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INTEGRATED APPROACH FOR HYDROGEOLOGIC INVESTIGATIONS IN AFRICA: INFERENCEs FROM SPACE-BORNE AND LAND-BASED GRAVITY, AEROMAGNETIC, GIS, AND REMOTE SENSING DATA

by

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Advisor: Mohamed Sultan, Ph.D.

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INTEGRATED APPROACH FOR HYDROGEOLOGIC INVESTIGATIONS IN AFRICA: INFERENCES FROM SPACE-BORNE AND LAND-BASED GRAVITY, AEROMAGNETIC, GIS, AND REMOTE SENSING DATA

Mohamed El Sayed Ahmed Ahmed
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An integrated (space-borne and land-based gravity, aeromagnetic, GIS, and remote sensing) approach was developed, tested, and applied to investigate the following on regional scales (Africa) or local scales (El Qaa Plain, Sinai): (1) assess the capability of GRACE data (04/2002 to 08/2011) for monitoring elements of hydrologic systems on the sub-basin level (Task 1); (2) address the nature and the controlling factors (e.g., climatic and/or human pressure-related) affecting the inter-annual GRACE-derived Terrestrial Water Storage (TWS) trends over a large suite of hydrologic systems and domains in (Africa) (Task 2); and (3) identify the structural, geologic and hydrologic settings as well as quantify the characteristics (e.g., geometry, volume of water, depletion rates) of the El Qaa Plain aquifer (Task 3).

Findings (1-7) for Tasks 1 & 2 include: (1) observed temporal mass variations are largely controlled by elements of the hydrologic cycle; (2) large sectors of Africa are undergoing statistically significant variations (+36 mm/yr to −16 mm/yr) due to natural and man-made causes; (3) warming of the tropical Atlantic ocean intensified Atlantic monsoons and increased precipitation and TWS over western and central Africa’s coastal plains, proximal mountainous source areas, and inland areas, whereas warming in the central Indian Ocean decreased precipitation and TWS over eastern and southern Africa; (4) the high frequency of negative phases of the North Atlantic
Oscillation (NAO) increased precipitation and TWS over northwest Africa; (5) deforestation in the Congo Basin decreased TWS in that area; (6) the construction of dams increased TWS in upstream Nile Valley countries; (7) consideration should be given to using GRACE TWS data as an alternative, viable index for measuring temporal and spatial variations in aridity. Findings (8 – 10) pertaining to structural, geologic and hydrologic settings and characteristics of El Qaa Plain (Task 3) include: (8) delineation of two connected sub-basins (average length: 48 km; average width: 16 km; maximum thickness: 3.75 km); (9) volume of water in storage was estimated at 40 to 56 km³; and (10) aquifers are being depleted through natural discharge and extraction at a rate of 19 x 10⁶ m³/yr.
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All praise is to God, the Almighty, for granting me the health and the patience to complete this work. To be honest, this is the most difficult part to write in my thesis. Thanking people who contributed to my success is not an easy task. Many people helped and supported me to reach this stage. For those groups of people that I do not specifically mention, I say thank you. I will try my best to pass my deepest thanks in a few words, but I know what they gave me is much more than words to be written.

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Mohamed E. Ahmed
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CHAPTER I
INTRODUCTION

1.1 Rationale

Access to potable fresh water resources is a human right and a basic requirement for economic development in any society. Fresh water resources are to be found in surface water systems (e.g., lakes, rivers, etc.) or in aquifers with varying areal extent (i.e., local or regional). The characterization and understanding of the geologic and hydrologic settings of, and the controlling factors affecting, these systems is gaining increasing importance due to the challenges posed by increasing population across the world.

In Africa, there are immense natural renewable and fossil fresh water resources. Yet, natural phenomena (e.g., rainfall patterns and climate change) together with human-related factors (e.g., population growth, unsustainable over-exploitation, and pollution) are increasingly threatening the sustainability of these resources, and hence the livelihoods of populations across the world. Hence, there is a real need for conducting scientific research to assess the current and projected impacts of these natural and anthropogenic parameters on hydrologic settings worldwide. Such research will generate datasets, tools, and interpretations that could be of use to the decision makers as they engage in developing plans and policies for the assessment, protection, and utilization of fresh water resources. In this study I target the hydrologic systems in the African continent.
I adopt an integrated approach including the space-borne and land-based gravity, aeromagnetic, remote sensing and GIS techniques to investigate the spatial distribution, geologic and hydrologic settings, and factors controlling the temporal and spatial distribution of fresh water resources in Africa at varying scales, namely on the regional (e.g., north, central, eastern and western Africa) and the local (El Qaa Plain, Sinai, Egypt) scales. The adopted integrated approach is different from many of the previous investigations; I adopt an integrated approach that utilizes a number of relevant data sets, whereas many of the previous investigations utilized individual data sets including geophysical, geochemical, climatic, or remote sensing data.

The main data set and methodology I utilize is gravity, a nondestructive geophysical tool that measures the differences in the earth’s gravitational field at a specific location (Telford et al., 1990). These measurements are used to determine the spatial variations in subsurface density. Types of gravity surveys include conventional land surveys, regional land surveys, shipboard gravity surveys, and satellite gravimetry (Lillie, 1999; Telford et al., 1990). For regional scale (Africa) applications, the satellite gravity data from the Gravity Recovery and Climate Experiment (GRACE) satellite mission were used. For smaller local scale (El Qaa Plain, Sinai, Egypt) applications, I resorted to the use of land-based gravity surveying along with aeromagnetic data.
1.2 GRACE

The GRACE satellite mission is a joint project between the National Aeronautic and Space Administration (NASA) in the United States and the Deutschen Zentrum für Luftund Raumfahrt (DLR) in Germany. GRACE was launched on March 17, 2002, as a five year mission, to map: (1) the temporal variations in Earth’s global gravity field on monthly basis with unprecedented accuracy and with a spatial resolution ranging from 400 to 40000 km, and (2) the mean Earth’s gravity field (Schmidt et al., 2008; Tapley et al., 2004a; Tapley et al., 2004b; Wahr et al., 2004).

The GRACE mission consists of twin (GRACE A and GRACE B) satellites flying in a polar orbit with an inclination of 89.5° and altitude of 500 km (at launch) and approximately 200 km apart from each other (Fig.1). The satellite altitude decays naturally (~30 m/day) so that the ground track does not have a fixed repeat pattern (Tapley et al., 2004b). The two satellites follow the same path and are interconnected by a K-band (frequency: 12 to 63 GHz) microwave link (Bettadpur, 2006; Flechtner, 2005). The orbits of the two spacecraft are affected differently due to the spatial and temporal variations in the Earth’s gravity field, which can increase or decrease the distance between them. These distance variations affect the travelling time of the microwave signals which are constantly being transmitted and received between satellites. The accuracy of the measured separation distance is as small as 1 micron, about 1 over 100th the thickness of a human hair. When approaching an area of higher gravity/mass, the front satellite, GRACE A, speeds up and then slows down when
flying above it. The second satellite, GRACE B, on the trail of GRACE A will have the same fate.

![GRACE Mission](image)

Figure 1: Gravity Recovery and Climate Experiment (GRACE) flight configuration. Source: http://www.csr.utexas.edu/grace/.

The disadvantage of having such a low (500 km) altitude is that GRACE experiences greater atmospheric drag, which can cause large and unpredictable changes in the inter-satellite range distance. To overcome this effect, each GRACE satellite has an on-board accelerometer to measure non-gravitational accelerations. Those measurements are transmitted to the ground where they are used to correct the
inter-satellite distance measurements. The orientation of the spacecraft in space is measured using star cameras. Each spacecraft also has an on-board GPS receiver, used to determine the orbital motion of each spacecraft in the global GPS reference frame and to improve the gravity field solutions at global-scale wavelengths (Bettadpur, 2006; Flechtner, 2005; Tapley et al., 2004b).

The GRACE raw data, collected from the two satellites, is calibrated and time-tagged in a non-destructive (or reversible) sense, and labeled Level-1A. The GRACE Level-1A data products are not distributed to public. The GRACE Level 1B data includes, among others, satellite-to-satellite distance, non-gravitational acceleration, spacecraft attitude. GRACE project also delivers Level-2 data products which consist of complete sets of harmonic coefficients out to maximum degree and order averaged over monthly intervals (Bettadpur, 2007, Tapley et al., 2004; Dunn et al., 2003). Level-2 data is distributed from three locations: (1) University of Texas Center of Space Research (UTCSR), (2) GeoForschungsZentrum (GFZ), and (3) Jet Propulsion Laboratory (JPL). For the convenience of GRACE users who would prefer to access GRACE data products as mass anomalies (e.g., water layer), the GRACE project also provide Level-3 data products in the form of equivalent water thicknesses.

The variability in the GRACE gravity field solutions represent geophysical responses associated with redistribution of mass at or near the Earth’s surface. That is, only in near surface locations, are mass variations likely to occur at the time scales (days to years) examined by GRACE measurements. Generally, the largest time-variable gravity signals observable in the GRACE data are expected to come from
changes in the distribution of water and snow stored on land (Wahr et al., 1998). The
temporal variations in the Earth’s gravity field can be used to make global estimates
of the Terrestrial Water Storage (TWS) (Wahr et al., 2004) in an area, a term that
refers to the total vertically-integrated water content in an area regardless of the
reservoir (e.g., surface water, groundwater, soil moisture and permafrost, snow and
ice, and wet biomass), in which it resides.

1.3 Problem Definition

1.3.1 GRACE: A Tool for Monitoring Elements of Hydrologic Systems on Regional
Scales

Since the launching of GRACE satellite mission, a new resourceful and
powerful tool became available to the scientific community, and numerous studies
addressing the potential hydrological applications of GRACE have been published.
The majority of these studies that utilize GRACE data for hydrological research and
applications (e.g., Chen et al., 2005; Rodell et al., 2004a) target very large watersheds
(areas of 450,000–6,000,000 km²), due to the fact that the accuracy of the recovered
mass variations from GRACE increases with increasing size of the monitored basin
(Wahr et al., 2006).

The problem is the majority of the world’s watersheds are much smaller
(<450,000 km²), and even for large watersheds one often needs to understand the
partitioning of water on the sub-basin level. In chapter III, I demonstrate that temporal
mass variations from the GRACE data acquired over the African continent, when
smoothed using a 250-km radius Gaussian function, are largely controlled by
elements of the hydrologic cycle, and have not been obscured by noise as previously thought.

1.3.2 GRACE: A Tool for Monitoring Water Availability

Drought indices (meteorological, agricultural and hydrological) define departures from normal conditions and are used as proxies for monitoring water availability (e.g., Dai, 2011b). There are uncertainties associated with such applications. Many of them exclude significant controlling factor(s) (e.g., Edwards and McKee, 1997), do not work well in specific settings and regions (Alley, 1984), and often require long (≥50 yr) calibration time periods and substantial meteorological data, limiting their application in areas lacking adequate observational networks (Smakhtin and Hughes, 2004). Additional uncertainties are introduced by the models used in computing model-dependent indices (Doesken et al., 1991). Aside from these uncertainties, none of these indices measure the variability in TWS. GRACE-derived TWS is, I believe, an adequate indicator for monitoring water availability, spatially and temporally, in varied hydrological and climatological settings.

