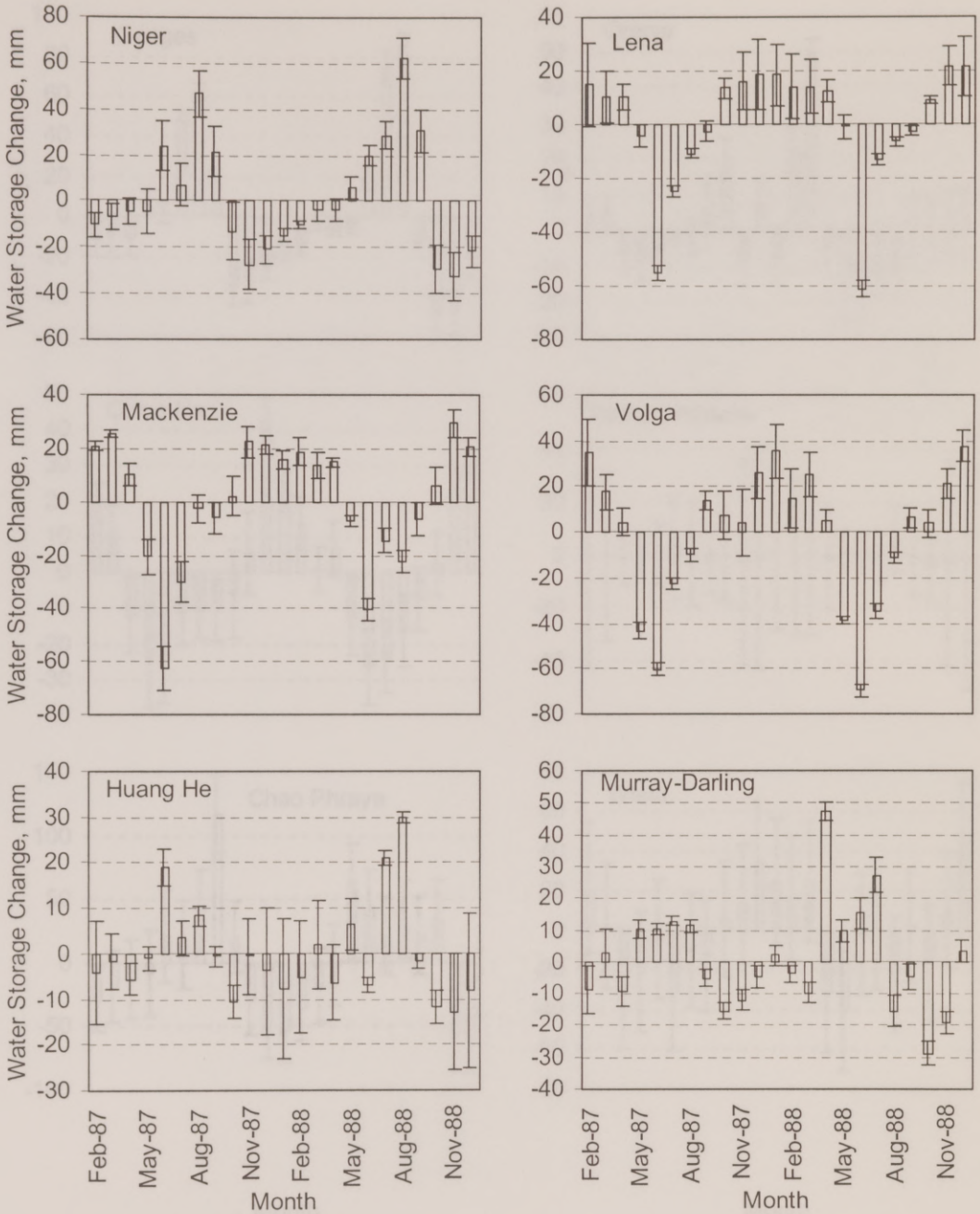

Figure 2.4: Part 2.

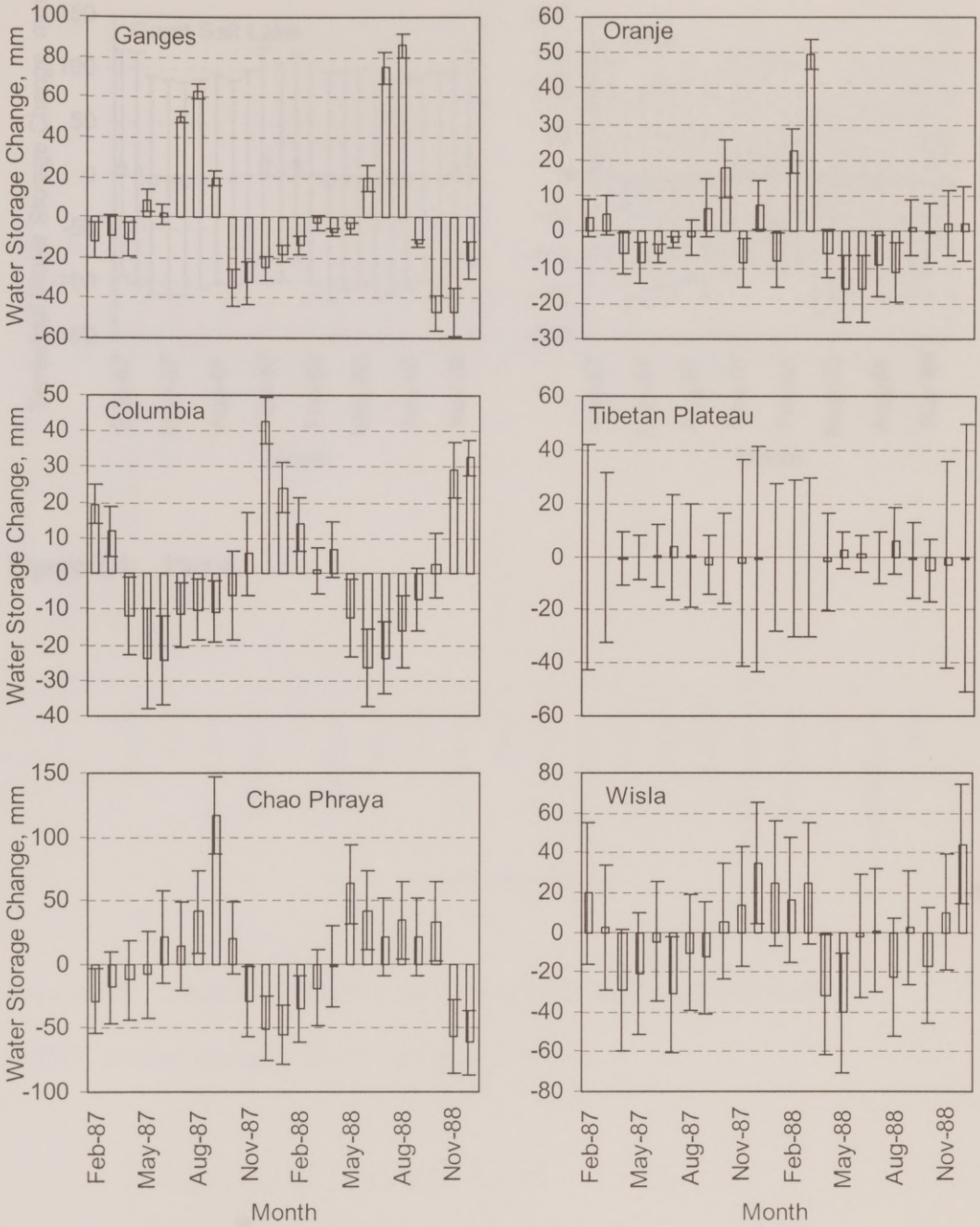

Figure 2.4: Part 3.

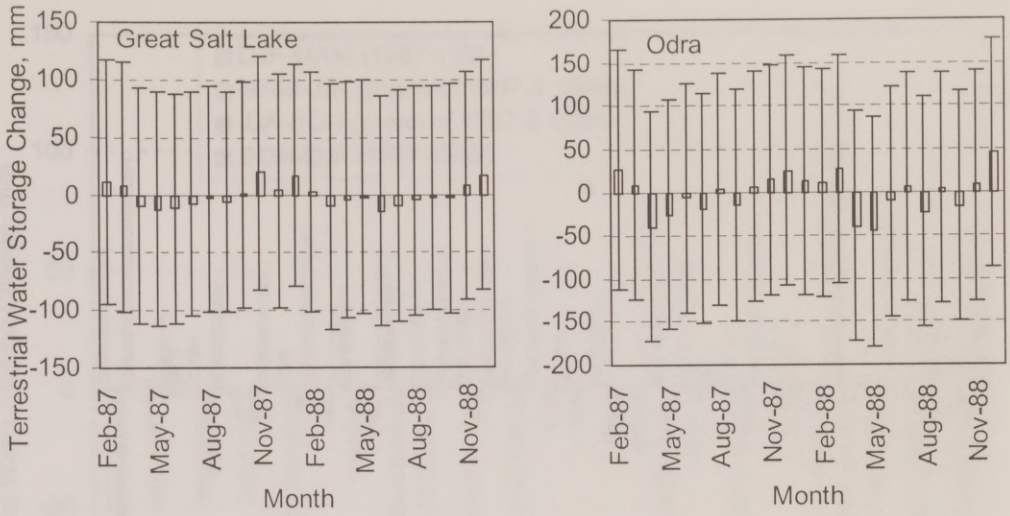

Figure 2.4: Part 4.

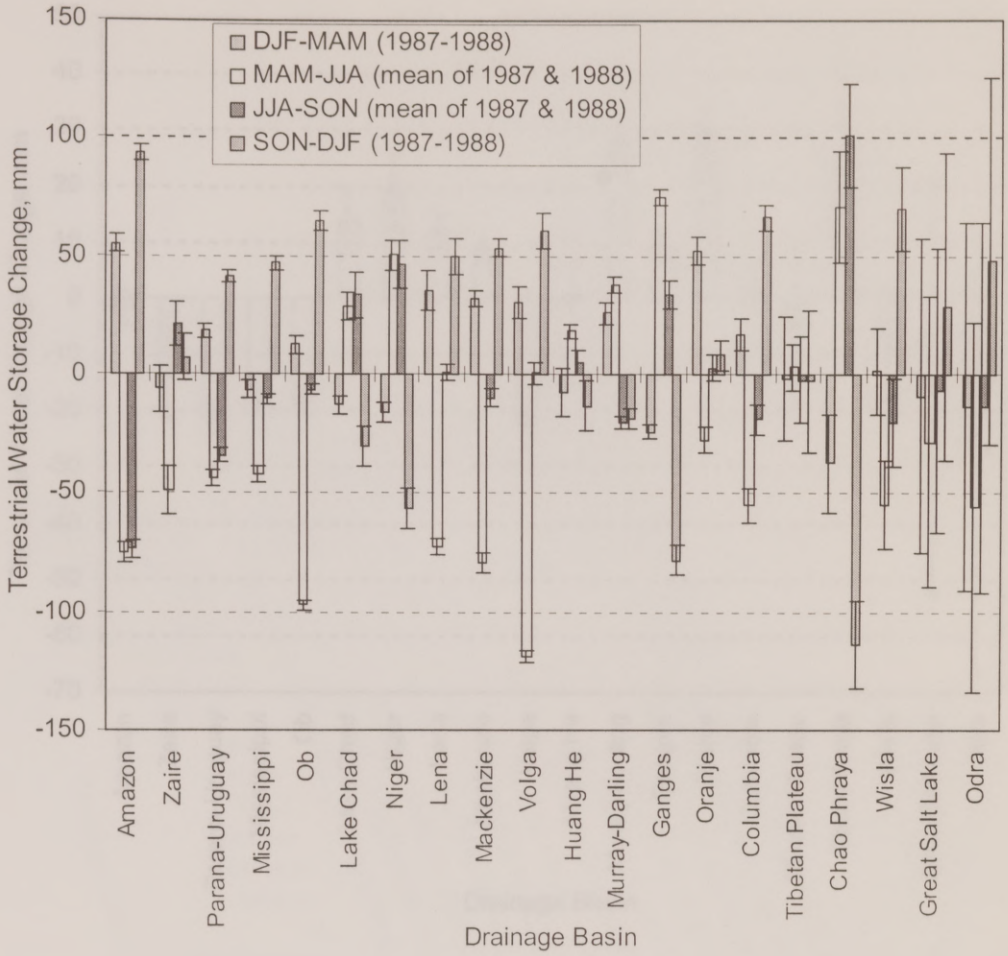

Figure 2.5: Seasonal terrestrial water storage changes and uncertainty. GSWP median changes. Error bars represent $\pm E_T$.

