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# The potential for satellite-based monitoring of groundwater storage changes using GRACE: the High Plains aquifer, Central US

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## Abstract

Groundwater storage in the High Plains aquifer has been steadily decreasing for 50 or more years due to withdrawals for irrigation. This trend has been documented in annually published United States Geological Survey reports of water level changes in the High Plains aquifer, but assessments of groundwater storage changes in other parts of the world are incomplete. NASA's gravity recovery and climate experiment (GRACE) soon may provide an alternative means for monitoring groundwater changes, via satellite remote sensing. That terrestrial water storage changes are likely to be detectable by GRACE satellites has been demonstrated by prior studies. This investigation builds on those studies 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. © 2002 Elsevier Science B.V. All rights reserved.

*Keywords:* Groundwater; Soil moisture; High Plains aquifer; Remote sensing; Gravity

## 1. Introduction

The High Plains aquifer underlies 450,000 km<sup>2</sup> of the central United States (Fig. 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. Thus, 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 50 years of the 20th 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<sup>3</sup> 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. From 1950 to 1980 the average rate of water table decline was about 0.10 m per year (Luckey et al., 1981).

In response to growing concerns about groundwater depletion, the United States Geological Survey

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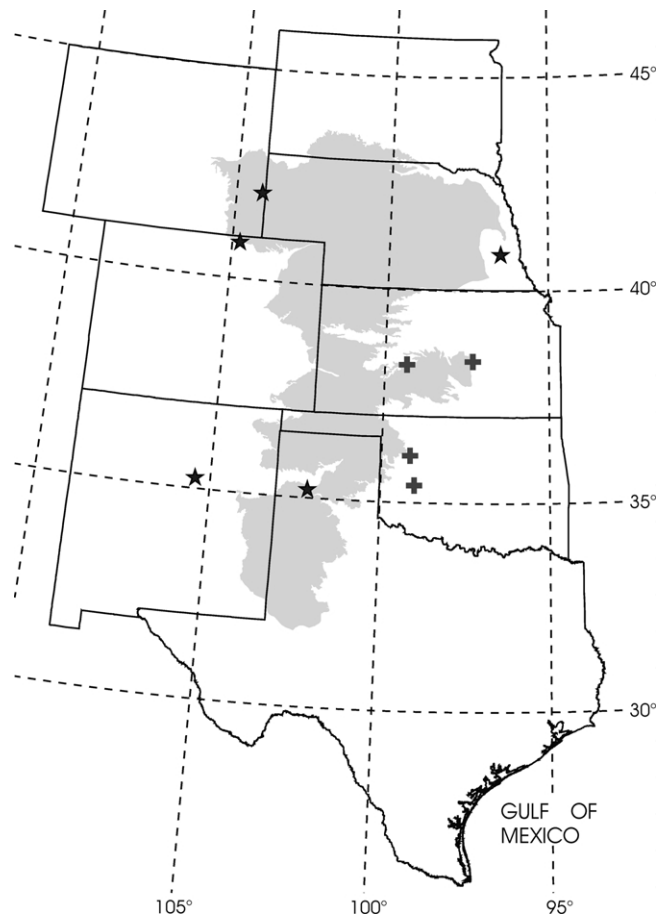


Fig. 1. Map of the central United States, High Plains aquifer shaded gray, showing locations of SCAN (stars) and ARM-SWATS (crosses) soil moisture monitoring sites.

(USGS), in cooperation with Federal, State, and local agencies, has been assessing changes in High Plains aquifer water levels annually since 1988. Measurements from over 6200 wells scattered across the region are used. 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 aquifer systems are not surveyed regularly and methodically, and without a centralized database of well records, 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 March,

2002, when NASA's gravity recovery and climate experiment (GRACE) commences. GRACE will launch two satellites into 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 5 year duration of the mission. By examining a GRACE-derived gravity model from each of two time periods (e.g. consecutive months), it will be possible to calculate, for a particular region, the change in mass 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. (It will not be possible to infer the absolute mass of water storage itself.)