In Chapter IV, I concluded that, given the 10-year monthly GRACE record of water availability data (represented by GRACE TWS) acquired on the sub-basin scale across the globe, and the plans underway for deployment of a GRACE follow-up mission (2016–2026), consideration should be given to using GRACE TWS data as an alternative, viable index for measuring temporal and spatial variations in aridity.
1.3.3 GRACE: A Tool for Monitoring Groundwater Depletion

Groundwater has been identified as one of the major sources of fresh water that can potentially meet the growing demands of Egypt’s population. In Egypt, the coastal plains along the Mediterranean Sea, the Gulf of Suez and the Gulf of Aqaba receive the highest amount of precipitation and are thus likely to have substantial aquifer storage. Along the Red Sea and Gulf of Suez coastal plains, many fresh water aquifers have been identified beneath the alluvium deposits covering the mid-Tertiary rift-related structural basins (Sultan et al., 2011b). The structural and hydrogeologic settings, the areal distribution and geometry, and the water volume of the El Qaa Plain aquifer were addressed in this study (chapter V) using both land-based gravity and aeromagnetic data. The aquifer depletion rates were estimated by addressing the temporal variations in GRACE monthly data.

1.4 Objectives

On the regional scale (Africa), GRACE data was used in conjunction with other readily available relevant data sets (e.g., topography, geology, remote sensing) to address the following tasks:

(1) Understand the nature and the controlling factors affecting observed temporal GRACE mass variations over the major African watersheds (i.e., the Nile, Niger, and Congo Basins) on sub-basin scales;
(2) Examine the capability of GRACE data for monitoring elements of hydrologic systems (i.e., runoff, recharge, groundwater flow) on the sub-basin level;

(3) Extract climatic inter-annual trends (e.g., wetness, dryness) from GRACE data over the African continent;

(4) Identify areas with significant GRACE-derived trends using various statistical techniques;

(5) Spatially correlate and compare the GRACE-derived and the TRMM-derived inter-annual trends;

(6) Investigate the nature (e.g., climatic and/or human pressures-related) and the controlling factors causing the variations in the GRACE-derived trends.

On the local scale (El Qaa Plain, Sinai, Egypt), land-based gravity, aeromagnetic data as well as GRACE data was used to identify the structural and hydrogeologic settings, the areal distribution and geometry, the approximate water volume, and the aquifer depletion rates in one of the Gulf of Suez rift-related basins, the El Qaa Plain basin.
CHAPTER II
METHODOLOGY

In this chapter, the main processing steps that were applied to GRACE and TRMM data during this study were discussed. The acquisition and processing parameters of the land-based gravity, and aeromagnetic data were also provided.

2.1 Processing of GRACE Data

I analyzed 108 gravity field solutions (RL04 unconstrained solutions) spanning the period April 2002 through August 2011 from the GRACE database provided by the University of Texas Center of Space Research (UTCSR) (available at: ftp://podaac.jpl.nasa.gov/allData/grace/L2/CSR). These solutions are represented in terms of fully normalized spherical harmonic decompositions up to degree \( l \) and order \( m \) of 60.

The time-variable component of the gravity field is obtained by removing the long-term mean \( (\overline{C}_{lm}(t), \overline{S}_{lm}(t)) \) of the Stokes coefficients from each monthly \( (C_{lm}(t), S_{lm}(t)) \) value to get the temporal variations in these coefficients \( ((\Delta \overline{C}_{lm}(t), \Delta \overline{S}_{lm}(t)) \). The reason for removing the mean field is that it is dominated by the static density distribution inside the solid Earth (Wahr et al., 1998).

After the temporal mean was removed, the correlated errors (long, linear, north to south oriented features) also called “stripes” were reduced by applying destriping methods developed by Swenson and Wahr (2006). The presence of stripes
implies spatial correlations in the gravity field coefficients at higher degree (short wavelength components). Swenson and Wahr (2006) show how, at a single higher \((m>8)\) order, the even (or odd) coefficients values at each degree tend to form smooth curves. Their filter, a moving window quadratic polynomial, was used to isolate and remove smoothly varying coefficients of like parity. In this study, a least square method was used to fit to a 4\(^{th}\) order polynomial for the odd degree and the even degree separately, then to subtract those polynomials from the original coefficients to leave the residuals \((\Delta \hat{C}_{lm}(t), \Delta \hat{\mathcal{C}}_{lm}(t))\).

At this point, I have a set of gravity coefficient anomalies for each month \((\Delta \hat{C}_{lm}(t), \Delta \hat{\mathcal{C}}_{lm}(t))\) that have been destriped and ready to compute mass anomalies. Several studies have used GRACE monthly solutions to make estimates of water storage variability, both on land and in the oceans (Chambers et al., 2004; Chen et al., 2005; Ramillien et al., 2005; Swenson and Milly, 2006; Tamisiea et al., 2005; Tapley et al., 2004b; Velicogna and Wahr, 2005; Wahr et al., 2004). I compute maps of water storage anomalies over the land \((\Delta \sigma(\theta,\phi,t))\) directly as:

\[
\Delta \sigma(\theta,\phi,t) = \frac{\rho_{\text{ave}}}{3} \sum_{i=0}^{60} \sum_{m=0}^{60} \mathcal{P}_{lm} W_i(\cos \theta) \left[ \frac{2l+1}{1+k_i} \right] \times (\Delta \hat{C}_{lm} \cos(m\phi) + \Delta \hat{\mathcal{C}}_{lm} \sin(m\phi)),
\]

\[
W_i = \exp \left[ -\frac{(lr/a_e)^2}{4\ln(2)} \right]
\]

Where \(\rho_{\text{ave}}\) is the average density of the Earth (5517 kg m\(^{-3}\)), \(a_e\) is the mean equatorial radius of the Earth, \(\theta\) is the geographic latitude, \(\phi\) is the longitude, \(\mathcal{P}_{lm}\) are
the fully-normalized Associated Legendre Polynomials of degree $l$ and order $m$, and $r$ is the Gaussian averaging radius, and $k_l$ are elastic/land Love numbers of degree $l$. The later was calculated via linear interpolation of the original data computed by Han and Wahr (1995).

The spatial averaging, or smoothing, of GRACE data is necessary to reduce the contribution of noisy short wavelength components of the gravity field solutions. In this study I generated $0.5^\circ \times 0.5^\circ$ equivalent water thickness grids using a Gaussian averaging radius of 250 km. Following these steps a number of derived GRACE by-products were generated including:

1. Standard deviation (SD) images were generated from the equivalent water thickness grids over periods of 2, 3, 4, 5, 6, 7, 8, 9, and 10 years;

2. Amplitude and phase of annual cycle images were generated from the equivalent water thickness grids. The amplitude of the annual cycle was generated by fitting a sinusoid model to each of the grid points with a period of 12 months.

3. Trend images were generated by fitting a time series at each grid point using an annual sine and cosine, semi-annual sine and cosine, mean, and linear trend;

4. Trend data were statistically analyzed using parametric techniques to identify areas that have statistically significant trends at 95% and at 65% levels of confidence, and areas where no, minimal, or insignificant trends are observed.
2.2 Processing of TRMM Data

The Tropical Rainfall Measuring Mission (TRMM) data (version: 3B43.006) that spans almost the same GRACE period from April 2002 through June 2011 were used in this study. TRMM is a joint mission between the National Space Development Agency (NASDA) of Japan and NASA of the United States, launched in 1997 as part of the Earth Observing System (EOS). TRMM provides global (50°N–50°S) data on rainfall using microwave and visible-infrared sensors. Instantaneous rainfall estimates are obtained every 3 hr with a 0.25° × 0.25° footprint and continuous coverage from 1998–present. The primary rainfall instruments of TRMM are the precipitation radar, TRMM microwave imager, and the visible/infrared scanner (Kummerow et al., 1998). The processing techniques applied for TRMM data include:

(1) Rescaling TRMM pixel size from 0.25° × 0.25° to 0.5° × 0.5° in order to match GRACE-derived water thickness grid size;

(2) Removal of the temporal mean from each grid point;

(3) Generation of the standard deviation (SD) images over periods of 2, 3, 4, 5, 6, 7, 8, 9, and 10 years;

(5) Generation of the amplitude and phase of annual cycle images;

(6) Generation of trend images as in the case of GRACE data;

(7) Statistically test the significance of TRMM-derived trend images as in the case of GRACE data.
2.3 Acquisition and Processing of Land-based Gravity Data

In the following sections I include the main acquisition and processing steps of land-based gravity data.

2.3.1 Acquisition of Land-based Gravity Data

A total of 381 ground gravity stations were collected throughout four field seasons (December 2003, May and December 2005, and September 2011) in the El Qaa Plain area. The 2003 and 2005 gravity data were collected by Dr. William Sauck and others. The collection of the 2011 data and the processing of the whole data (four seasons) were done by me. The Worden Master gravimeter with sensitivity of 0.01 mGal was used in the gravity data collection. Each surveying day started and ended with base station readings. Most of the gravity readings were taken along profiles trending N60E and were tied by NW-SE traverses and covered an 85 km x 20 km area. The station interval along a typical profile ranged from 0.5 to 1.0 km whereas the profile separation ranged from 1.0 to 5.0 km. Accurate horizontal (X and Y) and vertical (Z) positioning of the gravity stations was made using paired Magellan ProMark X and Trimble GPS systems.

2.3.2 Reduction of Land-based Gravity Data

Gravity data reduction started with the differential processing of the GPS data for accurate horizontal and vertical positioning of each gravity station. Following the accurate positioning, a conventional processing regime was applied to the raw gravity
reading including latitude, free-air, Bouguer, drift, terrain, and tide corrections and used to calculate the complete Bouguer anomaly.

2.3.2.1 Latitude Correction

Gravity is affected by the latitude. It is higher at the poles, because the flattening of the Earth makes the geoid closer to its center of mass. Therefore a correction, the latitude correction ($\Delta g_L$), is added to $g$ when moving towards the equator (Telford et al., 1990):

$$\Delta g_L / \Delta s = (1 / R_e) \Delta g / \Delta \phi = 0.811 \text{ mGal/km}$$

Where $\Delta g_L$ is the latitude correction, $\Delta s$ is the N-S horizontal distance, $\phi$ is the latitude, and $R_e$ is the radius of the Earth (=6368 km). Alternatively, this latitude effect is corrected for by applying the “normal” gravity expression, also called the theoretical sea level gravity ($g_t$), to the reduction process where we use the Geodetic Reference System 1967 (GRS67) equation:

$$g_t = 978031.846\left(0.005278895 \sin^2 \phi + 0.00002346 \sin^4 \phi\right) \text{ mGal}$$

2.3.2.2 Free-air Correction

The Free-air correction is used to correct for differences in elevation between ground-based stations and a uniform base level (datum), normally mean sea level (MSL). The correction value ($\Delta g_{fa}$) is added to the reading when the station is above the datum plane and vice-versa (Telford et al., 1990).

$$\Delta g_{fa} / \Delta R = 2g / R_e = 0.3086 \text{ mGal/m}$$
2.3.2.3 Bouguer Correction

The Bouguer correction corrects for the gravitational contribution introduced by the material located between the gravity station and the datum. In contrast to the free-air correction, the Bouguer correction ($\Delta g_B$) is subtracted when the station is located above the datum. Considering the station is located over a large horizontal slab of uniform density and thickness, the correction is calculated as follows (Telford et al., 1990):

$$\frac{\Delta g_B}{\Delta R} = \frac{2\pi \rho}{\gamma \rho} = 0.04192 \rho \quad \text{mGal/m}$$

Where $\rho$ is the density in gm/cm$^3$ and $\gamma$ is the universal gravitational constant ($6.672 \times 10^{-8}$ cm$^3$/g.sec$^2$).