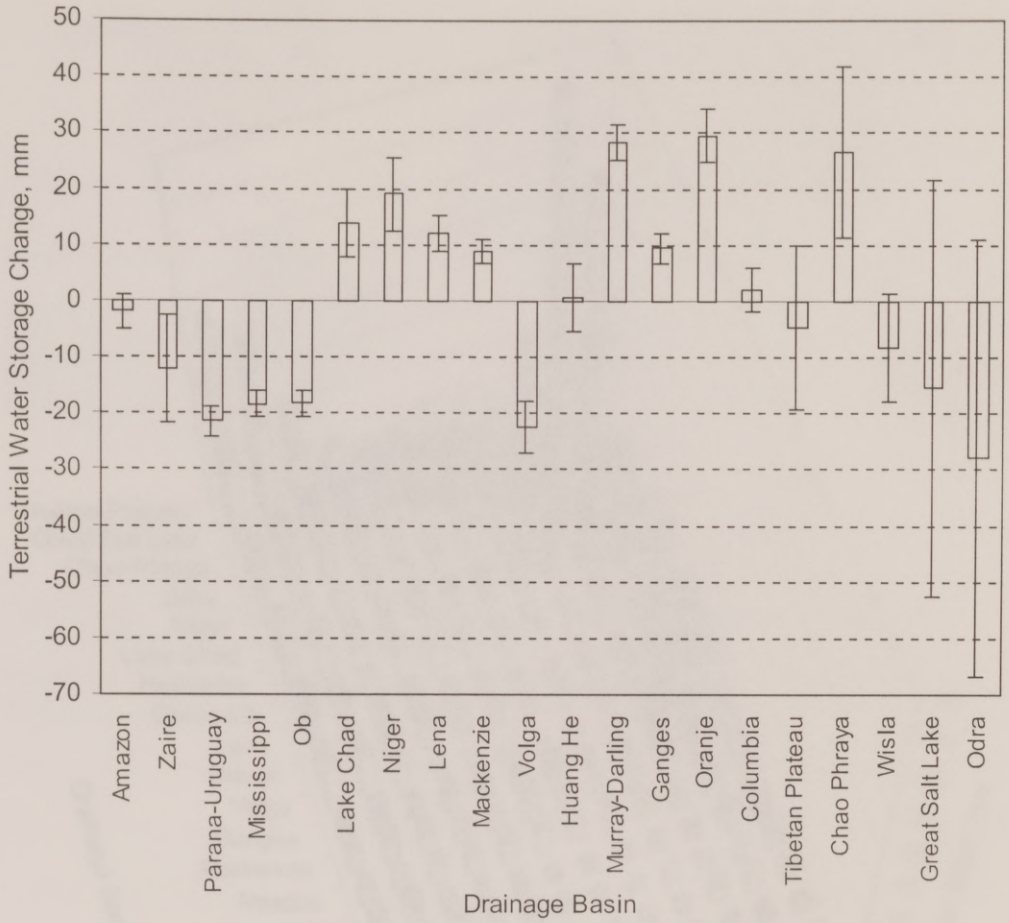

Figure 2.6: Annual terrestrial water storage changes and uncertainty.

GSWP median changes. Error bars represent $\pm E_T$.

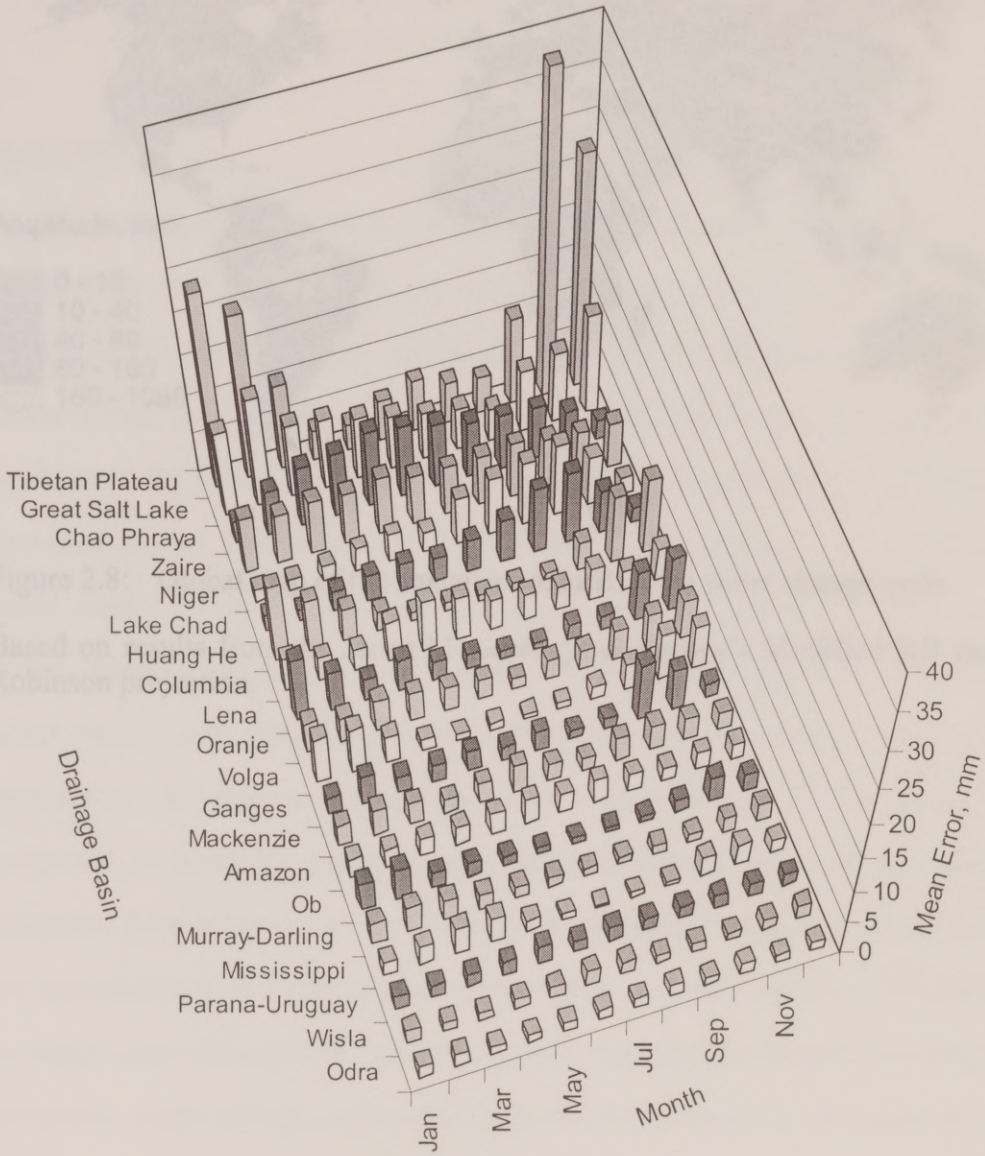

Figure 2.7: Mean monthly atmospheric errors.

Computed as the 15-year average difference between ECMWF and NCEP/NCAR model values divided by the square root of two, for each of the twenty drainage basins and each month of the year.

Figure 2.8: Global map of the annual amplitude of the water storage cycle.

Based on results from the Japan Meteorological Agency's Modified SiB model. Robinson projection.

Chapter 3: An Analysis of Terrestrial Water Storage Variations in Illinois with Implications for GRACE

SYNOPSIS

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 those 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.

Portions of this chapter have been accepted for publication as *Rodell and Famiglietti* (2000).

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*, 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 five years [*Tapley*, 1997]. The experiment will employ two satellites in a tandem orbit, 170-270 *km* apart, at about 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 redistribution at the Earth's surface, as would result from atmospheric and oceanic circulations and terrestrial water storage fluxes, is the main contributor to gravitational variations, the satellite observations will be manipulated 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, two-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 ($\geq$ 200,000 km^2) and the magnitude of the variations themselves (at least a few millimeters). This study 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, quality 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 two meters 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.

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 aquifers are unconfined and the water table usually exists at depths between one and ten meters. The primary 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 19 neutron probe monitoring stations. They calibrated the probes to soil moisture, quantified the uncertainty in the observations, and identified statewide patterns and relationships in ten years of data. They determined that uncertainty in the data was about 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 gauging 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 multi-month 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 one 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 about one month. They concluded that seasonal cycles of the hydrologic components were forced by the seasonal cycle of solar radiation, while interannual variability in the hydrologic 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 manipulated 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 five years and two 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 study addresses that deficiency and examines a longer, observation-based time series.

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 3.1) monitored by 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. Once-per-month observations often 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 eleven 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 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, 1999, personal communication), and various topographic maps. Records from five of the seven lakes with areas greater than 10 km^2 extended from 1983 or 1985 to 1999, the other two 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.

METHODS

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 eleven 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. However, 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 3.1) and had complete 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

$$IZ = H_{IZ} * \left(\frac{\theta + n}{2} \right), \tag{3.1}$$

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.

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 post glacial 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*, The University of Texas, 1998, personal communication). 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:

$$E_I = \sqrt{E_{I,1}^2 + E_{I,2}^2}\,, \tag{3.2}$$

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 in *Wahr et al.* [1998];

$$E_{A,i} = \frac{\left|\overline{P}_{ECMWF} - \overline{P}_{NCEP/NCAR}\right|}{\sqrt{2}}, \tag{3.3}$$

where $\overline{P}$ 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 $(\overline{P}_{ECMWF} - \overline{P}_{NCEP/NCAR})$, 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;

$$E_A = \sqrt{E_{A,1}^2 + E_{A,2}^2}, \tag{3.4}$$

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;

$$E_T = E_I + E_A. \tag{3.5}$$

RESULTS

Terrestrial Water Storage

Figure 3.2 depicts the entire 13 ½ 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 3.2) 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 to either the specific yield or the porosity, while the latter allows for a range of water contents.

Figure 3.2 demonstrates that there is significant seasonal and interannual variability in terrestrial water storage in Illinois. Changes in *SM* and *GW+IZ* (ΔSM and $\Delta GW+IZ$) 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. The figure suggests that there may be a cycle of terrestrial water storage in Illinois with a period of about 7 years, possibly linked to the El Nino 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 1988-89 and the wet summer of 1993 are obvious in Figure 3.2. 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. 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 3.3 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 $\Delta GW+IZ$ 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* 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 less than 1 *mm*.

Figure 3.4 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

$$\overline{\left|\Delta S_{T(April)}\right|} = \sum_{year=1983}^{1996} \left| S_{T(April,year)} - S_{T(March,year)} \right| \Big/ 14, \tag{3.6}$$

where $S_{T(April,year)}$ 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 detectability by GRACE. Simple mean changes misrepresent the average magnitudes.

In Figure 3.4 it is 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. Δ*SM* 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 less than 1 *mm* per month throughout the year.

Figure 3.5 shows mean absolute seasonal (i.e., DJF, MAM, JJA, 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 (4.6)). 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 about 30 *mm*. ΔSM dominates in winter and summer, averaging over 70 *mm*, but in spring and autumn it is close in magnitude to ΔGW, which averages about 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 3.6 shows changes in annual mean terrestrial water storage. For example, the 1984 change in soil moisture storage was calculated as

$$\Delta SM_{1984} = SM_{1984} - SM_{1983} \tag{3.7}$$

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 two 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 1988-89. The average magnitudes of annual changes in *SM* and *GW+IZ* were nearly identical, about 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 $\Delta GW+IZ$ 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 3.6 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 twelve years.

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 a hypothetical GRACE-derived estimate of ΔS_T, averaged for each month of the year. As previously mentioned, 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 upper left panel of Figure 3.7. 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. Five additional panels in Figure 3.7 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 3.8 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 3.9 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 seven of twelve years. ΔS_T was detectable in nine of twelve years for a 200,000 km^2 region, ten of twelve years for a 300,000 km^2 region, and eleven of twelve 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 non-change 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 $\Delta GW+IZ$ becomes larger for longer averaging periods; annual $\Delta GW+IZ$ is about the same magnitude annual ΔSM. ΔSN is occasionally significant on a monthly or seasonal basis, while ΔRS is never more than 2-3 *mm*, being largest seasonally. All types of water storage changes are occasionally as small as a 6 *mm* or less.

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 3.1 and 3.2 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 - greater than 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 time scales, but for any specific situation ΔS_T must be greater than the minimum for it to be detectable.

DISCUSSION

Table 3 lists the number of intervals when ΔS_T was detectable by GRACE (i.e., ΔS_T 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 was 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 about 18% of the ΔS_T values are 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 3.1 and 2 that the minimum seasonal ΔS_T and the mean seasonal E_T are both about 5 *mm* at that spatial scale. Annual ΔS_T was detectable nine times out of twelve for a 200,000 km^2 region and eleven times out of twelve for 500,000 km^2 and larger regions. As mentioned in the previous section, the 1992 annual change (-0.11 *mm*) was not realistically detectable at any scale.