Rodell and Famiglietti (2001) 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 2 m. 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<sup>2</sup> and larger regions. However, the likelihood of disaggregating the components of terrestrial water storage, including groundwater, was not explored explicitly. This paper evaluates the potential to estimate annual changes in the groundwater 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 also has 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 et al., 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 (scheduled to launch in 2002), the significance of this work is that GRACE will provide the first opportunity for monitoring deeper groundwater fluctuations from space. This capability cannot be proven until real data begins to flow from the mission, perhaps in early 2003, which is a fact that must be weighed when considering the results described herein. Nevertheless, it is important that hydrologists be aware of this potential new source of information.

## 2. Background

In this paper, 'High Plains' and 'High Plains region' will refer to the area underlain by the High Plains aquifer as shown in Fig. 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 500 mm with a range of about 300 mm 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 ranges from 1500 mm in the north to 2700 mm in the south. 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<sup>2</sup> 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 coarse material to replenish the aquifer (Dornbusch et al., 1995).

The High Plains aquifer is a 450,000 km<sup>2</sup> aquifer system that lies beneath parts of Nebraska, Texas, Kansas, Colorado, Wyoming, Oklahoma, New

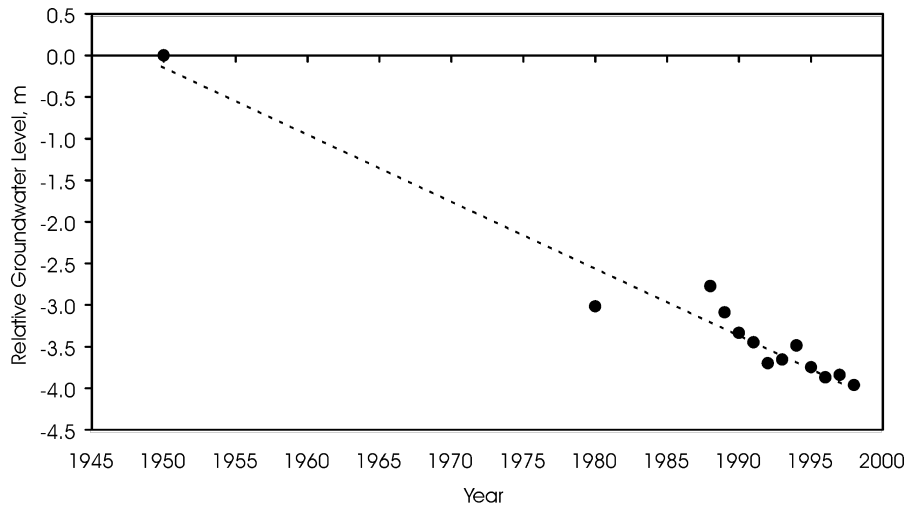


Fig. 2. Relative water levels in the High Plains aquifer between 1950 and 1998, based on estimates made by the USGS [Luckey et al., 1981; United States Geological Survey, 2000], with least-squares trend line.

Mexico, and South Dakota (Fig. 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 connected hydraulically 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 0–300 m, and the total volume of saturated material was estimated to be 26,800 km<sup>3</sup>

(Miller and Appel, 1997). Specific yield varies from less than 5% to greater than 30% and hydraulic conductivity varies from 7 to 90 m/day due to the non-uniformity of the aquifer materials. The water table is generally parallel to the land surface, sloping at about 1.9–2.8 m/km, so that water flows from west to east at about 0.30 m/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<sup>3</sup> of water from the High Plains aquifer to irrigate 56,000 km<sup>2</sup> (Weeks, 1986). Irrigation accounts for about 95% of the groundwater withdrawn. Prior to development, which began in the late 1800s and accelerated during droughts in the 1930s and 1950s, 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, withdrawals 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<sup>3</sup>/day to a 1960–1980 rate of 14,400,000 m<sup>3</sup>/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 beginning in 1950 (Fig. 2) show an average decline of 8 cm/year.