2.3.2.4 Terrain Correction

In and near mountainous areas, as in the case of Sinai, the local variation of the gravity field is affected (decreased) by topographic effects and thus the gravimeter readings must be corrected. Terrain corrections used to be performed manually through the use of topographical maps that were divided in sectors. The gravity effect is calculated for each sector and the total correction is the sum of the individual sector contributions (Telford et al., 1990). However, readily available commercial software are now used to calculate with high accuracy the topography-related corrections if digital topography data sets are available. In this study, the terrain correction was
calculated in the Geosoft Oasis Montaj environment using a 30 m horizontal resolution digital elevation model (DEM). The Oasis Montaj algorithm adopts the terrain correction methods described by Nagy (1966) and Kan (1962).

2.3.2.5 Drift Corrections

The gravity meters change their null reading value gradually with time due to the unidirectional creep in the gravimeter spring (Telford et al., 1990). The base station readings were used to obtain the drift correction values. I made a plot of readings at the base station as a function of time, linearly interpolated between the points, and subtracted the extracted correction values from subsequent readings as a function of their time of acquisition.

2.3.2.6 Tide Corrections

Because the geophysical definition of gravity includes all accelerations acting on a gravimeter, the time-variant tidal forces must also be accounted for. The solid-earth and ocean tide effects on gravity are due to the direct attraction of the Moon and the Sun (Telford et al., 1990), as well as the attraction of the oceanic and earth bulges that they cause. Tide corrections were calculated adapting the Longman (1959) equations and using the latitude and longitude of the station, and the time of acquisition of data.
2.3.2.7 Complete Bouguer Gravity Anomaly

All of the above mentioned corrections were used to calculate the complete Bouguer gravity anomaly \( g_{CBA} \) (Telford et al., 1990):

\[
g_{CBA} = g_{obs} - \gamma + (\Delta g_{FA} - \Delta g_B + \Delta g_T)
\]

Where \( g_{obs} \) is the station reading, \( \Delta g_{FA} \) is the free-air correction, \( \Delta g_B \) is the Bouguer correction, \( \Delta g_T \) is the terrain correction and \( \gamma \) is the theoretical sea level gravity (also termed “normal gravity”).

2.4 Acquisition and Processing of Aeromagnetic Data

A total of 800 km of airborne magnetic data were acquired in the northern El Qaa Plain area. The aeromagnetic data was a part of an airborne survey that was conducted by the exploration division of the Egyptian Nuclear Material Authority in March and April of 1998. A Scintrex Cesium magnetometer was used in data acquisition with in-flight sensitivity of 0.001 nT, sampling at 0.1 second, and at an average terrain clearance of 120 m. Twenty seven NW-SE –oriented paths covering a 65 km x 15 km area with a flight path separation of 1.0 km were acquired in the El Qaa Plain area.

A conventional processing regime was applied to the raw total magnetic field data by the Egyptian Nuclear Material Authority including removing the diurnal variation effects, heading corrections, lag correction, removing of the International Geomagnetic Reference Field (IGRF), and leveling (Elsirfe et al., 1998).
Post-processing techniques such as forward 2D modeling, horizontal and vertical derivatives, and analytic signal were used to define geometries and to locate the source (edges) of both gravity and magnetic anomalies. The physical concept behind these techniques was described in detail in several studies (e.g., Nabighian 1972; Roest et al. 1992; Macleod et al. 1993).

2.5 Building of GIS Database

The spatial distribution of GRACE, TRMM, land-based gravity, and aeromagnetic data products was compared to other relevant co-registered geologic, topographic, and hydrologic data in a Web-based GIS environment (available at www.esrs.wmich.edu/webmap).

2.5.1 Web-based GIS Structure

In this study, I used a web-based GIS system similar to that developed by Becker et al., (2012). The system consists of GoogleMaps, Python scripts, and an ArcGIS Server, which hosts all the services. The overall design of the web-based GIS system can be broken up into three distinct parts: (1) data storage; (2) GIS web server and rendering tiers; and (3) a front-end interface viewed by a user. Users of our web-based GIS can access standard navigation (pan, move, zoom, etc.) available with the Google Maps interface. The system also includes geoprocessing custom tools such as: (1) a graph tool that is used to graph certain datasets and to create elevation profiles; (2) a transparency tool that makes a displayed layer semi-transparent; (3) an identify
tool that provides the value at a specified point for an investigated layer; (4) a measure tool that measures selected distances, areas or perimeters; and (5) a time series tool that is used to generate a time series from GRACE or TRMM data at any selected location. This tool also provides opportunities to fit an annual function for both GRACE and TRMM datasets.

2.5.2 Web-based GIS Datasets

The GIS database (Fig. 2) includes the following co-registered digital data products:

(1) GRACE monthly (equivalent water thickness grids), Standard Deviation (SD), trend, and amplitude and phase of annual cycle images;
(2) TRMM monthly, average annual, total annual, SD, trend, and amplitude and phase cycle images;
(3) Simple and complete Bouguer anomalies, horizontal (X and Y) and vertical (Z) derivatives, and gravity analytic signal maps;
(4) Total magnetic intensity, horizontal (X and Y) and vertical (Z) derivatives, and magnetic analytic signal maps;
(5) Digital elevation data (DEM) extracted from Shuttle Radar Topography Mission (SRTM) data products (pixel size: 1 km);
(6) Slope data extracted from DEM (pixel size: 1 km);
(7) Geologic maps for Africa (scale: 1: 5000000; Choubert and Faure-Muret, 1987);
(8) False-color Landsat Thematic mapper (TM) data (pixel size: 90 m);
(9) Moderate Resolution Imaging Spectroradiometer (MODIS) images (pixel size: 500 m);

(10) Normalized Difference Vegetation Index (NDVI) data extracted from Landsat TM data (pixel size: 90m);

(11) Stream networks and watershed boundaries extracted (TOPAZ; Garbrecht and Martz, 1995) from the SRTM dataset;

(12) Distribution of surface water bodies extracted from SRTM and Landsat TM data, and geologic maps.

Figure 2: Web-based GIS end-user interface. The website available at: http://www.esrs.wmich.edu/webmap/.
CHAPTER III

GRACE: A TOOL FOR MONITORING ELEMENTS OF HYDROLOGIC SYSTEMS ON REGIONAL SCALES ACROSS AFRICA

3.1 Introduction

The majority of studies utilizing GRACE data for hydrological research and applications (e.g., Chen et al., 2005; Rodell et al., 2004a; Wahr et al., 2004) target large watersheds (areas of 450,000–600,000 km²), because the accuracy of the recovered mass variations increases with increasing size of the monitored basin (Wahr et al., 2006). Gaussian filters ranging from 600 to 1200 km in radius were applied in studies on the central U.S. High Plain Aquifer, the Amazon Basin, the upper Zambezi Basin, and Mississippi River Basin (Chen et al., 2005; Rodell and Famiglietti, 2002; Rodell et al., 2004a; Swenson and Wahr, 2009; Syed et al., 2005; Winsemius et al., 2006).

The majority of the world’s watersheds are much smaller, and even for large watersheds one often needs to understand the partitioning of water on the sub-basin level. Smoothing techniques, such as the isotropic Gaussian technique, are commonly applied to reduce the contributions from short-wavelength components. The smaller the radius of the Gaussian smoothing filter, the lower the accuracy in the Earth’s mass variations (Klees et al., 2007; Wahr et al., 2006). GRACE gravity field solutions smoothed using a small-radius Gaussian filter often display long, north–south, linear features, commonly referred to as stripes, that are more pronounced near the Equator (Swenson and Wahr, 2006).
3.2 Objectives

In this study, I demonstrate that temporal mass variations from the GRACE data acquired over the African continent, when smoothed using a 250-km radius Gaussian function, are largely controlled by elements of the hydrologic cycle, and have not been obscured by noise as previously thought. Such findings could have significant implications for the possibility of using readily available GRACE data to examine temporal responses of a much larger suite of hydrologic systems (e.g., watersheds, lakes, rivers, marshes, etc.) and domains (e.g., source areas, lowlands) within watersheds and sub-basins worldwide. Findings are based on the nature of the spatial and temporal patterns of GRACE monthly solutions in relation to observations extracted from co-registered relevant data sets and products (e.g., precipitation, topography, and geology) as examined in a GIS environment.
Figure 3: Standard Deviation (SD) image derived from equivalent water thickness grids (0.5° × 0.5°) from GRACE monthly (cm) solutions for a period of (A) two years (2002 and 2003), (B) three years (2002 through 2004), (C) four years (2002 through 2005), (D) five years (2002 through 2006), (E) six years (2002, through 2007), (F) seven years (2002 through 2008), (G) eight years (2002 through 2009), (H) nine years (2002 through 2010), and (I) ten years (2002 through 2011).
3.3 Observations and Findings

All GRACE-derived mass fields were interpreted as reflecting changes in water storage, given (1) the slow rates of the mass variations in the underlying solid Earth, (2) absence of large earthquakes and glacial isostatic adjustment in the region, (3) the small to negligible contributions related to mass fluctuations from the adjacent ocean (Wahr et al., 1998) and the corrections applied to remove the time-variable oceanic gravity signal from raw GRACE measurements (Tapley et al., 2004a; Tapley et al., 2004b), and (4) the fact that errors in the atmospheric pressure corrections applied to GRACE data are likely to be negligible in this region (Velicogna et al., 2001).

3.3.1 GRACE SD Images

Examination of the SD images that were generated from monthly GRACE solutions over periods of 2, 3, 4, 5, 6, 7, 8, 9, and 10, years (Fig. 3 A though I) showed persistent patterns. The persistent spatial characteristics of the SD anomalies contrast with those of the surface mass anomalies derived from the individual monthly GRACE solutions. The latter showed pronounced anomalies that vary in location and magnitude and are largely concentrated in sub-Saharan and tropical Africa. One interpretation for the observed persistent nature of the anomalous areas on the SD images is that they represent areas that are largely controlled by inherent mass variations (signal) that are modulated, but not obscured, by noise.
Figures 4A and 4B show a three-dimensional representation of SD images of both GRACE and TRMM data over the examined time period. Using arbitrary threshold SD values, I classified the mass anomalies displayed in Figures 3I and 2A into three major groups: (1) areas of high mass variations (SD >10 cm); (2) areas with intermediate variations (10 cm > SD > 6 cm); and (3) areas with low to no variations (SD < 6 cm). Areas displaying intermediate to high mass variations on SD images are located in sub-Saharan Africa, whereas Saharan Africa displays negligible mass variations. The Sahel region, which separates Saharan from sub-Saharan Africa, shows intermediate to small mass variation. Similar spatial precipitation patterns were observed over Saharan and sub-Saharan Africa and the Sahel region, which suggests a causal effect (Fig. 4B). One should not expect a one-to-one correspondence between mass variation and precipitation patterns, given that precipitation could readily be redistributed as runoff, recharge, evaporation, and transpiration, all of which could affect the spatial and temporal distribution of the precipitated water and hence the location and magnitude of the SD anomalies.