Because the lifetime of the GRACE mission will be five years it is worth examining the range of variability of ΔS_T for periods of five consecutive years. Mean E_T does not vary appreciably among five-year periods, so the series means in Table 3.2 are appropriate for comparison. For monthly changes the least variable five-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 five-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 five-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 more than 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, one of four annual water storage changes were detectable during the first period and three of four during the second period, seven of twelve changes being detectable over the course of the time series. For the other spatial scales, two or three out of four changes were detectable for the first period and four of four for the second, compared with nine to eleven out of twelve for the time series. Thus, the degree of water storage variability is not constant among five-year periods and will influence detectability during the five-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 3.1 to the uncertainty information in Table 3.2 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 time scales 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 $\Delta GW+IZ$ might be isolated in 300,000 km^2 regions, seasonal $\Delta GW+IZ$ in 200,000 km^2 regions, and annual $\Delta GW+IZ$ 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 assumption 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 hydrologic model is suggested. These advancements would expedite the delivery of useful GRACE-derived products to the hydrologic community.

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 scale, 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.

	Monthly ΔS, mm			Seasonal ΔS, mm			Annual ΔS, mm		
Component	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
ST	0.0	28.2	109.8	4.8	58.7	141.8	0.1	48.5	105.2

Table 3.1: Ranges of absolute changes in terrestrial water storage.

ΔS is change in water storage. *RS* is reservoir water storage, *SN* is snow water storage, *SM* is moisture storage in the top two meters of soil, *GW+IZ* is groundwater and intermediate zone storage, and S_T is total terrestrial water storage.

Region	Monthly E_T, mm			Seasonal E_T, mm			Annual E_T, mm		
	Min	Avg	Max	Min	Avg	Max	Min	Avg	Max
Illinois (145,800 km^2)	77.8	81.4	89.9	44.8	48.6	54.3	23.9	25.4	27.1
Area = 200,000 km^2	21.0	24.5	33.1	12.0	15.8	21.5	7.6	9.1	10.8
Area = 300,000 km^2	6.6	10.1	18.7	3.7	7.5	13.2	3.5	5.0	6.6
Area = 500,000 km^2	2.8	6.4	14.9	1.5	5.3	11.0	2.4	3.9	5.6
Area = 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

Table 3.2: Ranges of total uncertainty in GRACE-derived estimates.

E_T is total uncertainty.

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

Table 3.3: Rates of water storage change detectability.

The number of time intervals, ΔT, during which ΔS is detectable ($E_T/\Delta S < 1$) by GRACE.

Figure 3.1: Map of Illinois with locations of monitoring stations.

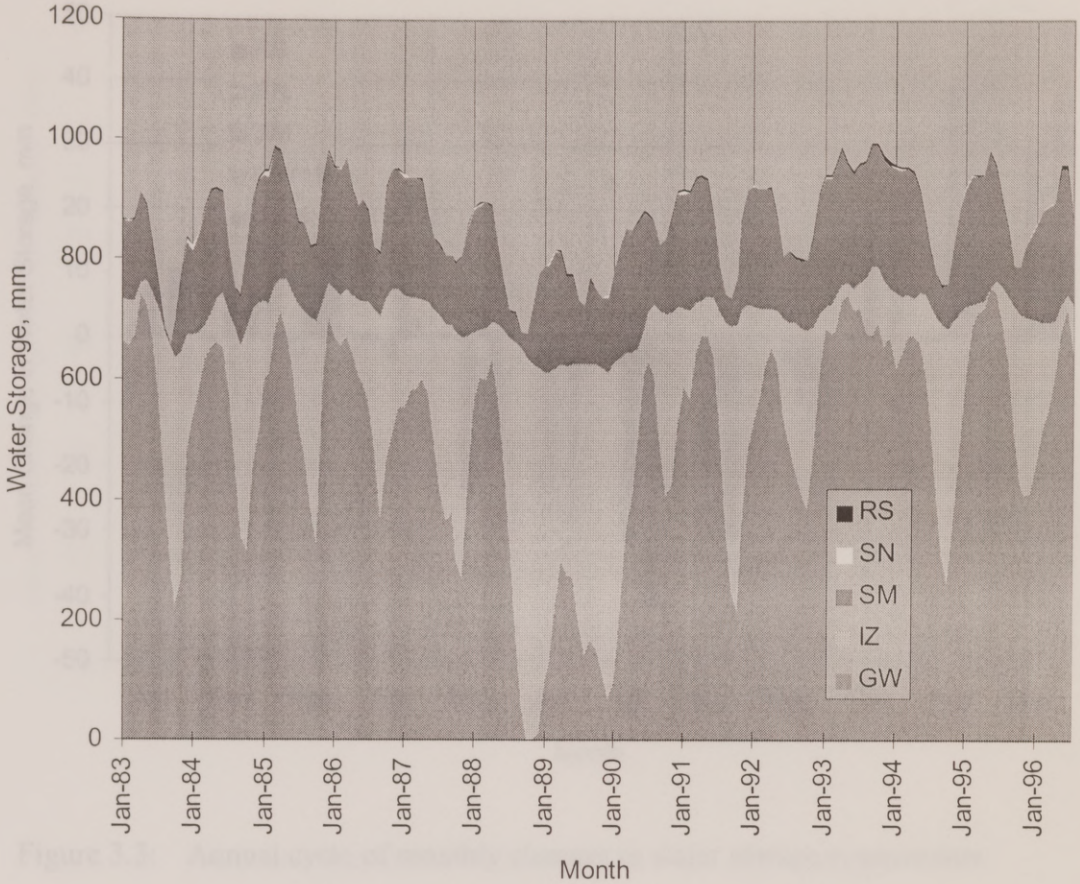

Figure 3.2: Time series of terrestrial water storage components.

Water storage is the equivalent height of water, averaged over Illinois, relative to each component's minimum value, which has been set to zero. RS is reservoir water, SN is snow, SM is soil moisture, IZ is intermediate zone moisture, and GW is groundwater.

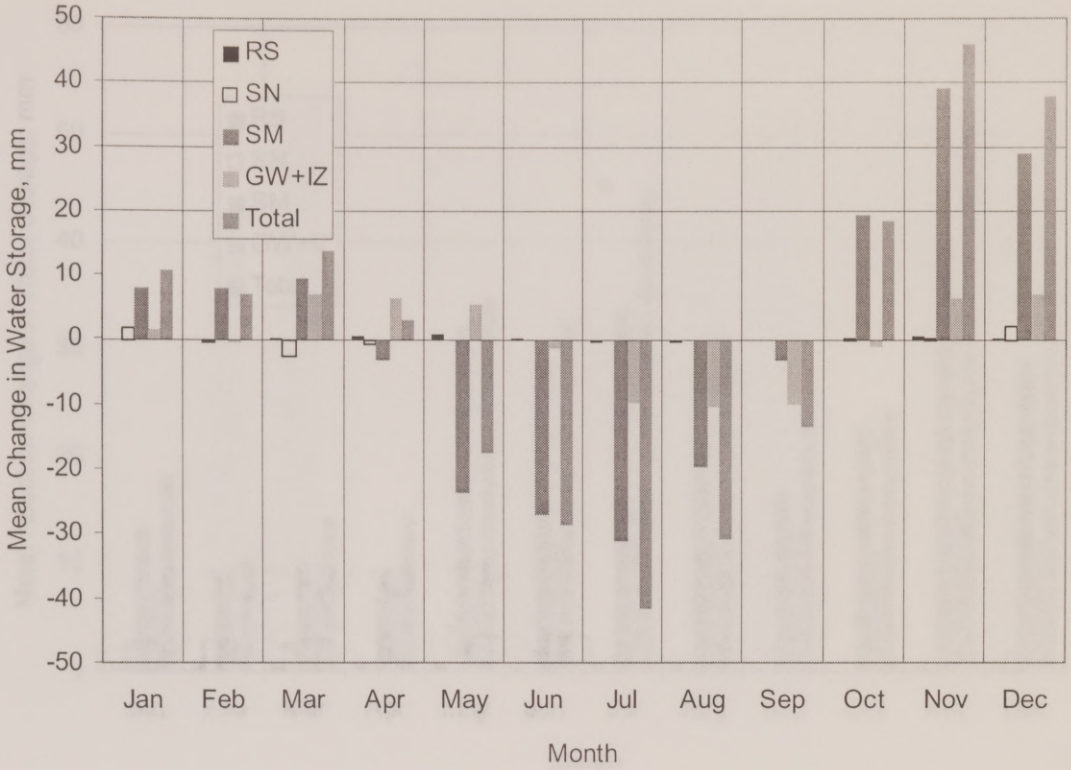

Figure 3.3: Annual cycle of monthly changes in water storage components.

RS is reservoir water, SN is snow, SM is soil moisture, IZ is intermediate zone moisture, and GW is groundwater.

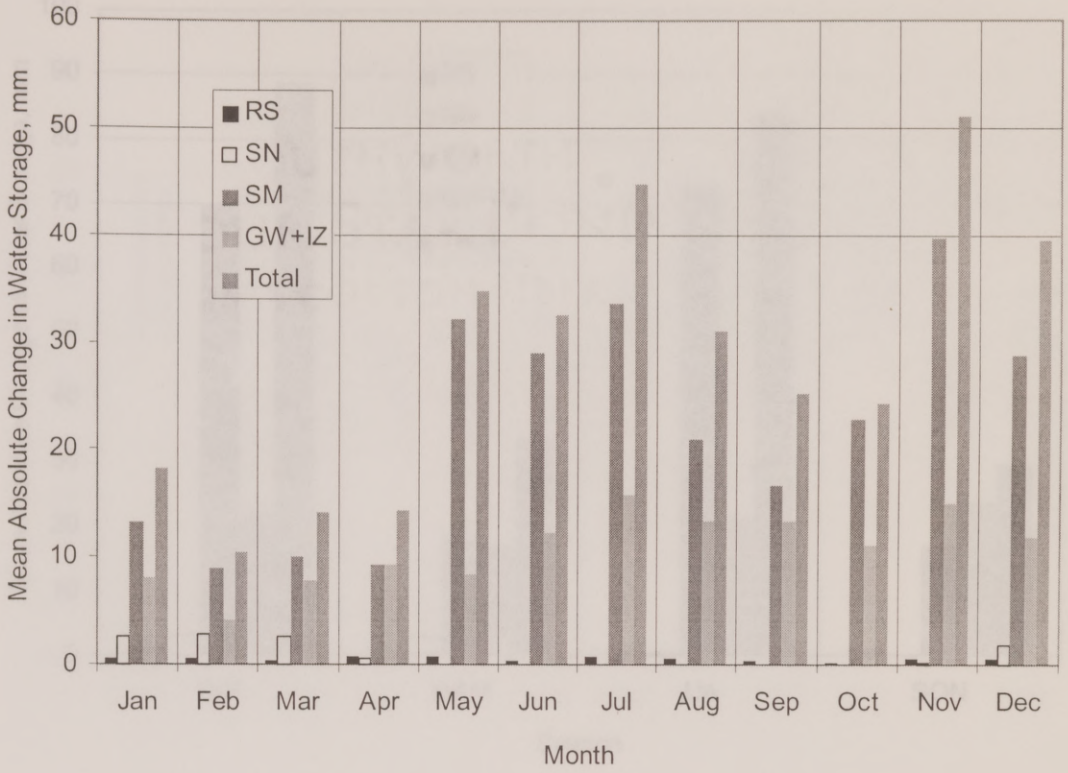

Figure 3.4: Mean magnitudes of monthly changes in water storage.

RS is reservoir water, SN is snow, SM is soil moisture, IZ is intermediate zone moisture, and GW is groundwater.