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 also was 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 6 years of satellite laser ranging to the Lageos 1 and 2 satellites to demonstrate that their orbits were measurably perturbed by changes in water storage on earth. Rodell and Famiglietti (2001) incorporated groundwater storage information in 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 paper will extend previous work by assessing the potential to isolate annual changes in High Plains aquifer groundwater storage from the other factors that will affect the GRACE gravity measurements, including variations in root zone soil moisture.

### 3. Data

The USGS 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). These assessments form the basis of

the current investigation. Information on soil moisture variations in the region and the ability to estimate those variations also was required in order to assess the potential to disaggregate groundwater from total water storage changes. Soil moisture observations from nine sites in the general region were used (Fig. 1). The soil climate analysis network (SCAN) of the national resource conservation service (Schaefer et al., 1995) collects 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 are 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 time series from the global soil wetness project (GSWP) (Dirmeyer et al., 1999) helped to gauge the precision of hydrological models in simulating soil water storage (see Section 4). GSWP is an ongoing global energy and water cycle experiment (GEWEX) collaboration 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, 10 contributing groups used different land surface schemes to produce time series of soil moisture, snow, and other variables on a 1° global grid. The 10 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 models and the results of the pilot phase of GSWP.

#### 4. Methods

Each USGS estimate of a change in annual average High Plains aquifer water level was converted to a change in groundwater storage,  $\Delta GW$ , 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

$$\Delta GW_{1998} = S_Y \Delta H_{WT1998}, \quad (1)$$

where  $S_Y$  is the specific yield and  $\Delta H_{WT}$  is the 1998 annual change in the mean height of the water table estimated by the USGS. Conceptually,

$$\Delta H_{WT1998} = \overline{H_{WT1998}} - \overline{H_{WT1997}}, \quad (1a)$$

where  $\overline{H_{WT}}$  is the average height of the water table in 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 most likely will be accomplished using modeled fields of atmospheric pressure (a surrogate for atmospheric mass) 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), although Velicogna and Wahr (2001) suggested that sufficiently dense barometric networks could provide the pressure fields. Errors in the pressure fields will propagate to the terrestrial water storage change estimates. This ‘atmospheric uncertainty’ and the limitations of the GRACE instruments themselves, the ‘instrument uncertainty’, will be the two most significant factors restricting the accuracy of terrestrial water storage change estimates (Rodell and Famiglietti, 1999). 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). Other gravitational effects, including planetary motions and solid earth tides, are well described and not considered significant sources of uncertainty. Diurnal atmospheric and

hydrologic mass variations may distort the GRACE-derived terrestrial water storage changes because satellite overpasses will not be frequent or regular. This effect is not considered here because it will be minimized by a long (annual) averaging period and mitigation techniques, which are under development. To isolate the groundwater component of terrestrial water storage, it will be necessary to estimate and remove the gravitational effect of changes in the other components, causing additional uncertainty.

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 recovered). Information on the expected instrument uncertainty was provided by the GRACE science team (S. Bettadpur, The University of Texas, 2000, personal communication) and used to calculate the error 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:

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

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, 2001) used data from the ECMWF Reanalysis (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

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

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

storage change estimates;