All of the pronounced anomalous areas (SD >10 cm) were found within relatively large- to medium-sized basins (e.g., Congo River Basin [CRB]; area 3,712,739 km²; Niger Basin: area 2,144,785 km²; Nile Basin: area 3,086,409 km²). Next, I show that these anomalous areas originate in mountainous source areas that receive high precipitation; they extend downslope toward the mountain foothills and often continue along the main channels, wetlands, or lakes draining these areas.
Figure 4: Three-dimensional GRACE image products overlain on digital topography extracted from SRTM data. (A) Standard deviation (SD) image displayed in Figure 3I. (B) Tropical Rainfall Measuring Mission (TRMM) derived SD image.
3.3.2 Congo River Basin (CRB)

The CRB receives the highest amount of precipitation (Average Annual from TRMM [AATRMM]: 2,000 mm/yr) of all the major watersheds in Africa and experiences the strongest SD anomalies. The GRACE anomalies within the CRB (Fig. 5) originate from mountainous areas that experience high precipitation, namely the Albertine Rift range (Anomaly Location [AL]: C1), the Ironstone Plateau (AL: C2), the Adamawa Plateau (AL: C3), the Lunda Plateau (AL: C4), and Muchinga Mountains (AL: C5). To a large extent, the anomalies then follow the main tributaries that drain these highlands and ultimately extend into the Congo River (length: 4,700 km) through the Democratic Republic of the Congo (DRC).

Precipitation over the Albertine Rift range (AATRMM: 2,000 mm/yr), Muchinga Mountains (AATRMM: 1,010 mm/yr), and the Lunda Plateau (AATRMM: 350 mm/yr) is channeled through the Lukuga (length: 320 km; AL: C6), Lualaba (length: 1,800 km; AL: C7), Lomami (length: 1,500 km; AL: C8), Kwango (length: 1,100 km; AL:C9), Kwilu (length: 600 km; AL: C10), and Kasai (length: 1,800 km; AL: C11) Rivers and the tributaries of the Kasai River, the Lulua (length: 420 km; AL: C12) and Sankuru (length: 1,230 km; AL: C13) Rivers. The Ironstone Plateau (Average Height [AH]: 600 m a.m.s.l) precipitation (AATRMM: 1,500 mm/yr) feeds the Ubangi River (length: 1,100 km; AL: C14) and its tributaries, whereas the Adamawa Plateau (AH: 1,000 m a.m.s.l) precipitation (AATRMM: 1,500 mm/yr) is channeled through the Sangha affluent (length: 850 km; AL: C15), a north-south-trending tributary of the Congo River.
Figure 5: Standard deviation (SD) image derived from equivalent water thickness grids (0.5° × 0.5°) from GRACE monthly solutions acquired for the Congo River Basin (CRB). Also shown are locations of major rivers (blue lines), highlands (dashed black lines), and basins (purple lines). The inset shows the CRB outline in red color.
3.3.3 Niger River Basin

The GRACE anomalies within the Niger Basin (Fig. 6) originate from the Fouta Djallon range (AH: 1,100 m a.m.s.l), which receives the highest amount of precipitation (AATRMM: >2,000 mm/yr) in Guinea, and the Nimba Range (highest point: 1,752 m a.m.s.l) along the borders of Guinea and the Côte d'Ivoire (AL: G1), which receives AATRMM of 3,000 mm/yr. The anomalies then extend northeast along the Niger River (AL: G2). The Niger River (length: 4,200 km) anomaly decreases with distance from the source area, but it is emphasized again (AL: G3) at its junction with the Sokoto River, which channels precipitation from the Jos Plateau (AH: 1,280 m a.m.s.l; AATRMM of 1,200 mm/yr). Other anomalous areas originate at the highlands of Cameroon (Adamawa Plateau), then follow the Benue River (AL: G4), the major tributary of the Niger River (length: 1,400 km), toward its junction with the Niger River, and extend over the massive delta where the two rivers discharge into the Atlantic Ocean (AL: G3) in the Gulf of Guinea.
Figure 6: Standard deviation (SD) image derived from equivalent water thickness grids (0.5° × 0.5°) from GRACE monthly solutions acquired for the Niger Basin. Also shown are locations of major rivers (blue lines), highlands (dashed black lines), and basins (purple lines). The inset shows the Niger Basin outline in red color.
3.3.4 Nile Basin

The GRACE anomalies within the Nile Basin (Fig. 7) originate from mountainous areas along the western margins of the Ethiopian highlands (AL: N1 and N2), the largest continuous area of its altitude (AH: >1,500 m a.m.s.l) in Africa and the northern parts of the Ironstone Plateau (AL: N3). Precipitation over the northwest part of the Ethiopian highlands (AATRMM: 1,400 mm/yr) drains into Lake Tana (elevation: 1840 m; area: 3,156 km$^2$), where the Blue Nile (length: 1,450 km) originates. The SD anomalies follow the Blue Nile from the highlands to its intersection with the White Nile (AL: N5), where the anomaly is emphasized. Anomalous areas on the SD images are also observed over the sub-basins that ultimately feed the White Nile; these anomalous areas start in the source highland areas, pass by lakes and wetlands, and follow the main tributaries up to the point where the White Nile emerges. Precipitation over the northern parts of the Albertine Rift and the Kenyan highlands (AATRMM: 2,000 mm/yr) are channeled (e.g., by the Kagera River) to Lake Victoria (elevation: 1,135 m a.m.s.l), Africa’s largest lake (area: 68,800 km$^2$) (AL: N4). The Victoria Nile exits Lake Victoria to Lake Kyoga (area: 1,720 km$^2$) and Lake Albert (area: 5,300 km$^2$), and then continues its journey toward the Sudan as the Albert Nile, where its name changes to Bahr al Jabal (length: 716 km) and it joins with Bahr al Jazal to form the White Nile. The Bahr al Ghazal and Sobat (length: 320 km) rivers are the two most important White Nile tributaries.
Figure 7: Standard deviation (SD) image derived from equivalent water thickness grids (0.5° × 0.5°) from GRACE monthly solutions acquired for the Nile Basin. Also shown are locations of major rivers (blue lines), highlands (dashed black lines), and basins (purple lines). The inset shows the Nile Basin outline in red color.
3.3.5 GRACE Time Series Analysis

There is a general correlation between precipitation patterns and SD anomaly patterns (Figs. 4A and 4B) and the spatial correlation of anomalous areas on the SD image with (1) mountainous source areas receiving high precipitation, and (2) elements of drainage systems, including rivers, lakes, and wetlands that channel and/or collect precipitation from the source areas. This supports the suggestion that the observed GRACE mass variations are related to elements of the hydrological cycle (e.g., infiltration, recharge, and/or surface runoff, and/or groundwater flow) observed at the sub-basin scales examined here.

This suggestion is supported by the seasonal patterns observed in the time-series analysis for the GRACE monthly gravity solutions. Figure 8 displays examples of this mass variability observed in monthly data (from April 2002 to August 2011) for anomaly locations C3 and C5 (highlands in CRB; Fig. 5), G1 (highlands in Niger Basin; Fig. 6), and N1 (highlands in Nile Basin; Fig. 7). Comparisons to precipitation time series for the same areas indicate that the GRACE solutions and precipitation data display similar seasonal patterns, but GRACE seasonal variations lag a month or two behind precipitation. In the Niger Basin (point G1; Fig. 8A), I find that the highest rainfall occurs between July and August and the largest increase in mass is between September and October. Similarly, maximum rainfall in the Nile Basin (point: N1; Fig. 8B) is between July and August, a month or two ahead of GRACE SD peaks (September to October). In the northern part of the CRB (point: C3; Fig. 8C), GRACE SD peaks lag a month or two behind rainfall (August to September);
they lag a month or two behind rainfall (December, January, and February) in the southern parts of the CRB (point: C5; Fig. 8D) as well. Findings are supported by reported observations for the timing of peak precipitation and flow in these three basins (Chishugi and Alemaw, 2009). For example, the peak river flows for the Ubangi River (Fig. 5) in northern CRB are from October to November (Orange et al., 1997) and those for the Luapula River (Fig. 5) in the southern CRB are in March and April (Munzimi, 2008).

One explanation is that with the onset of the precipitation period, which occurs mostly over the mountainous areas, a good fraction of this water is captured through initial losses, increasing soil moisture and creating local ponds and sinks, thus increasing the accumulated water and mass. With continued precipitation and mass accumulation, GRACE response will continue to rise and perhaps peak at a point where the soil moisture content approaches saturation. When precipitation starts to decline, mass accumulation gradually diminishes to the point at which water received in the soil profile from precipitation is less than that lost to evapotranspiration and to outflows (runoff, overland flow, interflow, and groundwater flow). This marks a halt in additional increases in water thicknesses and the onset of the effects of mass deficiency. As the rainy season comes to an end, precipitation tapers off and factors such as evaporation, transpiration, and runoff progressively decrease accumulated water, reducing the GRACE response until it bottoms out.
Figure 8: Time series plots for GRACE and TRMM data and their three-point moving averages for anomaly locations shown in Figures 5, 6, 7. (A) Location G1; Fig. 6. (B) Location N1; Fig. 7. (C) Location C3; Fig. 5. (D) Location C5; Fig. 5.
If this conceptual model is true, one would expect the amplitude of annual cycle from GRACE data to decrease with distance from the source areas because (1) the source areas by nature receive the highest amounts of precipitation, and (2) only a portion of this precipitation moves toward the lowlands downstream as runoff, overland flow, interflow, and groundwater flow in shallow aquifers. One would also expect a progressive shift in phase of annual cycle from GRACE data with distance from the source area because of the time it takes for runoff, overland flow, interflow, and/or groundwater flow in shallow aquifers to move the water from the highlands to the lowlands. The shift here refers to the shift of the peak and/or trough observed in a single annual cycle in the monthly GRACE solutions.