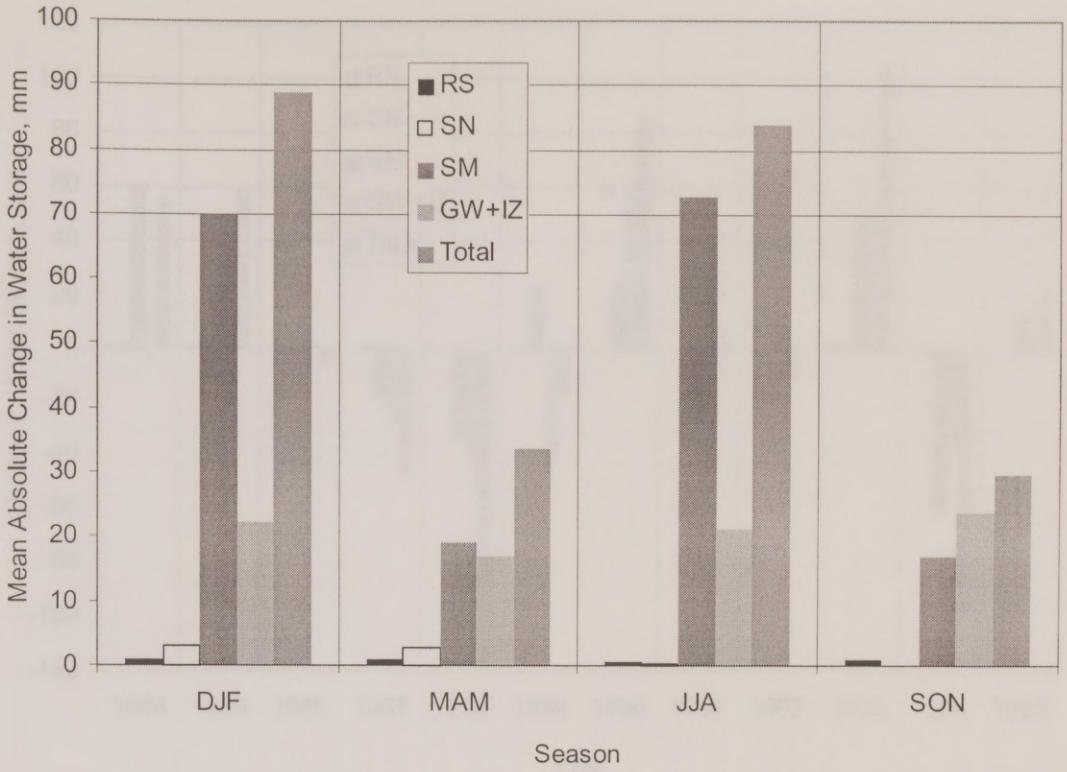

Figure 3.5: Mean magnitudes of seasonal changes in water storage.

RS is reservoir water, SN is snow, SM is soil moisture, IZ is intermediate zone moisture, and GW is groundwater.

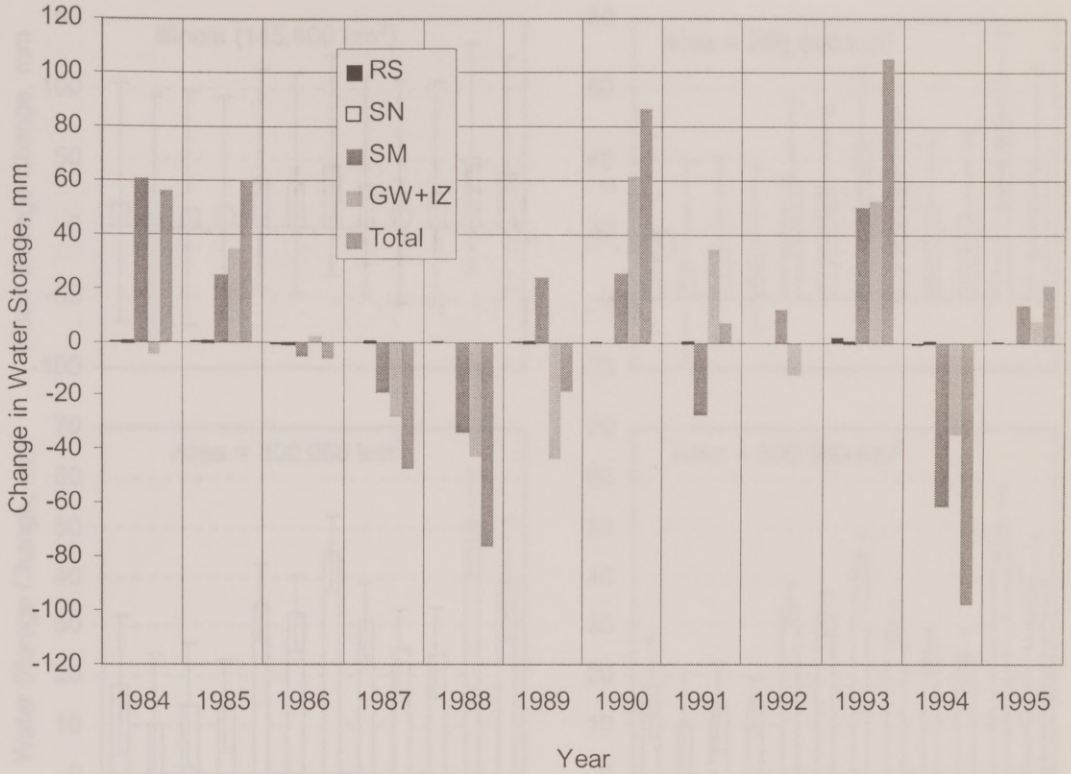

Figure 3.6: Annual changes in water storage components.

RS is reservoir water, SN is snow, SM is soil moisture, IZ is intermediate zone moisture, and GW is groundwater.

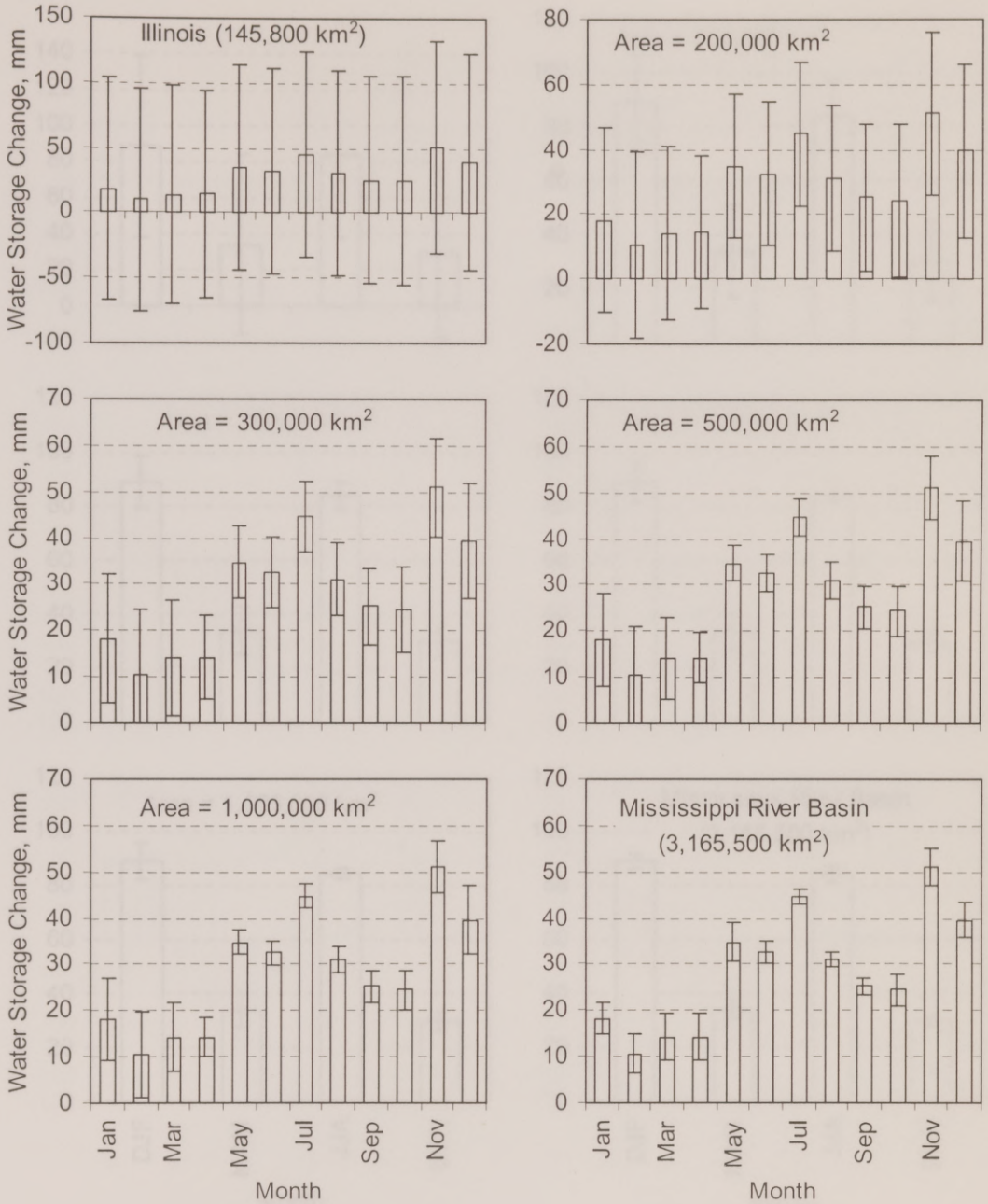

Figure 3.7: Monthly changes in total water storage and estimated uncertainty. Changes are mean magnitudes for Illinois. Error bars represent $\pm E_T$.

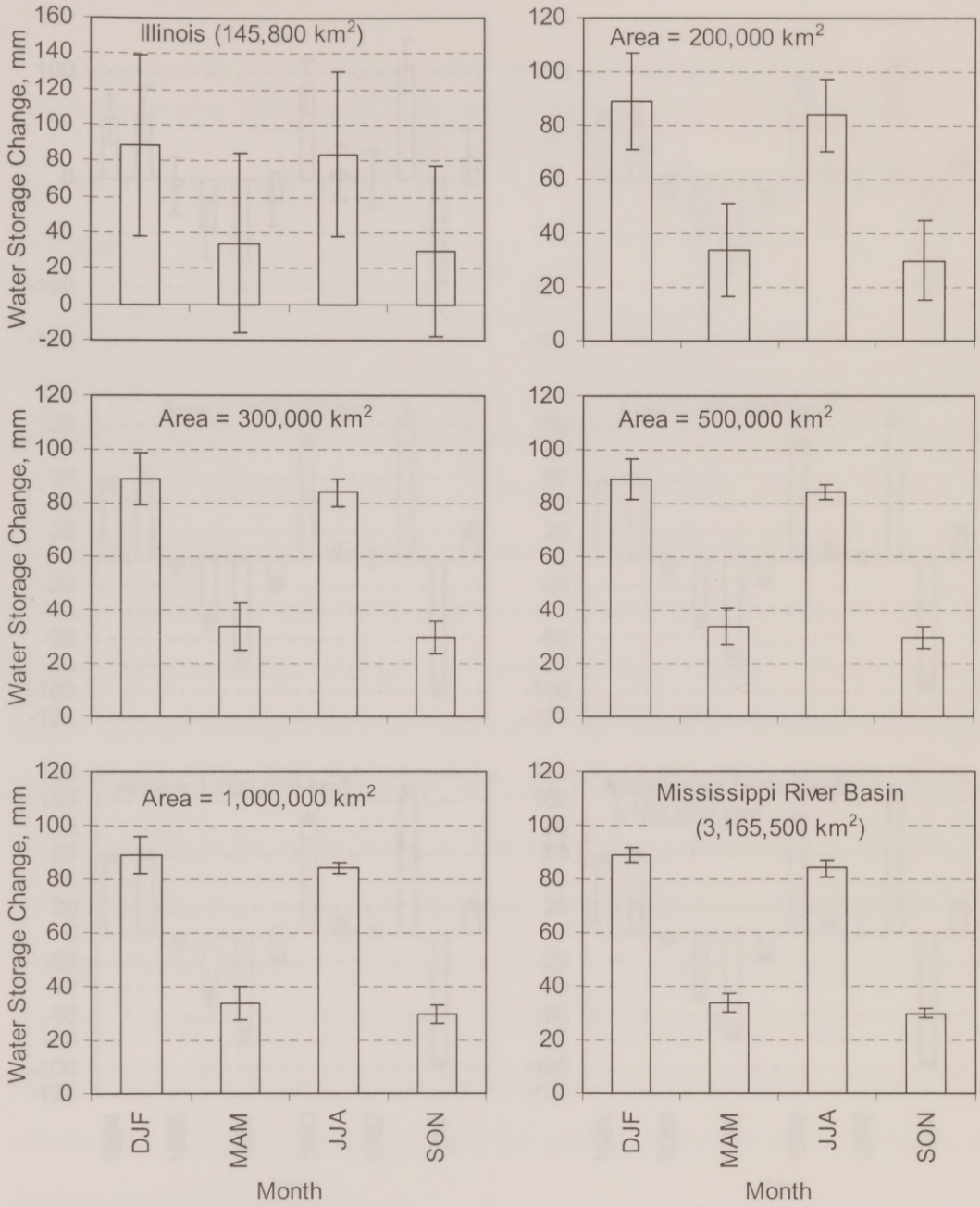

Figure 3.8: Seasonal changes in total water storage and estimated uncertainty. Changes are mean magnitudes for Illinois. Error bars represent $\pm E_T$.