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

where  $E_A$  is the atmospheric error in an estimate of water storage change, and  $E_{A,1}$  and  $E_{A,2}$  are the atmospheric errors in GRACE measurements for averaging periods 1 and 2, respectively. However, Eqs. (3) and (4) are less appropriate for approximating annual errors than monthly or seasonal errors, because over longer periods mean surface pressure is less variable, and hence, more accurately modeled, so that uncertainty is diminished. In fact, the models exhibit greater discrepancies between estimates of annual average surface pressure than estimates of the change in annual average surface pressure. A probable explanation for this behavior is that the models run on different spatial grids and use different elevation definitions, which would affect absolute pressure but not pressure changes. In the present investigation, atmospheric uncertainty was based on modeled surface pressure changes. Using the same logic as for Eq. (3), uncertainty in an estimate of the annual change in pressure was computed here as

$$E_A = \frac{|\Delta\bar{P}_{\text{ECMWF}} - \Delta\bar{P}_{\text{NCEP/NCAR}}|}{\sqrt{2}}, \quad (5)$$

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

In order to isolate groundwater storage changes from future GRACE-derived total water storage changes, changes in the other components will have to be estimated and removed. Rodell and Famiglietti (2001) demonstrated that, in Illinois, reservoir and snow water storage changes are typically insignificant, especially on an annual basis, relative to soil moisture and groundwater storage changes. It was assumed that the same is true in the High Plains, leaving only soil moisture changes to extract. The uncertainty in performing that extraction was simulated as described in the remainder of this section. For each of the nine monitoring locations, 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 fractional volumetric water content by the height of the layer. Total soil water storage was taken as the sum over all of the observed layers. Due

to a lack of deeper observations, changes in soil water were not quantified in the zone below the lowest observation layer and above the top of the water table. Effectively, those ‘intermediate zone’ changes were lumped together with groundwater storage changes in this investigation. When the time comes to analyze actual GRACE-derived total water storage variations, intermediate zone water storage will have to be modeled, measured, or acknowledged as a component of groundwater storage in order to account for its contribution.

The average absolute annual change in soil water storage,  $|\overline{\Delta\text{SM}}|$ , was estimated for the region. Developing a regional mean time series was infeasible because the soil moisture observations were not temporally congruent in spanning 2–4 years between 1996 and 2000. For each monitoring location,  $L$ , 12 months,  $m$ , of total soil water storage values,  $\text{SM}_{L,m}$ , 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 observation, the annual change was computed as the difference between the later and earlier 12 month averages. The absolute values of all end-month annual changes then were 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:

$$|\overline{\Delta\text{SM}}| = \sum_{L=1}^9 \left[ \sum_{M=24}^{23+T_L} \left| \left( \sum_{m=M-11}^M \text{SM}_{L,m}/12 \right) - \left( \sum_{m=M-23}^{M-12} \text{SM}_{L,m}/12 \right) \right| / T_L \right] / 9. \quad (6)$$

Land surface models likely will produce the fields necessary to remove the effect of soil moisture from the gravity signal, and errors in those fields will propagate to the GRACE-derived groundwater storage changes. To estimate the errors, the coefficient of variation of the 10 GSWP-modeled estimates of the 1987–1988 annual change in root zone soil water in the High Plains was computed and used as a guide in selecting the uncertainty coefficient for soil moisture changes. The uncertainty coefficient then was