3.3.6 Amplitude and Phase of the Annual Cycle

Inspection of the amplitude of the annual cycle image (years 2003 to 2010) (Fig. 9A) shows that the amplitude of the annual cycle over mountainous areas that receive high precipitation (e.g., areas labeled N, C, C1, G; Fig. 9A) is high and declines downstream with distance from the highlands (e.g., traverse N-N', C-C', C1-C1', G-G'; Fig. 9A). The phase of the annual cycle image of monthly TRMM precipitation data and for monthly GRACE data for the same year are displayed in Figures 9B and 9C, respectively; on these images, peak precipitation or mass are assigned values ranging from 1 (January) to 12 (December). Over the mountainous source areas in the Nile, Congo (northern part), and Niger basins, areas labeled N, C, and G in Figure 9B, the peak precipitation occurs largely in the months of July,
August, and September, respectively, and is monsoonal in origin (Chishugi and Alemaw, 2009; Kebede et al., 2006; Mbagwu et al., 2002; Olivry, 2002; Semazzi and Song., 2001; Sutcliffe and Parks, 1999). Peak monthly GRACE values progressively shift to October, November, and December, respectively, with distance from the mountainous areas. Progressive shift in phase with distance from the mountainous source areas is observed; the steeper the source areas, the smaller the distance over which mass variations are observed (e.g., traverse N-N', C-C' C1-C1'; Fig. 9C). That is to be expected, because the steeper the gradient is, the faster the water will move out of it.

3.4 Conclusions

Results show that temporal and spatially smoothed (250 km; Gaussian) mass variations are largely controlled by elements of the hydrologic cycle such as runoff, infiltration, and groundwater flow, and that these mass variations are probably modulated, but not obscured by noise as previously thought. If true, our findings suggest that it is possible to use GRACE to investigate temporal responses of a much larger suite of (smaller) hydrologic systems (watersheds, lakes, rivers, marshes, etc.) and domains (e.g., source areas, lowlands) within watersheds and sub-basins worldwide.
Figure 9: 3D GRACE and TRMM products overlain on digital topography extracted from SRTM data. (A) Amplitude of annual cycle derived from monthly GRACE solutions. (B) Phase of annual cycle for the TRMM data. (C) Phase of annual cycle for GRACE monthly solutions. Also shown is the distribution of drainage networks, selected locations on mountainous areas (G, C, C1, N) and traverses (G-G, C-C, C1-C1, N-N) along rivers channeling precipitation from these areas.
CHAPTER IV

GRACE: A TOOL FOR MONITORING CLIMATE AND MAN-MADE INDUCED VARIATIONS IN WATER AVAILABILITY ACROSS AFRICA

4.1 Introduction

It is common practice for researchers engaged in research related to climate change to examine the temporal variations in relevant climatic parameters (e.g., temperature, precipitation, evaporation, transpiration, soil moisture, humidity, etc.) and to extract and examine drought indices reproduced from one or more such parameters. For prediction purposes, the outputs of climatic models remain the most useful tool in this regard. These tools could have obvious societal benefits, informing the public across the globe about observed climatic changes, as well as their nature and severity, and forecasting such changes. Whether areas are getting wetter or drier, warmer or cooler, or more arid or more humid has serious societal implications; such conditions bear on the availability of water for humans and sustainability of ecosystems, especially in the case of arid and semi-arid parts of the world.

4.2 Drought Indices: Proxies for Water Availability

Drought indices are commonly used as proxies for monitoring water availability in an area (Dai, 2011a, b). Indices that measure recurring extreme climatic events on land that are characterized by below normal precipitation over periods of months to years are commonly used to monitor and quantify droughts. Drought indices used to define departures from normal conditions include: (1) meteorological
drought indices that identify periods with below normal precipitation and above 
normal temperatures (Dai, 2011b), (2) agricultural drought indices define periods with 
dry soils resulting from below average precipitation, intense but less frequent rain 
events, or above normal evaporation, all of which lead to reduced crop production and 
plant growth (Mannocchi et al., 2004); and (3) hydrological drought indices define 
periods when river stream flow and water storages in aquifers, lakes, or reservoirs fall 
below long-term mean levels (Rathore, 2004; Tallaksen and Van Lanen, 2004). In 
quantifying meteorological drought indices, precipitation is the primary variable, 
surface air temperature the secondary factor; soil moisture content and stream flow 
data are the primary variables used in estimating the agricultural and hydrological 
drought indices respectively.

There are uncertainties associated with the use of drought indices as proxies 
for water availability in an area. Many of these indices exclude significant controlling 
factor(s), do not work well in specific settings, and often require complex 
computations. Evaporation is excluded in the case of the Standardized Precipitation 
Index (SPI) and Rainfall Deciles (RD) (Edwards and McKee, 1997; McKee et al., 
1993) and antecedent conditions in the case of the Palmer Moisture Anomaly Index 
(Z-index) (Karl, 1986; Palmer, 1965). The Palmer Drought Severity Index (PDSI) 
does not work well in mountainous and snow covered settings (Alley, 1984), yields 
inconsistent values across diverse climatological regions (Alley, 1984; Guttman, 
1992; Heddinghaus and Sabol, 1991; Karl, 1983, 1986), and requires long (≥ 50 yr) 
calibration periods (Karl, 1986) and substantial meteorological data. The latter limits
its application in areas lacking adequate observational networks (Smakhtin and Hughes, 2004). Additional uncertainties are introduced by the models used in computing model-dependent-indices. The Computed Soil Moisture (CSM) index is an output of a land surface model that is forced by observed precipitation, temperature and other additional atmospheric forcing factors (Huang et al., 1996; Qian et al., 2006; Wang et al., 2009). Similarly, the Surface Water Supply Index (SWSI) for river basins is driven by precipitation, snowpack, and stream flow, yet is also affected by basin-dependent formulations (Doesken et al., 1991; Shafer and Dezman, 1982).

Aside from these uncertainties, none of these indices measure the variability in the terrestrial water storage (TWS) in an area, a term that refers to the total vertically-integrated water content in an area regardless of the reservoir (surface water, groundwater, soil moisture and permafrost, snow and ice, and wet biomass), in which it resides. The variability in the TWS is spatially-dependent, dominated by snow and ice in polar and alpine regions, by soil moisture in mid-latitudes, and by surface water in wet, tropical regions (Bates et al., 2007; Rodell and Famiglietti, 2001). The TWS, I believe, is the most adequate indicator for monitoring the temporal changes in water availability, spatially and temporally, in varied hydrological and climatological settings.

GRACE has some shortcomings when it comes to measuring TWS. The majority of studies utilizing GRACE data for hydrological research and applications target large watersheds (areas of 450,000–6,000,000 km²), because the accuracy of the recovered mass variations increases with increasing size of the monitored basin.
Gaussian filters ranging from 600 to 1200 km in radius were applied in studies on the central U.S. High Plain Aquifer, the Amazon Basin, the upper Zambezi Basin, and Mississippi River Basin (Chen et al., 2005; Rodell and Famiglietti, 2001; Rodell et al., 2004a; Syed et al., 2005; Winsemius et al., 2006).

4.3 Objectives

I showed that, when smoothed using a 250-km radius Gaussian function, temporal mass variations from the GRACE data acquired over northern and central Africa and as far as 10° south of the Equator are largely controlled by elements of the hydrologic cycle (e.g., runoff and recharge), and have not been obscured by noise as previously thought (Ahmed et al., 2011).

In this study, I capitalize on such findings and use GRACE data to examine temporal responses of a much larger suite of hydrologic systems (e.g., watersheds, lakes, rivers, marshes, etc.) and domains (e.g., source areas, lowlands, etc.) within watersheds, sub-basins, and aquifers across Africa. Specifically I utilize GRACE dataset in conjunction with other readily available climatic and relevant remote sensing data sets for monitoring the spatial and temporal long term inter-annual trends in TWS over a time period of 10 years and investigate the nature (e.g., climatic and/or human pressures-related) and the controlling factors causing these variations.
4.4 Observations and Findings

Examination of trends in the GRACE-derived estimates of TWS across Africa (Fig. 10A) shows that large sectors of the African continent are undergoing significant TWS variations, positive variations indicative of increasing TWS, or negative trends in areas experiencing decreasing TWS. Three regions are shown: areas experiencing trends that are statistically significant at 95%, and 65% levels of confidence, and areas where no, minimal or insignificant trends are observed. These are the remaining areas in Africa. One would expect many of these TWS trends to be controlled by either natural causes such as changes in the patterns and amounts of precipitation (Fig. 10B) (scenario I) or by man-made structures and practices that could be as influential as the natural causes (scenario II).

4.5 Mass Variations Scenarios

4.5.1 Scenario I: Mass Variations Induced by Natural Climatic Changes

In a simple system, one would expect areas that are witnessing an increase in precipitation rates and/or a decrease in temperatures with time to experience an increase in TWS due to increased accumulation of water in various reservoirs, and decreased evaporation, and vice versa. One should not expect a one-to-one correspondence between mass variation and precipitation patterns, given that precipitation could readily be redistributed as runoff, recharge, evaporation, and transpiration, all of which could affect the spatial and temporal distribution of the precipitated water and hence the location and magnitude of the mass anomalies. In
addition, the spatial resolution of GRACE data (Gaussian radius: 250 km) is substantially less than that for TRMM (0.25° x 0.25°). Thus, precipitation anomalies appear as well-defined domains on TRMM trend images, whereas TWS anomalies appear as diffuse zones on GRACE trend images.

![Figure 10: Linear trend images generated over the African continent: (A) Monthly (04/2002 to 08/2011) GRACE data. (B) Monthly (04/2002 to 06/2011) TRMM data. Also shown are the statistically significant areas at 95 % (dashed back polygon), and 65 % (black dots) level of confidence. The red polygons with labels (A-H) represent areas of similarities/differences between the two images.](image)

Most of the precipitation over Africa occurs over areas affected by the Intertropical Convergence Zone (ITCZ), a zone (latitude: 23.4° N to 23.4°S) characterized by low pressure, trade wind convergence and confluence, increased cloudiness, and high surface temperature, rainfall (McGregor and Nieuwolt, 1998;
Nicholson, 2009). In sub-Saharan Africa, the source of precipitation is largely controlled by Atlantic (Fontaine et al., 1998; Giannini et al., 2003; Janicot et al., 1998; Reason and Rouault, 2006; Trzaska et al., 1996; Vizy and Cook, 2002) and Indian monsoons (Conway, 2009; O’hare et al., 2005; Saji et al., 1999), whereas precipitation over northern Africa, is largely controlled by westerlies that have Atlantic and Mediterranean sources (Dunkeloh and Jacobit, 2003; Hurrell, 1995; Hurrell and Van Loon, 1997; Lamb and Peppler, 1987; Rodo et al., 1997).

Global warming influences sea and land surface temperatures, which in turn affect precipitation rates and patterns. An increase in temperature on land increases evaporation, dries land surfaces, intensifies droughts, and prolongs drought duration; over oceans, the water-holding capacity of air increases, storms intensify, and precipitation over the coastal plains and mountainous source areas along storm trajectories increase (Trenberth, 2011; Trenberth and Shea, 2005). One would expect that the increased precipitation would give rise to positive TWS over the coastal plains, and over the source areas, stream networks and outlets of proximal watersheds. This is observed in areas outlined by polygons A, B, and C on Fig 10. In these areas a positive trend in precipitation and TWS was observed. The positive precipitation trend is related to the intensification of Atlantic monsoons (Mohino et al., 2011; Rodriguez-Fonseca et al., 2011) reaching as far inland as central Chad (Levin et al., 2009; Nicholson, 1996). The contiguous patterns of increased precipitation and TWS throughout areas defined by polygon “A” supports the
suggestion (Levin et al., 2009; Nicholson, 1996) that the intensified Atlantic monsoons impacted areas as far inland as central Chad (Fig. 10).