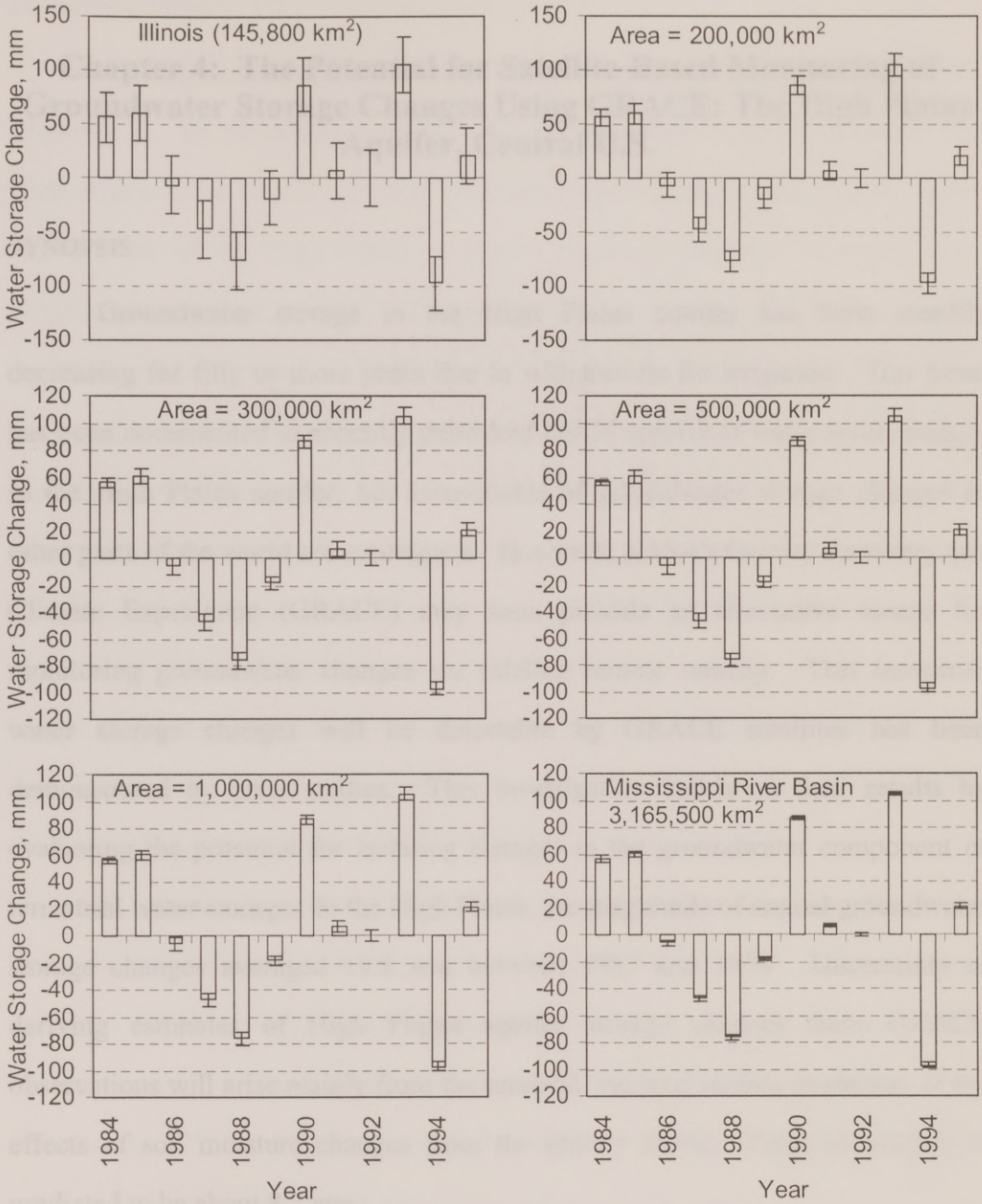

Figure 3.9: Annual changes in total water storage and estimated uncertainty. Time series of changes averaged over Illinois. Error bars represent $\pm E_T$.

Chapter 4: The Potential for Satellite-Based Monitoring of Groundwater Storage Changes Using GRACE: The High Plains Aquifer, Central U.S.

SYNOPSIS

Groundwater storage in the High Plains aquifer has been steadily decreasing for fifty or more years due to withdrawals for irrigation. This trend has been documented in annually published USGS reports of water level changes in the High Plains aquifer, but assessments of groundwater storage changes in other parts of the world are incomplete. However, NASA's Gravity Recovery and Climate Experiment (GRACE) may soon provide an alternative means for monitoring groundwater changes via satellite remote sensing. That terrestrial water storage changes will be detectable by GRACE satellites has been demonstrated by prior studies. This investigation builds on their results by evaluating the potential for isolating changes in the groundwater component of terrestrial water storage. In the High Plains, the magnitude of annual groundwater storage changes averaged 19.8 *mm* between 1987 and 1998. Uncertainty in deriving estimates of High Plains aquifer storage changes from GRACE observations will arise mainly from the removal, via land surface modeling, of the effects of soil moisture changes from the gravity signal. Total uncertainty is predicted to be about 8.7 *mm*.

INTRODUCTION

The High Plains aquifer underlies 450,000 km^2 of the central United States (Figure 4.1). By providing a steady supply of water for irrigation, it enables the High Plains to be one of the most productive agricultural regions in the world. The climate of the High Plains is largely semi-arid, so that, in many years, irrigation is vital to the success of crops. Following droughts in the 1930s (when the southern part of region became known as the "Dust Bowl") and the 1950s, farmers greatly increased their usage of groundwater. In the last fifty years of the twentieth century discharge from the aquifer typically exceeded recharge, so that average groundwater levels steadily declined. It has been estimated that, from the time when development began in the late 1800s until 1980, about 205 km^3 of water were removed from storage [*Miller and Appel*, 1997], or an equivalent height of 456 *mm* averaged over the area, causing the water table to descend more than 30 *m* in some parts. Between 1950 to 1980 the average rate of water table decline was about 10 *cm* per year [*Luckey et al.*, 1981].

In response to growing concerns about groundwater depletion, the United States Geological Survey (USGS), in cooperation with Federal, State, and local agencies, began annually assessing changes High Plains aquifer levels in 1988. Measurements from over 6200 wells scattered across the region are used in the calculations. Ninety-eight percent of the wells are monitored manually, usually once or twice per year, while the others contain continuous recorders [*McGuire and Fischer*, 2000]. Owing to the spotty nature of the measurements and the need for quality control, the task is difficult, but manageable. Most of the world's

aquifers are not surveyed regularly and methodically, and without a centralized database of well records, if they even exist, regional assessment of groundwater level change is tedious or impossible.

However, a new era of water storage change estimation, via remote sensing, will begin in June, 2001, when NASA's Gravity Recovery and Climate Experiment (GRACE) commences. GRACE will utilize two satellites in a tandem, near-polar orbit, 170-270 *km* apart, at an altitude of approximately 480 *km* [*Tapley*, 1997]. Using precise measurements of the distance between the two satellites recorded by an onboard microwave tracking system, GRACE scientists will produce a new model of the Earth's gravity field every 30 days throughout the five year duration of the mission. Given a GRACE-derived gravity model from each of two time periods, it will be possible to calculate the change in mass over a region of interest that would have been necessary to cause the observed change in the gravity field. Over land, temporal variations in the gravity field are caused mainly by changes in terrestrial water storage (groundwater, soil moisture, snow and ice, lakes and rivers, and water contained in biomass) and atmospheric mass. Therefore, given information on the other major contributors to mass variations, it may be possible to infer changes in groundwater storage. Note that it will not be possible to infer the absolute mass of water storage itself.

Rodell and Famiglietti (2000) demonstrated that annual (i.e., year to year) variations in groundwater in Illinois were similar in magnitude to annual variations in soil moisture in the top two meters. Both contributed significantly to annual changes in total water storage, while snow and reservoir water storage did

not. It was concluded that terrestrial water storage changes were likely to be detectable by GRACE in 200,000 km^2 and larger regions, however, the likelihood of isolating a single component of terrestrial water storage, such as groundwater, was not explored. This chapter evaluates the potential to estimate annual changes in the water stored in the High Plains aquifer using satellite-based gravity observations from GRACE. It builds on prior results by analyzing a region that is experiencing a long-term trend of groundwater level decline and by assessing the potential to isolate the groundwater trend from variations in soil water storage. The method for evaluating the uncertainty in the GRACE technique has also been modified from past investigations. While C-band and L-band microwave remote sensors have been tested for near-surface (0-5 *cm*) soil water retrieval [e.g., *Jackson*, 1999] and the Advanced Microwave Scanning Radiometer (AMSR-E) [*Spencer*, 2000], a C-band instrument, will be on board NASA's Earth Observing System (EOS) Aqua satellite, which is set for launch in December, 2000, the significance of this work is that GRACE will provide the first opportunity for monitoring deeper groundwater fluctuations from space.

BACKGROUND

In this chapter, "High Plains" and "High Plains region" will refer to the area underlain by the High Plains aquifer as shown in Figure 4.1. The High Plains region is a grassland biome with a mid-latitude, dry continental climate, although the weather can vary substantially on any time scale from hours to decades [*Kromm and White*, 1992]. Atmospheric moisture comes from the Gulf

of Mexico in the "wet" months from April to September, while dry, polar air masses descend upon the region in the winter. Annual precipitation averages 50 *cm* with a range of about 30 *cm* across the High Plains, the north and east receiving more than the south and west. Natural runoff and gains to storage are minimal, as evapotranspiration returns most of the precipitated water to the atmosphere: typical annual Class A pan evaporation rates range from 150 *cm* in the north to 270 *cm* in the south. However, flash-flooding is not uncommon, nor are any of nature's extremes, including frosts, heat waves, high winds and tornadoes, hail, and droughts. But the High Plains are not inhospitable, for although the region is sometimes called the Great American Desert, it is also known as the Breadbasket of the World, due in large part to the steadying effect of the groundwater (the "underground rain", as *Green* [1973] termed it).

Other than the Sandhills of Nebraska, where 52,000 km^2 of sand dunes decorate the landscape, the High Plains are treelessly flat. Western portions of the region lie more than 1200 *m* above sea level while some eastern portions are below 600 *m*, but the slope is imperceptible. Ephemeral and occasionally perennial playa lakes exist in some parts of the central and southern High Plains. A network of streams formed the plain, which was subsequently uplifted, by carrying sediments eastward from the Rocky Mountains. Mollisols, naturally fertile soils which develop under grasses where a seasonal moisture deficit occurs, are prevalent [*Kromm and White*, 1992]. Due to the high potential rate of evaporation, sandy areas such as the Sandhills are important as recharge zones

where rain is able to percolate quickly through the coarse material to replenish the aquifer [*Dornbusch et al.*, 1995].

The High Plains aquifer is a 450,000 km^2 aquifer system that lies beneath parts of Nebraska, Texas, Kansas, Colorado, Wyoming, Oklahoma, New Mexico, and South Dakota (Figure 4.1). It is generally unconfined and consists of several units of Quaternary and Tertiary age [*Gutentag and Weeks*, 1980]. The principal of these is the Ogallala Formation, which constitutes 77% of the system's horizontal extent. The Ogallala is a Miocene-aged alluvial deposit composed of unconsolidated gravel, sand, silt, and clay. Other water yielding units include the Brule Formation in the northwest, a massive siltstone whose ability to hold water is limited to secondary porosity; the Arikaree Group, a fine-grained sandstone which overlies the Brule; unconsolidated deposits overlying the Ogallala in the east, which are actually reworked material from that formation; valley-fill deposits along the channels of streams that are hydraulically connected to the aquifer; loess deposits composed of eolian silt; and the previously mentioned Sandhills [*Weeks and Gutentag*, 1988]. Tertiary to Permian aged sandstone, siltstone, shale, gypsum, anhydrite, dolomite, limestone, and halite underlie the aquifer.

The water table is less than 30 *m* below ground in about half of the High Plains region and approaches the surface near hydraulically connected rivers such as the Platte and the Arkansas. Locally, the depth to water can be 120 *m* or more. In 1992 the average saturated thickness of the aquifer was about 58 *m*, with a range of 0 to 300 *m*, and the total volume of saturated material was estimated to

be 26,800 km^3 [*Miller and Appel*, 1997]. The water table is generally parallel to the land surface, sloping at about 1.9-2.8 meters per kilometer, so that water flows from west to east at about 30 *cm* per day [*Weeks and Gutentag*, 1988]. However, *Larkin and Sharp* [1988] indicated that, in Texas, groundwater development has produced mounds and sinks in the potentiometric surface, whose configuration now reflects the irregular base of impermeable bedrock more closely than surface topography.