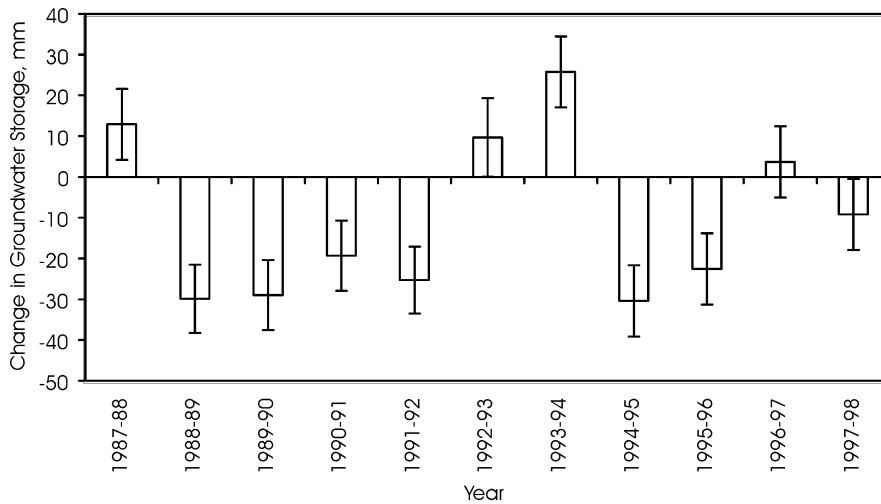


Fig. 3. Annual changes in High Plains aquifer water storage with error bars that represent the total uncertainty in GRACE-derived estimates.

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 suppose that the change in soil water storage between 1987 and 1988 equaled 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:

$$E_T = E_I + E_A + E_{SM}. \quad (7)$$

## 5. Results

The expected instrument uncertainty for a change in the gravity field on a 365-day averaging period in a 450,000 km<sup>2</sup> region is 0.80 mm (equivalent height of water). Between 1987 and 1993 the estimated atmospheric uncertainty ranged from 0.38 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 used in the total uncertainty calculation.

Using Eq. (6),  $|\overline{\Delta SM}|$  was estimated to be 24.0 mm in the High Plains. For reference, the 1987–1988 change in root zone soil moisture estimated by the

GSWP models averaged  $-32.5$  mm in the High Plains. Considering that 1988 was the first year of a drought, so that a greater than normal loss of water storage would be expected, these two values complement each other well. The coefficient of variation of the GSWP-modeled 1987–1988 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 extremes 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 to 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 earlier, 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 Fig. 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

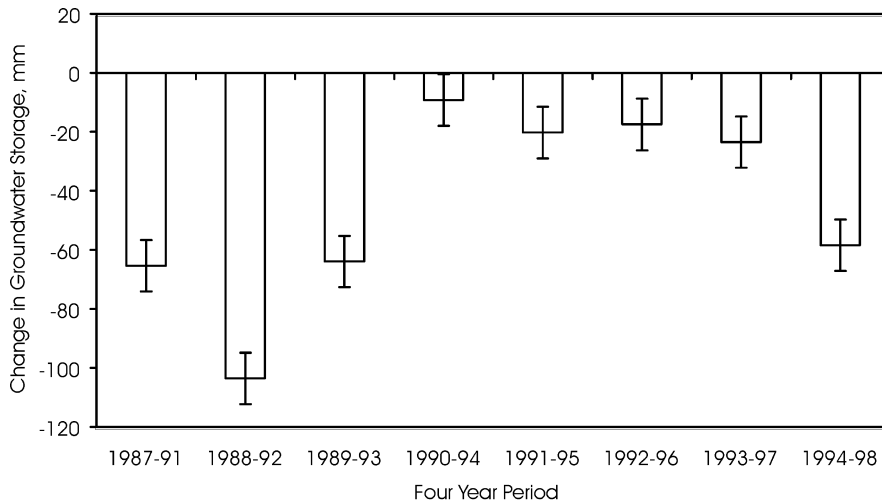


Fig. 4. Four year changes in High Plains aquifer water storage with error bars that represent the total uncertainty in GRACE-derived estimates.

year, resulting in a total loss of 113.7 mm over the 11 year period. The average absolute change was 19.8 mm. The error bars in Fig. 3 represent  $\pm E_T$ , the estimated total uncertainty in a hypothetical GRACE-derived estimate of the change in groundwater storage. For annual changes in the High Plains,  $E_T$  averaged 8.7 mm, so that the relative uncertainty,  $E_T/\Delta GW$ , was 0.44 on average. 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 5 years, so it is worth examining the detectability of 4 year changes in High Plains aquifer water storage. Water storage changes were computed for each of the eight periods of five consecutive years between 1987 and 1998. These are shown in Fig. 4, with error bars to depict the uncertainty in GRACE-derived estimates. Although the changes happen over the course of 4 years, the magnitude of the uncertainty is the same as that of the annual changes because the averaging period for each measurement is still 1 year. Each 4 year change in the timeframe of the study was a loss ranging in magnitude from 9.2 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. Therefore, using 5 years of GRACE data in this manner will hint at long term,

regional groundwater storage trends in the High Plains and other parts of the world.