The observed positive trends in TWS over the coastal plains of Nigeria, Benin, Togo, and Ghana (C1; 18.9 mm/yr; Fig. 11A), Gabon, Congo, Democratic Republic of Congo, and Angola (C2; 13.6 mm/yr; Fig. 11A), and Namibia, and Angola (C3; 18.2 mm/yr; Fig. 11A) are apparently related to increasing rates of precipitation (C1: 4.9 mm/yr; C2: 6.9 mm/yr; C3: 6.1 mm/yr; Fig. 11B).

Positive trends in GRACE TWS were also observed over the mountainous source areas and across the basins(s) they feed and their stream networks. The increase in TWS over the Bié Plateau (average height [AH] 1,780 m a.m.s.l), the source area for the Zambezi and Okavango Basins (S1; 36 mm/yr; Fig. 12A), and the Téna Kourou mountain range (AH 749 m a.m.s.l) the source area for the Niger Basin (S2; 11.6 mm/yr; Fig. 12A) are apparently largely related to increasing rates of precipitation (Zambezi: 3.1 mm/yr; Okavango: 5.6 mm/yr; and Niger: 3.5 mm/yr; Fig. 12B). The reported satellite-based positive trends in precipitation and TWS are supported by field observations including: (1) a rise in water levels in the Zambezi (up to 3 m), and Niger (up to 1 m) basin streams (Cretaux et al., 2011), and (2) floods in the Zambezi Basin, which are typically cyclical in nature, became unpredictable and damaging (Kirchhoff and Bulkley, 2008) and flooded areas in the Okavango Delta (Fig. 12) within the Okavango Basin, increased in size (from $5 \times 10^3$ km$^2$ to 12 $\times 10^3$ km$^2$) (Wolski et al., 2006).
The warming of the central Indian Ocean has the opposite effect on eastern Africa. It disrupts the onshore moisture transport, decreases continental precipitation and causes droughts (Funk et al., 2008). This could explain the observed general decrease in precipitation and TWS along the eastern and southern margins of the African continent. Decreased precipitation over eastern Africa has apparently given rise to the negative TWS trends over the mountainous source areas. The Matopa Hills (AH 1,021 m a.m.s.l.) in Mozambique, the source areas for the Limpopo and Mozambique north-east coast basins witnessed decreased precipitation rates (S3: -3.6 mm/yr, Fig. 13B) and TWS (S3: -7.5 mm/yr; Fig. 13A). Similarly, the Aberdare

Figure 11: Linear trend images generated over the African continent: (A) Monthly (04/2002 to 08/2011) GRACE data. (B) Monthly (04/2002 to 06/2011) TRMM data. The figure shows that the increase in the TWS over the coastal plains (C1, C2, and C3) is probably related to increase in precipitation rates along the same areas.
Range (AH: 3,500 m a.m.s.l) in Kenya, the source area for the Shebelli and Juba Basin, and the East Central Coast Basin experienced similar negative trends in precipitation (S4: -5.4 mm/yr; Fig. 13B) and TWS (S4: -11.4 mm/yr; Fig. 13A) rates.

Figure 12: Linear trend images generated over the African continent: (A) Monthly (04/2002 to 08/2011) GRACE data. (B) Monthly (04/2002 to 06/2011) TRMM data. The figure shows that the increase in the TWS over the source areas is related to increase in precipitation rates over the same areas.

The reported satellite-based negative trends in precipitation and TWS are supported by field observations: (1) decreasing rates of precipitation (S3: 2-10% [Bambaige, 2007]; S4: up to 100 mm (Funk, 2010)) and increasing temperatures (S3: 1.8-3.2°C [1960-2005]; S4: 0.7-1.3°C [1960-2009]) in Mozambique and Kenya over the past decade (Bambaige, 2007; Funk, 2010; Osbahr et al., 2008), (2) Mozambique witnessed a series of droughts (1981-1984, 1991-1992, 1994-1995, 2002-2003, 2005-
2006, and 2008-2009) (Sacramento et al., 2010), (3) Three droughts were reported from Kenya’s drylands during the last ten years in 2000, 2004-2005 and 2008-2009 (Ndathi et al., 2011).

The observed negative TWS trend over southern Africa (Polygon L1; -3.9 mm/yr; Fig. 14A) is probably caused by rising temperatures. The mean annual temperature averaged over South Africa, increased by 0.6°C between 1960 and 2006, at an average rate of 0.14°C/decade and the daily temperature observations show significantly increasing trends in daily temperature extremes (McSweeney et al., 2010).

Figure 13: Linear trend images generated over the African continent: (A) Monthly (04/2002 to 08/2011) GRACE data. (B) Monthly (04/2002 to 06/2011) TRMM data. The figure shows that the decrease in the TWS over the source areas of the Eastern Africa watersheds is related to decrease in precipitation rates.
The positive TWS trend over the Atlas Mountain (L2; 3.2 mm/yr; Fig. 14A) is related to increasing rates of precipitation (1.6 mm/yr; Fig. 14B) over the past decade. The increased precipitation was attributed to the higher frequency of negative phases of the North Atlantic Oscillation (NAO), where these phases are associated with higher precipitation over the Atlas Mountain (Garci´a Herrera et al., 2001; Lamb et al., 1997; López-Moreno et al., 2011; Rodo et al., 1997; Ulbrich et al., 1999; Ward et al., 1999; Zorita et al., 1992).

Figure 14: Linear trend images generated over the African continent: (A) Monthly (04/2002 to 08/2011) GRACE data. (B) Monthly (04/2002 to 06/2011) TRMM data. The figure shows that the increase (decrease) in the TWS over the Atlas Mountains (South Africa) is related to increase (decrease) in precipitation rates.
4.5.2 Scenario II: Mass Variations Induced by Man-made Interventions

In some areas, anthropogenic ongoing practices and construction could exemplify the impacts of mass variations related to climate change. A few examples are cited. I attribute the positive (L3; 24.9 mm/yr; Fig. 15) TWS trend over the Lake Volta reservoir to increased precipitation rates (3.7 mm/yr; Fig. 15) and ponding of water behind the Akosombo Dam (capacity: 148 km$^3$) that flooded part of the Volta River Basin, and created Lake Volta, one of the world's largest man-made reservoirs (area: 8,502 km$^2$). The increase in precipitation was associated with an increase of up to 10 m in lake levels over the past decade (Cretaux et al., 2011).

Figure 15: Linear trend images generated over the African continent: (A) Monthly (04/2002 to 08/2011) GRACE data. (B) Monthly (04/2002 to 06/2011) TRMM data. The figure shows that the increase in TWS over Lake Volta is related to the increase in precipitation rates as well as the construction of the Akosombo Dam.
Inspection of Figure 16A shows a general decrease in TWS across large sections of the Sahara region (area: $9.4 \times 10^6$ km$^2$) that I attribute to one or more of the following factors: decrease in precipitation, rise in temperature (Hulme et al., 2001), and increase in groundwater extraction from the underlying fossil aquifers. The North-Western Sahara Aquifer System (NWSAS; area: $1 \times 10^6$ km$^2$), an aquifer that extends across Algeria, Tunisia and Libya experienced a negative TWS trend (-2.6 mm/yr; Fig. 16A), the Nubian Sandstone Aquifer System (NSAS; area: $2.2 \times 10^9$ km$^2$) in Egypt experienced a decline of -3.5 mm/yr (Fig. 16A). Extraction rates from the NWSAS are on the rise (year 1970: $0.60 \times 10^9$ m$^3$/yr; 2000: $2.5 \times 10^9$ m$^3$/yr) (Al-Gamal, 2010; OSS, 2008) and currently exceed modern recharge estimated at $1 \times 10^9$ m$^3$/yr (Besbes et al., 2002; OSS, 2008). Similarly, extraction from the Nubian aquifer in Egypt is on the rise. During the last 40 years the groundwater table declined by over 60 m in the oases, and all of free-flowing wells and springs have been replaced by deep wells (Bakhbakhi, 2006) and is not compensated for, by aquifer recharge (0.8 mm/yr) in Sudan and northward groundwater flow towards Egypt (Sultan et al., 2012).

Man-made effects can reverse the natural climatic effects. For example, building dams will impound surface water in artificial lakes upstream from the dams, induce infiltration from the lake, and increase recharge from the lakes to groundwater. One such example, is the River Nile Basin, specifically in areas downstream from the source areas of the White and Blue Nile (3.9 mm/yr; Fig.17A). Examination of Figure 17B, shows a decrease in precipitation rate (-2.2 mm/yr) that is not matched by the
GRACE trend data. To the contrary, an increase in TWS trends is observed. I attribute the latter to the construction of a number of dams throughout the GRACE life time (March 2002 till present).

Examples (green triangles; Fig. 17) include: (a) Merowe High Dam (D1; location [LOC]: Sudan; construction period [CP]: 2004-2009; reservoir capacity [RC]: $12.5 \times 10^9$ m$^3$), (b) Tekezé Dam (D2; LOC: Ethiopia; CP: 1999-2009; RC: $31 \times 10^9$ m$^3$), (c) Amerti-Neshi Dam (D3; LOC: Ethiopia; CP: 2007-2010; RC: $0.5 \times 10^9$ m$^3$), (d) Beles Dam (D4; LOC: Ethiopia; CP: 2005-2010), (e) Gilgel Gibe I Dam (D5; LOC: Ethiopia; CP: 1986-2004; reservoir area [RA]: 63 km$^2$), and (f) Gilgel

Figure 16: Linear trend images generated over the African continent: (A) Monthly (04/2002 to 08/2011) GRACE data. (B) Monthly (04/2002 to 06/2011) TRMM data. The figure shows that the negative TWS trend over the Sahara aquifers is related to the decrease in precipitation and over extraction rates.
Gibe II Dam (D6; LOC: Ethiopia; CP: 2004-2009). The observed TWS trend is expected to continue given the dams currently in the planning phase or under construction. Examples (red triangles; Fig 17) include: (a) Kajbar dam (D8; LOC: Sudan, construction began [CB]: 2010; RA: 110 Km$^2$), (b) Dal Dam (D9; LOC: Sudan; CB: 2009), (c) Rumela and Burdana dams (D10; LOC: eastern Sudan, CB: 2010; RC: 3 x 10$^9$ m$^3$), (d) Shereik Dam (D11; LOC: Sudan, CB: 2010), (e) Baro Dams (D12; LOC: Ethiopia), (f) Gilgel Gibe III Dam (D13; LOC: Ethiopia; CB: 2007; RA: 200 km2), (g) Karadobi Dam (D7; LOC: Ethiopia; RC: 32.5x10$^9$ m$^3$), and (h) Grand Ethiopian Renaissance (Millennium) Dam(D14; LOC: Ethiopia; CB: 2011; RC: 63 x 10$^9$ m$^3$).