In 1980, approximately 170,000 wells withdrew an estimated 22 km^3 of water from the High Plains aquifer to irrigate 56,000 km^2 [*Weeks*, 1986]. Irrigation accounts for about 95% of the groundwater withdrawn. Prior to development, which began in the late 1800's and accelerated during droughts in the 1930's and 1950's, the High Plains aquifer was in equilibrium, as the rate of recharge by infiltration equaled, on average, the rate of discharge to streams, springs, and seeps and by evapotranspiration. However, pumpage from wells and subsequent redistribution in canals and by irrigation have altered the patterns of recharge and discharge and disturbed the balance. Irrigation increased infiltration from an estimated predevelopment rate of 680,000 m^3/day to a 1960-1980 rate of 14,400,000 m^3/day [*Alley et al.*, 1999]. Still, most of the water pumped for irrigation is lost to the atmosphere, so that the net results are loss of storage and a consequential decrease in natural discharge. The sustainability of current irrigational practices is often called into question [e.g., *Flores*, 1995], as assessments of the average water table elevation made by the USGS show a long-term trend of decline of roughly 8 *cm/yr* (Figure 4.2).

Dickey et al. [1997] proposed that specialized satellites be used to monitor the Earth's gravity field from space, and theorized that those gravity measurements would be useful for estimating changes in terrestrial water storage. After their proposal begot GRACE, *Wahr et al.* [1998] and *Rodell and Famiglietti* [1999] evaluated the theory using modeled soil moisture and snow data, by comparing the magnitudes of the water storage changes to estimates of the uncertainty in the technique. The studies agreed that terrestrial water storage changes were potentially detectable by GRACE on monthly or longer time intervals, depending on the size of the region of interest and the magnitude of the changes themselves. The subject was also discussed by *Dickey et al.* [1999], and *Wahr et al.* [2000] showed how GRACE data could be used to help monitor ice mass changes in Antarctica. *Nerem et al.* [2000] used data from six years of satellite laser ranging to the Lageos 1 and 2 satellites to demonstrate their orbits were measurably perturbed by changes in water storage on Earth. *Rodell and Famiglietti* (2000) incorporated groundwater storage information into an observation-based study of projected water storage detectability by GRACE. One of their conclusions was that, on an annual basis, changes in soil water in the upper 2 *m* were matched in magnitude by changes in deeper water storage. This study extends previous work by assessing the potential to isolate annual changes in High Plains aquifer groundwater storage from other factors that will affect the GRACE satellite gravity observations, including variations in soil moisture.

DATA

The USGS has assessed water level changes in the High Plains aquifer for each year between 1987 and 1998, relying on data from a well monitoring program [*USGS*, 2000]. Information on soil moisture variations in the region and the ability to estimate those variations was required in order to assess the potential to disaggregate groundwater from total water storage changes. The investigation used soil moisture observations from nine locations in the general region (see Figure 4.1). The Soil Climate Analysis Network (SCAN) of the National Resource Conservation Service [*Schaefer et al.*, 1995] collected and continues to collect observations at five of these locations. At each SCAN site, a suite of sensors automatically monitors and records meteorological and soil conditions. Measured soil properties include the dielectric constants at depths of 5, 10, 20, 51, and 102 *cm*, which are used to determine soil water contents. Observation at these five sites began between November, 1996, and June, 1997, and, at the time of this study, were available through June, 2000. Observations at the other four sites were collected by the Department of Energy's Atmospheric Radiation Measurement Soil Water and Temperature System (ARM-SWATS) [*Schneider and Fisher*, 1997]. Two sets of instruments are installed at each site, including heat dissipation sensors which automatically monitor and record matric potential at depths of 5, 15, 25, 35, 60, 85, 125, and 175 *cm*. Matric potential is converted to soil water content using site-specific relationships. The ARM-SWATS

observations used in this investigation were made between April, 1996, and March, 1999.

Modeled data from the Global Soil Wetness Project (GSWP) [*Dirmeyer, et al.*, 1999] helped to gauge the precision of hydrologic models in simulating soil water storage (see "Methods", below). GSWP is an ongoing Global Energy and Water Cycle Experiment (GEWEX) project whose original purpose was to test the feasibility of producing a global data set of soil wetness for use in climate model initialization. In the pilot phase, ten contributing groups used different land surface schemes to produce time series of soil moisture, snow, and other variables on a 1° global grid. The ten groups used a common data set of model parameters, meteorological observations, and analyses to drive the models. The products span the 24 month period beginning January 1, 1987. The reader is referred to *Dirmeyer, et al.* [1999] for a detailed description of the GSWP models and the results of the pilot phase.

METHODS

Each USGS estimate of a change in annual average High Plains aquifer water level was converted to a change in groundwater storage, ΔGW, using a specific yield of 15.1% [*Gutentag et al.*, 1984]. For example, the mean water storage change in the High Plains aquifer between 1997 and 1998, as an equivalent height of water, was computed as

$$\Delta GW_{1998} = S_Y \times \Delta H_{WT\,1998}, \tag{4.1}$$

where S_Y is the specific yield and ΔH_{WT1998} is the 1998 annual change in the mean height of the water table estimated by the USGS. Unlike in Chapter 3, the use of a specific yield to compute groundwater storage changes was chosen in this study because of the lack of neighboring soil moisture monitoring stations. Conceptually,

$$\Delta H_{WT1998} = \overline{H_{WT\ 1998}} - \overline{H_{WT\ 1997}}, \tag{1a}$$

where $\overline{H_{WT}}$ is the height of the water table averaged over a given year. This is consistent with GRACE, which will compute a change in the gravity field as the difference between gravity observations averaged over two separate time periods.

In order to use GRACE satellite observations to estimate changes in terrestrial water storage, the effect of atmospheric mass redistribution will have to be removed from the gravity signal. This will be accomplished using modeled atmospheric pressure (a surrogate for atmospheric mass) 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). Errors in those fields will propagate to the terrestrial water storage change estimates. The atmospheric models and the limitations of the GRACE instruments themselves will be the main sources of uncertainty in the technique [*Rodell and Famiglietti*, 1999]. Furthermore, to isolate the groundwater component, estimates of changes in the other components of terrestrial water storage will be required.

Uncertainty in GRACE gravity observations will be inversely related to both the area of the region of interest and the length of the measurement

averaging period (the time period during which GRACE observations contributing to a single global gravity field are retrieved). Information on this "instrument uncertainty" was provided by the GRACE science team (*S. Bettadpur*, The University of Texas, 2000, personal communication) and used to calculate the uncertainty in a 365-day GRACE measurement over the High Plains region, following the technique employed by *Rodell and Famiglietti* [1999]. Two GRACE measurements will be required to identify a change; therefore the following relation was used to calculate the expected instrument error, E_I, in an estimate of the change in water storage:

$$E_I = \sqrt{E_{I,1}^2 + E_{I,2}^2}\,, \tag{4.2}$$

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

Prior studies [*Wahr et al.*, 1998; *Rodell and Famiglietti*, 1999 and 2000] used data from the ECMWF Re-analysis [*ECMWF*, 1996] and the NCEP/NCAR Reanalysis [*Kalnay et al.*, 1996] to estimate atmospheric errors. The error in a single modeled pressure estimate was computed as

$$E_{A,i} = \frac{\left|\overline{P}_{ECMWF} - \overline{P}_{NCEP/NCAR}\right|}{\sqrt{2}}, \tag{4.3}$$

where $\overline{P}$ 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 $(\overline{P}_{ECMWF} - \overline{P}_{NCEP/NCAR})$, 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 water storage changes;

$$E_A = \sqrt{E_{A,1}^2 + E_{A,2}^2}\,, \tag{4.4}$$

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. This convention produced error estimates on the order of 30% of the model-predicted atmospheric pressure changes. However, (4.3) and (4.4) are less appropriate for estimating annual errors, because mean surface pressure is more stable from year to year than from month to month, and the two models agree more closely about the annual changes, despite discrepancies among individual values of total surface pressure. Using (4.3) and (4.4), annual atmospheric uncertainty in the High Plains region is predicted to be greater than 125% of the modeled pressure changes, despite the fact that those changes agree to within 15% on average. Therefore, in the present investigation, atmospheric uncertainty is based on the modeled surface pressure changes. Using the same logic as for (4.3), uncertainty in an estimate of the annual change in pressure is computed here as

$$E_A = \frac{\left|\Delta\overline{P}_{ECMWF} - \Delta\overline{P}_{NCEP/NCAR}\right|}{\sqrt{2}}\,, \tag{4.5}$$

where $\Delta\overline{P}$ is a modeled year-to-year change in atmospheric pressure.

Rodell and Famiglietti (2000) demonstrated that, in Illinois, reservoir and snow water storage changes were typically insignificant, especially on an annual basis, relative to soil moisture and groundwater storage changes. It is assumed that the same is true in the High Plains. Hourly and/or six-hourly soil moisture observations were averaged to produce monthly values, which were then converted to an equivalent depth of water in each soil layer by multiplying the

volumetric water content by the height of the layer. Total soil water storage was taken as the sum over all of the layers.

The average absolute annual change in soil water storage, $\overline{|\Delta SM|}$, was estimated for the region as follows. Computing a time series was infeasible because the soil moisture observations were not temporally congruent and spanned only two to four years at each monitoring location. For each monitoring location, L, 12 months, m, of total soil water storage values, $SM_{L,m}$, were averaged to produce annual means for each month that was at the end of a consecutive 12 month period of observations. Then, for each month, M, that was at the end of a consecutive 24 month period of observations, the annual change was computed as the difference between the latter and former 12 month averages. The absolute values of all of these end-month annual changes were then averaged for each location, where T was the number of such changes. The regional average absolute annual change in soil water storage was taken to be the arithmetic mean of the averages from the nine observation locations:

$$\overline{|\Delta SM|} = \sum_{L=1}^{9}\left[\sum_{M=24}^{23+T_L}\left|\left(\sum_{m=M-11}^{M} SM_{L,m}/12\right)-\left(\sum_{m=M-23}^{M-12} SM_{L,m}/12\right)\right|\Big/T_L\right]\Big/9 \qquad (4.6)$$

The effect of soil moisture changes will have to be removed from the GRACE gravity fields in order to isolate the groundwater signal. Land surface models will likely produce the required estimates, and errors in those estimates will propagate to the GRACE-derived groundwater storage changes. Therefore, the coefficient of variation of the ten GSWP-modeled estimates of the 1987 to 1988 annual change in root zone soil water in the High Plains was computed. Using that number as a guide, a coefficient of uncertainty for soil moisture

changes was selected. This coefficient of uncertainty was then multiplied by the observation-based $\overline{|\Delta SM|}$, resulting in an estimate of "soil moisture uncertainty", E_{SM}, as an equivalent height of water. This method was preferred to simply taking E_{SM} as the standard deviation of the model estimates, as was done to determine E_A, because the latter would necessitate a baseless assumption that the change in soil water storage between 1987 and 1988 was close to the average annual change. For each year, the total uncertainty, E_T, in a GRACE-derived change in groundwater storage was estimated conservatively by summing the three error components:

$$E_T = E_I + E_A + E_{SM}. \tag{4.7}$$

RESULTS

The expected instrument uncertainty for a change in the gravity field on a 365-day averaging period in a 450,000 km^2 region is 0.80 *mm* (equivalent height of water). Between 1987 and 1993 the estimated atmospheric uncertainty ranged from 0.38 *mm* to 1.63 *mm*. ECMWF Re-analysis data were not available after 1993, so, beginning with 1994, the average atmospheric uncertainty from 1979 to 1993, 0.72 *mm*, was applied.

Using (4.6), $\overline{|\Delta SM|}$ was estimated to be 24.0 *mm* in the High Plains. For reference, the 1987 to 1988 change in root zone soil moisture estimated by the GSWP models averaged -32.5 *mm* in the High Plains. Considering that 1988 was a drought year, so that a greater than normal loss of water storage would be expected, these two values compliment each other well. The coefficient of

variation of the GSWP-modeled 1987 to 1988 soil moisture changes in the High Plains was 0.48. However, two of the modeled changes were much larger than the other eight: -62.7 and -59.4 *mm*, compared to a mean of -32.5 *mm*. Eliminating the first of these from the calculation reduces the coefficient of variation to 0.42, and eliminating both reduces it to 0.20. According to *Entin et al.* [1999], the magnitude of the differences among the GSWP models is typically similar or larger than the magnitude of the differences between observations and any one model. Therefore, using an uncertainty coefficient of 0.30, which is in the range of the coefficients of variation computed above, is reasonable. Multiplying that value by the observation-derived $\overline{|\Delta SM|}$ produces an E_{SM} of 7.2 *mm*. Hence, E_{SM} is an order of magnitude larger than the instrument and atmospheric errors and will dominate the uncertainty in the technique.