## 6. 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 about 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 intriguing because it is not labor intensive and does not require an extensive network of wells.

Much of the world is experiencing a water crisis that is better attributed to mismanagement than scarcity of water, therefore, any new and objective method for monitoring water resources will be valuable to the world's people and international organizations such as the United Nations Food and Agriculture Organization (World Water Council, 2000). Significant groundwater level declines have been reported not only in the US High Plains, but in aquifer systems in northern China, India, and Saudi Arabia, all caused by withdrawal for irrigation (Postel, 1993). The

GRACE technique may reveal groundwater depletion occurring in parts of the world where it is either not systematically documented or not disclosed for political or economic reasons.

It has been demonstrated by Wahr et al. (1998), Rodell and Famiglietti, 1999, 2001) 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 will help to develop understanding of seasonal cycles of groundwater and the sensitivity of regional water table levels to variability in precipitation and other climatic variables, groundwater pumping, and irrigation. GRACE also may be able to discern storage changes in aquifer system subregions. There will be a tradeoff between spatial resolution and the accuracy of the estimates, but some sections of the High Plains aquifer experience much larger storage changes than others (e.g. McGuire et al., 1997), so that certain scenarios may produce an acceptable level of accuracy.

Soil moisture error,  $E_{SM}$ , will dominate the uncertainty in GRACE-derived estimates of groundwater storage changes. Fortunately, several factors will increase the accuracy of soil moisture simulations in the near future, thereby reducing  $E_{SM}$ . The EOS AMSR-E remote sensor will deliver global observations of surface (about 1 cm depth) soil moisture. These remotely sensed observations will be assimilated into land surface models, helping to constrain estimates of  $|\overline{\Delta SM}|$  (e.g. Walker and Houser, 2001). Monitoring networks such as SCAN and ARM-SWATS have come online very recently. Ground-based soil moisture observations will facilitate the evaluation and enhancement of land surface model performance. Furthermore, region-specific optimization, advances in modeling, and greater computing power all will improve estimates of  $|\overline{\Delta SM}|$ .

A shortcoming of this investigation and the technique in general is that groundwater storage must be lumped together with intermediate zone soil water storage 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 will not be clear to what extent the residual storage changes are attributable to water table variations rather than

unsaturated storage changes above the water table but below the root zone. Rodell and Famiglietti (2001) concluded that intermediate zone soil water storage is a potentially significant component of terrestrial water storage which is not well understood at this time.

As a final note, plans for a follow-on to the GRACE mission already are being developed. The leading candidate mission would replace the onboard microwave tracking system, which is the key to GRACE measurements, with a laser interferometer, thereby greatly reducing instrument errors. (Watkins et al., 2000). Although soil moisture errors were the limiting factor in this study, Rodell and Famiglietti (2001) indicated that instrument errors increase rapidly as the spatial scale decreases below 300,000 km<sup>2</sup>, so that the smallest area in which water storage changes are resolvable by GRACE will be about 200,000 km<sup>2</sup>. The follow-on mission would have a minimum resolvable area between 1000 and 20,000 km<sup>2</sup>.

## 7. 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 current airborne and satellite active and passive microwave remote sensing techniques cannot detect water storage changes below the top few centimeters of soil. Time series of observed and modeled soil moisture were used to predict the accuracy of soil moisture simulations, which will be used to remove the effects of soil water variations from GRACE gravity measurements. Uncertainty in the soil moisture simulations, 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 groundwater 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 likely will provide estimates of groundwater storage changes for all parts of the world, including regions where well-monitoring networks do not exist.

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