Annual groundwater storage changes in the High Plains are shown in Figure 4.3. They range in magnitude from a minimum absolute change of 3.7 *mm*, between 1996 and 1997, to a maximum absolute change of 30.4 *mm*, between 1994 and 1995. The average change was a decrease in storage of 10.3 *mm* per year, resulting in a total loss of 113.7 *mm* over the eleven year period. The average absolute change was 19.8 *mm*. The error bars in Figure 4.3 represent $\pm E_T$, the total uncertainty in a hypothetical GRACE-derived estimate of a change in groundwater storage. For annual changes in the High Plains, E_T averaged 8.7 *mm*, so that the relative uncertainty, $E_T/\Delta GW$, averaged about 0.44. Ten of the eleven annual changes between 1987 and 1998 would be detectable by GRACE, given that detectability is achieved when the relative uncertainty is less than one.

The minimum duration of the GRACE mission will be five years. Therefore it is worth examining the detectability of 4-year changes in High Plains aquifer water storage, which also better reveal the trend of decline. Water storage changes were computed for each of the eight periods of five consecutive years between 1987 and 1998. These are shown in Figure 4.4, with error bars to depict the uncertainty in GRACE-derived estimates. Although the changes in question happen over the course of four years, the magnitude of the uncertainty is the same as that of the annual changes because the averaging period for each measurement is still one year. Each 4-year change in the timeframe of the study was a loss ranging in magnitude from 9.2 *mm* to 103.6 *mm*, with a mean absolute change of 45.2 *mm*. In six of eight cases the relative uncertainty was 0.50 or less, being 0.34 on average.

DISCUSSION

This investigation concludes that, with the aid of computer simulations for predicting atmospheric pressure and soil moisture, annual groundwater storage changes in the High Plains aquifer will be estimated to within approximately 8.7 *mm* using GRACE observations of the gravity field. It is unclear whether or not this is an improvement in accuracy over the well-monitoring-based method employed by the USGS. However, the prospect of satellite-based monitoring of groundwater is exciting and the GRACE technique does offer some significant advantages over the conventional method. First, it is not labor intensive and does not require an extensive network of wells. For this reason GRACE will be useful

for monitoring groundwater in parts of the world where overdraft is occurring but is not systematically documented. Significant groundwater level declines have been reported in northern China, India, and Saudi Arabia, all caused by overpumping for irrigation [*Postel*, 1993]. Similarly, GRACE will be used to estimate long-term changes in the volume of water stored in surface water bodies such as the Aral Sea, the volume of which decreased 65% between 1926 and 1990 [*Gleick*, 1993]. Second, it has been demonstrated by *Wahr et al.* [1998] and *Rodell and Famiglietti* [1999, 2000] that GRACE will be able to identify monthly and seasonal changes in terrestrial water storage. Therefore the potential exists for assessing sub-annual changes in groundwater storage, which may help to develop understanding of the seasonal cycle of groundwater level and the sensitivity of an aquifer, in terms of recharge and drawdown, to seasonal variability in precipitation and other climatic variables, groundwater pumpage, and irrigation. Third, GRACE may be able to discern storage changes in aquifer subregions. There will be a tradeoff between spatial resolution and accuracy, but some sections of the High Plains aquifer experience much larger storage changes than others [e.g., *McGuire et al.*, 1999], so that certain scenarios may produce a level of relative uncertainty similar to that described here.

Soil moisture error, E_{SM}, will dominate the uncertainty in GRACE-derived estimates of groundwater storage changes. Fortunately, several factors should improve the accuracy of soil moisture estimates in the near future, thereby reducing E_{SM}. Monitoring networks such as SCAN and ARM-SWATS have just recently come online. The AMSR-E remote sensor will begin to deliver global

observations of surface (about 1 *cm* depth) soil moisture after launch in December, 2001. Observations from these systems will be assimilated into models and will facilitate the evaluation of model performance. Furthermore, region-specific optimization, advances in modeling, and greater computing power should all contribute to improved soil moisture estimates.

A shortcoming of this investigation and the technique in general is that groundwater storage must be lumped together with intermediate zone water storage (i.e., water stored in the soil above the water table and below the soil measurement level) whenever the latter is not specifically observed or modeled. In other words, if observations and models are used to remove root zone (e.g., 0-2 *m* depth) moisture changes from the GRACE-derived terrestrial water storage change estimates, it still will be unclear to what extent the residual storage changes are attributable to water table variations rather than storage variations in the unsaturated region above. *Rodell and Famiglietti* (2000) concluded that intermediate zone water storage is a potentially significant component of terrestrial water storage that it is not well understood at this time. However, this situation should not be considered a major obstacle, because, above the field capacity, the water in the intermediate zone tends to drain to the water table. In effect, it is "groundwater in waiting".

As a final note, plans for a follow-on to the GRACE mission are already being discussed. The leading candidate mission, presently dubbed NASA EX-5, would replace the onboard microwave tracking system, which is the key to GRACE measurements, with a laser interferometer, thereby reducing errors by an

estimated four orders of magnitude [*Watkins et al.*, 2000]. Although soil moisture errors were the limiting factor in this study, *Rodell and Famiglietti* (2000) indicated that instrument errors begin to dominate the uncertainty budget as the spatial scale decreases below 300,000 km^2.

SUMMARY

Annual water table variations in the High Plains aquifer were estimated by the USGS from 1987 to 1998. These were converted to equivalent changes in groundwater storage for the purpose of evaluating the potential to estimate such changes using satellite-based gravity observations from NASA's GRACE mission. This prospect is enticing because airborne and satellite active and passive microwave remote sensing techniques currently in use can not detect water storage changes below the upper few centimeters of soil. Time series of observed and modeled soil moisture were used to assess the expected accuracy of simulated soil water storage, which simulations will be used to remove the effects of soil water variations from GRACE gravity measurements. Uncertainty in the soil moisture information, as well as uncertainty in modeled atmospheric fields and errors inherent to the GRACE instrument, will contribute to the total uncertainty in GRACE-derived estimates of changes in groundwater storage. Total uncertainty for annual estimates of storage changes in the High Plains aquifer was determined to be about 8.7 *mm*, whereas the magnitudes of the changes themselves averaged 19.8 *mm* on an annual basis and 45.2 *mm* for 4-year changes. From its satellite platform, GRACE is expected to provide estimates of

groundwater storage changes in parts of the world where well-monitoring networks do not exist, and assessment of sub-annual and sub-aquifer changes may be possible.

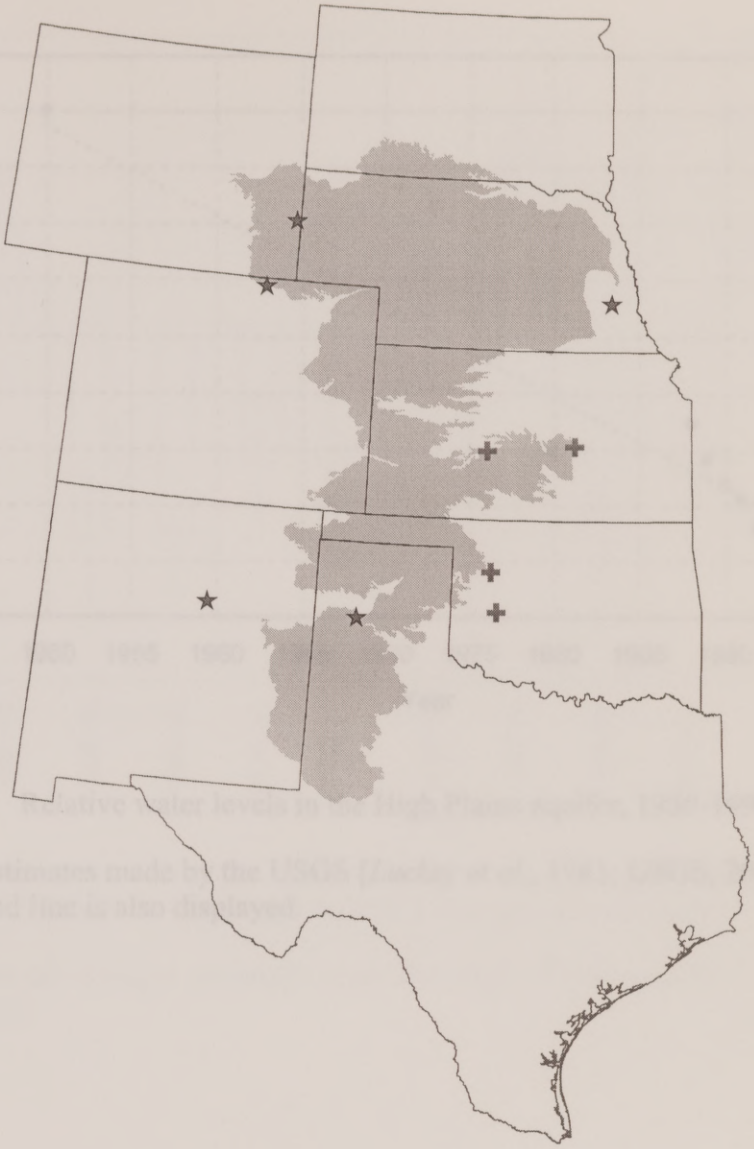

Figure 4.1: Map of the High Plains aquifer within the central United States.

The aquifer is shaded gray. Locations of SCAN (stars) and ARM-SWATS (crosses) soil moisture monitoring sites are marked.

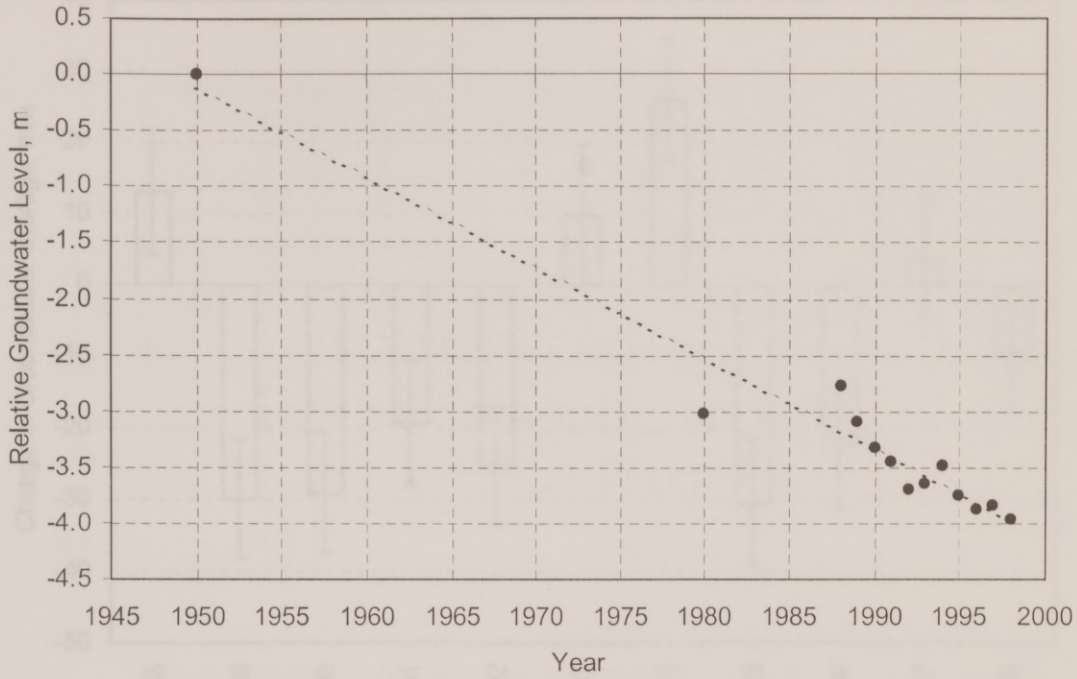

Figure 4.2: Relative water levels in the High Plains aquifer, 1950-1998.

Based on estimates made by the USGS [*Luckey et al.*, 1981; USGS, 2000]. Least-squares trend line is also displayed.

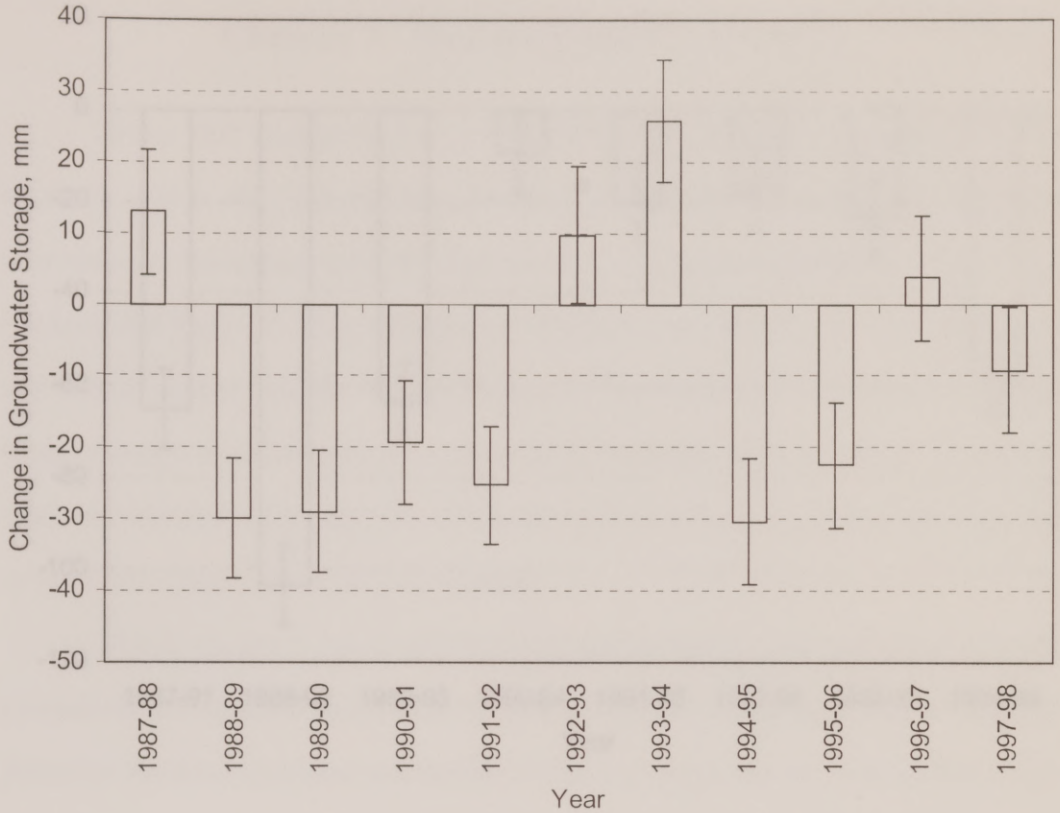

Figure 4.3: Annual changes in groundwater storage and uncertainty.

Time series of changes averaged over the High Plains aquifer. Error bars represent $\pm E_T$.

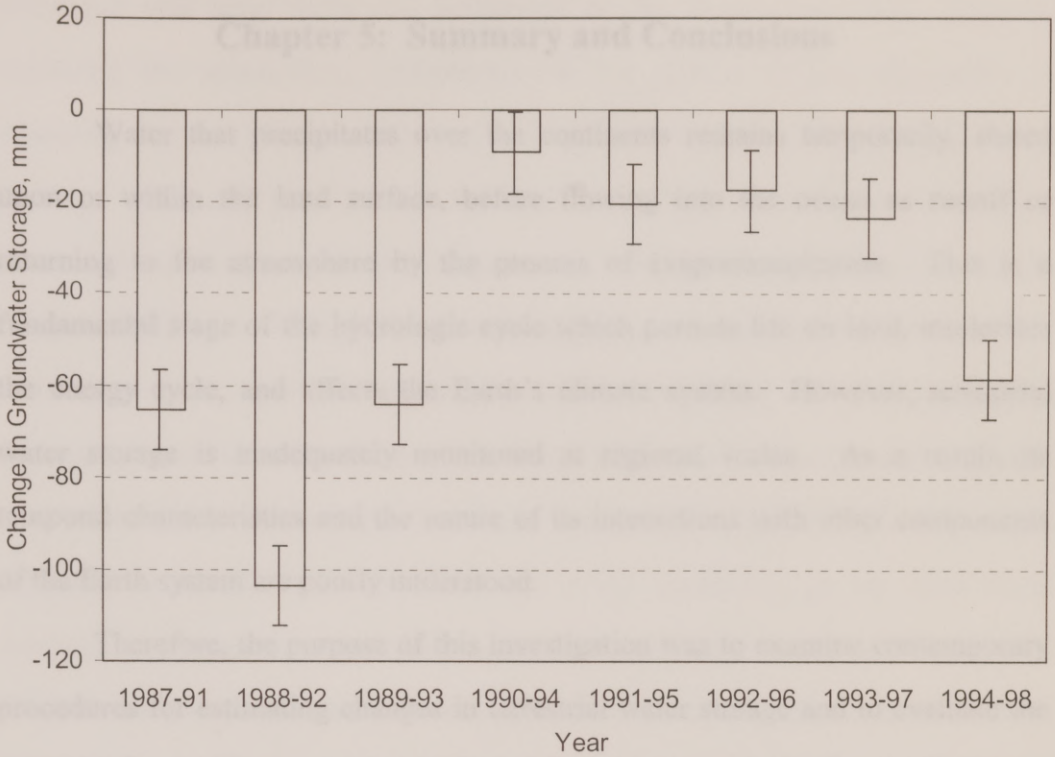

Figure 4.4: Four-year changes in groundwater storage and uncertainty.

Time series of changes averaged over the High Plains aquifer. Error bars represent $\pm E_T$.

Chapter 5: Summary and Conclusions

Water that precipitates over the continents remains temporarily, stored upon or within the land surface, before flowing into the ocean as runoff or returning to the atmosphere by the process of evapotranspiration. This is a fundamental stage of the hydrologic cycle which permits life on land, moderates the energy cycle, and affects the Earth's climate system. However, terrestrial water storage is inadequately monitored at regional scales. As a result, its temporal characteristics and the nature of its interactions with other components of the Earth system are poorly understood.

Therefore, the purpose of this investigation was to examine contemporary procedures for estimating changes in terrestrial water storage and to evaluate the potential for satellite-based monitoring of those changes via the Gravity Recovery and Climate Experiment (GRACE). The GRACE satellite mission, set for launch in 2001, will produce a new model of the Earth's gravity field, with unprecedented accuracy, at monthly and longer intervals. Because changes in terrestrial water storage cause variations in the Earth's gravity field, it will be possible to generate estimates of those changes from the gravity observations made by GRACE.

The investigation consisted of three separate studies. In the first of these, the best available modeled time series of soil moisture and snow were evaluated and then used to predict monthly, seasonal, and annual changes in terrestrial water storage in twenty continental-scale drainage basins. These were chosen to encompass a range of climates and spatial extents. Uncertainty in the GRACE

technique will arise from the limitations of the instrument itself, errors in removing the atmospheric contribution to the gravity signal, and errors in removing the contribution of post glacial rebound (PGR). Therefore, for each drainage basin and time period, the total error from these three sources was computed and compared to the estimated water storage change. Primary controls on detectability were determined to be the area of the region, which affects instrument errors, and the magnitude of the water storage changes themselves. Secondary controls were uncertainty in the atmospheric models and temporal resolution, which also affects instrument errors. Detectability thresholds were generally in the 200,000 – 500,000 km^2 range, depending on the latter three controls. Atmospheric errors dominated the uncertainty in regions larger than 400,000 km^2, while instrument errors dominated in regions smaller than 200,000 km^2. Uncertainty due to PGR was insignificant in most parts of the world.

Objectives of the second study were to examine groundwater and surface water variations, in addition to soil moisture and snow water variations, and to utilize a longer, observation-based time series. Therefore, more than 13 years of measurements of depth to groundwater, soil water content in the upper two meters, snow depth, and reservoir water level were obtained from observation stations in Illinois, which is possibly the only region of the world where all of these types of observations have been collected and made available. Water stored in the intermediate zone of the soil was estimated, and monthly time series of all five variables were constructed. Uncertainty in the GRACE technique was computed as before, except that PGR model errors were ignored. Because Illinois

is only 145,800 km^2 (instrument errors associated with a region that small are prohibitive) and to extend the usefulness of the data, it was assumed that the observed water storage changes were representative of five progressively larger areas. Detectability analyses were conducted for each. It was concluded that GRACE has the potential to detect terrestrial water storage changes of similar magnitude to those observed in Illinois if they occur in 200,000 km^2 and larger regions. Variations in water storage below two meters depth were determined to be of nearly the same magnitude as soil moisture changes in the upper two meters. Snow and reservoir water storage changes in Illinois were generally negligible relative to soil moisture and groundwater storage changes.

The aim of the third study was to evaluate the potential for isolating changes in groundwater from changes in the other components of terrestrial water storage, given observations from GRACE and additional information. The study relied on eleven years of annual water table elevation variations, averaged over the High Plains aquifer, which were estimated by the USGS based on well measurements. In order to isolate groundwater storage variations, the contribution of soil moisture will have to be removed from the total water storage signal, and uncertainty will arise from that procedure. Therefore, time series of observed and modeled soil moisture were used to assess the degree of uncertainty in modeled soil moisture. Based on the results of the second study, variations in snow and surface water storage were assumed to be insignificant. Annual changes in groundwater storage in the High Plains aquifer, of which the average magnitude was 19.8 *mm* between 1987 and 1998, were predicted to be detectable

by GRACE in most years, given that the uncertainty in resolving those changes was estimated to be 8.7 *mm* on average. Accordingly, it was anticipated that the GRACE technique could provide estimates of groundwater storage changes, which would be particularly valuable in parts of the world where well monitoring networks do not exist. Assessment of sub-annual and sub-aquifer changes, although not specifically explored, also may be achievable.

In addition to analyzing the potential to derive hydrologic quantities from GRACE satellite observations, the major contribution of this body of work is its treatment and characterization of terrestrial water storage as a singular component within the hydrologic cycle. Myriad studies have described the spatial and temporal variability of groundwater, soil moisture, snow, or surface water storage. Few have considered the combined effects of these components. Nevertheless, terrestrial water storage is one of the three stocks in the most basic description of the hydrologic cycle: water evaporates from the ocean, flows through the atmosphere as water vapor, precipitates onto the land surface, and eventually returns to the ocean as runoff. Characterization of the dynamics of terrestrial water storage should be a fundamental pursuit in the study of water.

Figure 3.2 provides an innovative perspective of the temporal behavior of terrestrial water storage, by superposing the shallower components on the deeper ones. The figure is based entirely on observations, which is an achievement in itself. Soil moisture and groundwater stand out as the dominant controls on total terrestrial water storage variability in Illinois. Furthermore, Figure 3.2 shows the manner in which variations in the individual components combine during events

such as the drought of 1988 and 1989 and the flood of 1993. It is noteworthy that the four components do not normally attain peak values (or lowest values) simultaneously. Figure 3.3 demonstrates that monthly changes in water storage below two meters depth lag behind changes in the upper soil zone. Figure 3.6 shows that the same may be true for annual changes.

Another unique contribution of this research is its analysis of the soil zone below two meters depth and above the top of the water table (i.e. the intermediate zone). The intermediate zone is a temporary reservoir for water that percolates down from the near-surface soil zone and that which remains as a residual when the water table descends. As such, it balances changes in these two adjacent stocks. This is apparent in Figure 3.2. The scientists who observe and model hydrological processes should pay more attention to intermediate zone water storage (it is typically ignored) if they wish to better quantify the terrestrial water balance and develop understanding of the cycles of flood and drought.

That the annual or multiannual change in terrestrial water storage is negligible, so that the left side of the water balance equation (1.1) can be set to zero, is an overly simplistic assumption that is nonetheless common. The results presented in Chapters 2, 3, and 4 establish that the regional average annual or multiannual change in water storage is significant more often than not. Figures 2.6 and 3.6 make it abundantly clear that the annual change is terrestrial water storage is rarely small enough to be considered insignificant, and Figures 4.2 and 4.3 demonstrate that, even on the scale of decades, the assumption is not reliable.

Because this research indicate that it will be possible to derive useful hydrologic information from future GRACE observations, suggestions are presented here for expediting the delivery of that information and maximizing its value. First, an automated technique should be developed to use each new GRACE product, as it becomes available, to estimate terrestrial water storage changes in strategically located regions. Second, sources of groundwater, soil moisture, snow depth, and surface water observations should be identified around the world. These will aid in decomposing GRACE-derived terrestrial water storage changes. Finally, employing all GRACE-derived and auxiliary data as constraints in a hydrologic model is recommended in order to optimize the final product.

If GRACE is a success, follow-on missions, such as NASA EX-5, will likely be funded. Owing to the replacement of microwave ranging with laser interferometry and other technological advancements, these future missions are expected to have accuracies that are orders of magnitude better than GRACE [*Watkins et al.*, 2000]. Furthermore, the evolution of global climate models, through improved computational capacities, better representation of meteorological processes, and assimilation of observations from new sources, will reduce the uncertainty in simulating atmospheric mass redistribution. Therefore, if the method continues to be researched and refined, hydrologic data may become a standard derivative of satellite-based gravity measurements in the near future.

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