Figure 17: Linear trend images generated over the African continent: (A) Monthly GRACE data. (B) Monthly TRMM data. Dams name: the Merowe High Dam (D1), Tekezé (D2), Amerti-Neshi (D3), Beles (D4), Gilgel Gibe I (D5), Gilgel Gibe II (D6), Karadobi (D7), Kajbar (D8), Dal (D9), Rumela and Burdana (D10), Shereik (D11), Baro (D12), Gilgel Gibe III (D13), and Melennium (D14).
In central Africa, specifically over the Congo Basin, there has been an increase in precipitation rates (14.5 mm/yr; Fig. 18B) over the investigation period. Such a trend could have started at a much earlier date as heavy rainfall events were reported to have increased in frequency in the period 1931 and 1990 over Angola, Mozambique, Malawi and Zambia (Boko et al., 2007; Sivakumar et al., 2005). The increase in precipitation I report is not reflected in GRACE TWS trends. To the contrary, a decreasing trend (-16.4 mm/yr; Fig. 18A) was observed. I attribute this discrepancy between the precipitation and GRACE TWS trends to the increase in the deforestation and the forest degradation rates. Deforestation promotes runoff at the expense of infiltration and recharge, giving rise to an overall decrease in GRACE TWS. This suggestion is supported by the spatial correlation of the distribution of the heavily deforested areas and those of decreasing TWS. The net annual deforestation rates for central Africa counties for the period 2000-2005 are: Cameroon: 0.03%, Congo: 0.07%, Central Africa Republic: 0.06%, Democratic Republic of Congo: 0.22%, and Congo River Basin: 0.17% (de Wasseige et al., 2010; Duveiller et al., 2008; Ernst et al., 2010).

Similarly, the Miombo woodlands in southern Tanzania show positive precipitation trends (6.4 mm/yr; Fig. 18B), yet experience a negative TWS trend (L4; -8.2 mm/yr; Fig. 18A). Again I attribute these contrasting trends to deforestation (1990-2000: 1%; 2000-2010: 1.1 %) (FAO, 2011).
4.6 Conclusions

Results indicate that large sectors of the African continent are undergoing significant TWS variations (+36 mm/yr to -16 mm/yr), where observed positive and negative trends are indicative of increasing and decreasing mass, respectively. The fact that these trends are statistically significant over large sectors of Africa (20% of Africa at 95% level of confidence; 65% at 65% level of confidence) supports our earlier findings (Ahmed et al., 2011) that smoothed (250 km; Gaussian) GRACE mass variations are probably modulated, but not obscured by noise as previously
thought. Hence, it is possible to use GRACE to investigate temporal responses of a large suite of (smaller) hydrologic systems (watersheds, lakes, rivers, marshes, etc.) and domains (e.g., source areas, lowlands) within watersheds and sub-basins across Africa and world-wide.

GRACE is now providing, for the first time, unique opportunities to examine temporal and spatial variations in TWS on a global scale with consistent observational parameters. I have shown that the observed temporal and spatial trends in TWS and annual precipitation over the past decade in central, eastern, western, and southern Africa are consistent with recent interpretations that call on: (1) warming in tropical regions of the Atlantic Ocean that intensifies Atlantic monsoons and increases precipitation over western and central Africa’s coastal plains, proximal mountainous source areas, and inland as far as central Chad; (2) warming in the central Indian Ocean that disrupts onshore moisture transport decreases continental precipitation, and causes droughts over eastern and southern Africa; and (3) more frequent recurrence of the negative NAO index values that bring increased precipitation over northwest Africa. Not only does the GRACE TWS analysis reflect the impacts of ongoing natural climatic changes, our analysis has also shown that locally, man-made interventions (e.g., deforestation, dam construction) could amplify, modulate, and in some cases reverse the impacts of ongoing climatic changes when it comes to the temporal and spatial changes in TWS. Many of the existing climatic indices are insensitive to the impacts of human interventions on local or regional climates and none of them measures TWS, a sensitive measure of aridity or wetness.
in an area. I here advocate that consideration should be given to the use of GRACE data as an additional or alternative climatic indicator, where consideration should be given to the utility of the TWS as a valued and consistent measure for the wetness and aridity of areas across the globe.

It is intriguing that scientists can now start utilizing GRACE observations to make futuristic projections pertaining to the impacts of climate change, and human interventions on the wetness or aridity of areas world-wide, and to test the validity of climatic trends extracted global climatic models. Our success will largely depend on our ability to acquire long range gravity records and to improve on the spatial and temporal gravity measurements. It is encouraging that the scientific community has now agreed to launch a follow up mission of GRACE in 2016-2017. The GRACE Follow-on mission has the advantages of using more accurate ranging system through the use of interferometric laser ranging methods which in turn will improve the spatial resolution of the acquired data (Sheard et al., 2011; Stephens et al., 2011).
CHAPTER V
THE HYDROGEOLOGIC AND STRUCTURAL SETTING OF THE EL QAA PLAIN, SINAI, EGYPT

5.1 Introduction

The development of the arid and semiarid areas in the Middle East is hindered by the scarcity of its fresh water resources and the absence of a comprehensive understanding of the geologic and hydrogeologic settings of fresh water aquifers in these areas. Such an understanding provides the foundation for the development and the preservation of the natural resources in these arid and semi-arid areas. One such area is the country of Egypt, a country that is seeking additional fresh water resources to support its increasing population and is in need of such water resources to pursue its plans for modernization and development. The Nile River has been the vital surface fresh water resource for Egypt’s population and has been used for the development of its agricultural and industrial sectors. However, Egypt is currently using its total annual allocation of River Nile water estimated at $55 \times 10^9 \text{ m}^3/\text{yr}$ and alternative water resources have to be identified if Egypt is to sustain the needs of its increasing population is to continue its progressive plans for development.

Given the scarcity of surface fresh water resources in Egypt, and the difficulties and expenses entailed in channeling River Nile waters to regions that are distant from the Nile River Valley and Nile Delta in Egypt, groundwater aquifers remain one of the largest potential resources that could address Egypt’s deficiencies in fresh water resources and its growing demands. Examples of areas where the
Identification and development of groundwater aquifers could be quite valuable for Egypt's population include the Eastern and Western Deserts and the Sinai Peninsula. Egypt's economy depends largely on the tourism industry, especially the tourism along the Sinai which has been growing over the years. Along the Sinai coastlines, new cities, resorts, and tourism facilities are being constructed to support this important industry. The principal limitation on the development of such facilities is the paucity of fresh water supplies. The identification and further development of groundwater aquifers in these areas provides viable and cost-effective alternatives to the construction of desalinization plants and extensive pipelines to channel fresh water from the distant River Nile valley.

5.2 Objectives

Fortunately, in the Red Sea and Gulf of Suez coastal areas, many fresh water aquifers are concealed beneath the sands covering the mid-Tertiary rift-related structural basins. In this study, an integrated approach using ground and space-borne gravity (GRACE) and airborne magnetic data has been used to study rift-related aquifers along the coastal plain of the Gulf of Suez in Sinai, not all of which are beneath the waters of the Red Sea or Gulf of Suez. A significant example of rift-related aquifers is found in the El Qaa Plain in Sinai. Specifically, the study presented here addresses the identification of the structural and hydrogeologic setting of aquifers, their areal distribution, geometry, aquifer storage (water volume), and depletion rates.
5.3 Geologic Setting

The Gulf of Suez rift, the northward extension of the Red Sea rift (Said, 1990), is an inactive intercontinental rift (length: 300 km; width: 80 km) created by the stretching and collapse of the continental crust (Garfunkel and Bartov, 1977) between the Sinai micro plate and the African plate (Landon, 1994; Moustafa, 2002) (inset; Fig. 19A). The rift started in the Oligocene (Robson, 1971), Oligocene–Miocene (Garfunkel and Bartov, 1977), or early Miocene (Moustafa, 1993; Patton et al., 1994) and was associated with uplift; the rift shoulders were elevated by as much as 4 km, exposing the underlying crystalline rock and the overlying thick (up to 2.5 km) sedimentary successions to extensive erosion (Garfunkel and Bartov, 1977). The rifting was largely accommodated by extensional normal faults that strike north and northwest forming a complex array of half-grabens and asymmetric horst (Pivnik et al., 2003). The uplift removed the thick sedimentary successions, whereas the extensional faults preserved these successions as subsided blocks (half-grabens) under the Gulf of Suez, and the marginal coastal plains. The subsided blocks were subsequently filled by syn- and post-rift clastics shed from the uplifted blocks. An example of such basins is the El Qaa Plain basin which is bounded to the east by a major extensional fault (Said, 1990) (Fig. 19). The rift-related basins are separated by accommodation zones (Bosworth, 1985; Faulds and Varga, 1998; Lambiase and Bosworth, 1995; Rosendahl et al., 1986). The latter defined as zones of along-strike displacement transfer between oppositely dipping rift-border and intrarift faults, also termed “transfer zones” (Morley et al., 1990; Moustafa, 1997, 2002), “interference
accommodation zones” (Versfelt and Rosendahl, 1989), or “hinge zones” (Alsharhan, 2003). Two accommodation zones (inset; Fig. 19) were recognized north and south of the El Qaa Plain area, namely, the Zaafarana and Morgan accommodation zones (Moustafa, 1997).

The pre-rift stratigraphic section of the Gulf of Suez rift system consists of Precambrian basement that is unconformably overlain by Cambrian to late Eocene sedimentary successions (Patton et al., 1994; Said, 1990; Schutz, 1994). Thick sequences of Lower Cambrian–Lower Cretaceous clastics, dominantly shallow marine to fluvial sandstones (Allam and Khalil, 1989) also called the Nubian Sandstone (Said, 1962) are overlain by interbedded sandstones, limestones, and shales of the Raha, Wata, Matulla and Duwi formations (Ghorab, 1961). These in turn are overlain by thin Paleocene shales of the Esna formation and early to middle Eocene carbonates of the Thebes, Darat, Mokattam, Tanka, and Maadi formations (McClay et al., 1998; Patton et al., 1994). The pre-rift strata are unconformably overlain by late Oligocene–early Miocene red beds and volcanics that were deposited in isolated, fault-bounded basins (Said, 1962, 1990; Sharp et al., 2000). These successions were followed by early to middle Miocene thick (up to 6 km) (Bosworth et al., 1998) synrift clastics of the Gharandal and Ras Mallab groups (McClay et al., 1998; Sharp et al., 2000). The post-rift strata consist of Pliocene clastics and Quaternary wadi and sabkha deposits (Moustafa, 2004).
Figure 19: (A) Geologic map of the El Qaa Plain region showing the distribution of Precambrian outcrops and the Phanerozoic rock units along the Gulf of Suez eastern margin (modified from Klitzsch et al., [1987] and Moustafa [2004]). Also shown are the rift boundary (thick black line), main cities (red circles), main road (pink line), field gravity data (polygon a), aeromagnetic data (polygon b), GRACE spatial average area (polygon c), transfer zones (inset: red and blue lines), and thermal springs (inset: colored crosses). (B) Geologic cross sections along lines X-X’ and Y-Y’ shown in Figure 19A, modified from Moustafa (2004).
The El Qaa Plain basin has four aquifers (Quaternary alluvium, Lower Miocene clastics, Nubian Sandstone, and Precambrian fractured basement) and three aquitard units (massive Basement, Middle Calcareous Division, and Upper Miocene Evaporites) (Gorski and Ghodeif, 2000). The Nubian Sandstone aquifer is considered to be a nonrenewable aquifer, commonly referred to as a “fossil” aquifer. Fossil aquifers are believed to have been recharged under previous wet climatic periods (Sturchio et al., 2004) but may have also received modern meteoric contributions in intervening dry climatic periods such as at present (Sultan et al., 2008; Sultan et al., 2007). Using continuous rainfall runoff models for the period 1998-2007, a conservative estimate for the average annual modern recharge of the Nubian Aquifer in Sinai was estimated at 13 x 10^6 m^3/yr (Sultan et al., 2011a). The fossil groundwater in the Nubian Sandstone aquifer discharges along the extensional faults bounding the Gulf of Suez at high temperatures (up to 70°C) within the coastal plain of the Gulf of Suez and along its coastline (Sturchio et al., 1996). The groundwater flow directions are northward, toward the Mediterranean Sea, and westward, toward the Gulf of Suez, where discharge occurs along the faults defining the Gulf (Gorski and Ghodeif, 2000; Issar, 1979). Thus, the groundwater in the aquifers within study area is probably largely composed of fossil precipitation over the mountains in southern Sinai. Additional contributions to these aquifers, especially the unconfined aquifers, from modern precipitation over the coastal plain could not be ruled out.
5.4 Data Acquisition and Data Processing:

The ground-based gravity and aeromagnetic data (Fig. 20) acquisition parameters and processing steps were discussed in details in Chapter II (sections II.3 and II.4).

Figure 20: Location map showing the spatial distribution and the measurement time of the gravity stations (colored circles) and the locations of the flight paths for the aeromagnetic survey (black lines). Also shown are the locations of the 2D modeling profiles (purple lines).
5.5 Basin Geometry and Boundaries Outlines

Both gravity and magnetic anomalies reveal the spatial distribution and the relief of the subsurface basement rocks in the study area. The anomalies of the deeper sources are characterized by longer wavelength and are smooth over considerable distances, while those of the shallow sources are sharper and of shorter wavelengths (Cloke et al., 1999).

Examination of the complete Bouguer gravity anomaly map (Fig. 21A; Contour interval [CI]: 2 mGal) shows a NW–SE trending gravity low, parallel to the Gulf of Suez, occupying the central and northern parts of the map area. This anomalous feature indicates a thick sedimentary section (i.e., basin) overlaying a complex trough of basement rocks. Contours are closer spaced along the ENE side of the basins, indicating the location of the controlling fault of the half-graben. The gravity low anomaly feature expressed as two elliptical shapes (north: -45 mGal; south: -48 mGal) represents two sub-basins connected by a narrow saddle (-42 mGal). The two basins are connected to the Gulf of Suez by N-S trending channel-like feature of low (-35 mGal) gravity. Examination of the southern part of the map shows a gravity high (9 mGal) which indicates the presence of shallow basement rocks to the south, or the presence of a high-density intrusive body within the country rock.

The residual magnetic anomalies map (Fig. 21B; CI: 10 nT) shows a NW-SW magnetic low parallel to the Gulf of Suez. The magnetic low is represented by three (north: -90nT, central: -100 nT, and south: -70 nT) near circular magnetic anomalies separated by two narrow saddles. These anomalous features represent three basins.
with a thick sedimentary section overlying deep basement rocks. The central and southern anomalies are equivalent to those shown in the complete Bouguer anomaly map (Fig. 21A). The difference in the geometries between gravity and magnetic anomalies is here attributed to one or more parameters including induced and remnant magnetization(s), differing magnetic susceptibility of rocks with the same density, and the depth, strike, and magnetic inclination and declination of the source body.

Figure 21: (A) Complete Bouguer anomaly (Contour interval [CI]: 2 mGal), and (B) Total magnetic intensity (CI: 10 nT) maps of the El Qaa Plain.
The analytic signal and the derivatives (horizontal and vertical) were used to outline the areal distribution of the sources causing the gravity and/or magnetic anomalies. Horizontal derivatives were used in resolving composite and complex anomalies into their individual components. Features striking perpendicularly or at high angles to the direction of the applied derivatives were enhanced and vice versa for features that strike in all other directions. The vertical derivative was used to increase the resolving power of the shallow features at the expense of the deeper anomalies (Telford et al., 1990). The analytic signal (using all 3 derivatives) was used to define and delineate the main boundaries of the features causing the gravity and magnetic anomalies regardless of the structural dip of these features, and independent of the direction of the induced and/or remnant body magnetizations. The boundaries of these features are located directly beneath the analytic signal maxima (MacLeod et al., 1993; Nabighian, 1972; Roest et al., 1992).

Inspection of the horizontal and vertical derivatives maps of both gravity and magnetic data clearly shows the boundaries of the subsurface basins causing the observed anomalies. The horizontal derivative in X-direction (dx: Figs. 22A and 23A) clearly shows the basin boundaries that are striking in direction perpendicular to X-direction (black dashed line; Figs. 22A and 23A). Similarly, the horizontal (vertical) derivative in Y (Z)-direction (dy: Figs. 22B and 23B; dz: Figs. 22C and 23C) shows the basin boundaries that are striking in a direction perpendicular to Y (Z)-direction (black dashed line; Figs. 22B,C and 23B,C). Examination of the analytic signal maps (Figs. 22D and 22D) of both gravity and magnetic data clearly shows the boundaries.
of the subsurface basins located beneath the maximum amplitude of the analytic signal (black dashed line; Figs. 22D and 23D). Based on the examination of the gravity and magnetic field derivatives and the analytic signal (Figs. 22 and 23) a subsurface basin striking NW–SE direction, parallel to the Gulf of Suez, with a length of 48 km and an average width of 16 km has been revealed.

Calculation of the depth to the basement rock is an important step in defining the spatial variations in the thickness of the sedimentary cover and delineating the structural relief of the basement and its effect on the overlying sedimentary units. In this study a two-dimensional (2D) modeling was carried out simultaneously for both gravity and magnetic data to outline the shape and the surface of the underlying basement and to define the thickness and geometry of the overlying sedimentary cover in the selected basins. The forward modeling was carried out after removing the linear regional field. A trial and error technique was used to adjust the initial model to match the observed anomalies, where the configuration of the model layers was iteratively modified for best fit between the observed and the computed anomaly data. An understanding of the subsurface geology as well as knowledge of the physical properties (e.g., density) and relevant geophysical characteristics (e.g., magnetic susceptibility, and the total magnetic intensity, inclination, and declination of the magnetic field) for the investigated sequences are needed to develop sound 2D models.
Figure 22: (A) Horizontal derivative in the X direction, (B) Horizontal derivative in Y direction, and (C) Vertical derivative in Z direction and (D) the analytic signal of the gravity field measured at the El Qaa Plain.
Figure 23: (A) Horizontal derivative in the X direction, (B) Horizontal derivative in Y direction, and (C) Vertical derivative in Z direction and (D) the analytic signal of the magnetic field measured at the El Qaa Plain.
The density and magnetic susceptibility data used in this study were compiled from previously published studies (Abdelrahman et al., 1988; El-Gezeery, 2006; Ismail, 1998; Makris, 1977; Omran, 1982; Rabeh and Miranda, 2008; Rabeh et al., 2009; Selim, 2002; Setto, 1991; Sultan et al., 2009). The parameters of the geomagnetic field were extracted from the National Oceanic and Atmospheric Administration (NOAA) National Geophysical Data Center (NGDC) website (available at: www.ngdc.noaa.gov/geomagmodels/IGRFWMM.jsp). I created a three-layer, 2-D model with two clastic layers overlying a layer of basement rock with densities (from top to bottom) of 2.27 (syn- and post-rift clastics), 2.55 (pre-rift sediments), and 2.67 gm/cm$^3$, respectively, and magnetic susceptibilities of 0, 0, 2.5 x 10$^{-3}$ cgs, respectively where the total magnetic field, inclination and declination were found to be 42187 nT, 40.86° and 2.35° respectively.

Four 2D gravity and magnetic profiles, trending NE-SW and NW-SE, covering the anomalous features were selected during this study (Fig. 24). Examination of the cross sections A-A’ and C-C’ (Fig. 24) shows a dipping and thickening of the sedimentary section toward the east direction. The thickness of the sedimentary section increases from the west (A-A’: 0.15 km; C-C’: 0.7 km) to the east (A-A’: 2.7 km; C-C’: 3.75 km). Inspection of the cross sections B-B’ and D-D’ (Fig. 24) shows a thick (B-B’: 2.9 km; D-D’: 0.55 km) sedimentary section at the central part of the profile which decreases in thickness toward the eastern (B-B’: 1 km; D-D’: 0.1 km) and the western (B-B’: 1.5 km; D-D’: 0.05 km) margins of the profile.
Figure 24: 2D modeling of both gravity and magnetic data along profiles A-A’, B-B’, C-C’, and D-D’. The profile locations are shown on Figure 20. D – density and k – magnetic susceptibility.
5.6 Volume of Water

While an infinite number of geometries and density distributions can give rise to a single residual gravity map of an area, a remarkable property of the gravity method is that the total anomalous mass is precisely determined by the areal integration of the residual gravity anomaly. The total anomalous (ΔM; missing in this case) mass is calculated using Gauss’s Theorem (Ramsey, 1949) and it is directly related to the integral of the residual gravity anomalies (∆g(x, y)) over each unit cell (dxdy) included in the selected basin.

\[
\Delta M = \frac{1}{2\pi G} \int \int_{-\infty}^{\infty} \Delta g(x, y) dxdy
\]

If I assume that the missing mass is due entirely to two layers, air-filled pore space in the vadose zone, and water-filled pores below, it is possible to calculate the mass, and hence the volume of water in the basin. The adopted model depicts a basin of an area (S) filled with alluvium sediments; the lower surface of the alluvium is bound by bedrock. Above the water table, the alluvium sediments with a thickness (h) have a matrix density (ρ) and the pore spaces within the sediment are filled with air (density contrast [Δρ] -2.6), whereas below the water table, the pore spaces are saturated with water (density contrast -1.6). The total anomalous missing mass (ΔM) is the sum of the missing mass below the water table (ΔM₁: saturated zone) and the missing mass above the water table (ΔM₂: vadose zone).
In this study, I used a contour line of -36 mGal to define the basin boundary (0 mGal after separating the residual anomaly). The basin area \( S \), the integral of the residual gravity anomalies within that basin \( \int_{-\infty}^{\infty} \int_{-\infty}^{\infty} \Delta g(x, y) \, dx \, dy \), and the total mass \( \Delta M \) was estimated at 518.5 km\(^2\), -2.76 x 10\(^9\) mGal.m\(^2\), and -6.59 x 10\(^{13}\) kg, respectively. Knowing the matrix density, porosity (\( \phi \)) and the depth to water (\( h \)) one can calculate the vadose zone mass (\( \Delta M_2 \)). I used a matrix density of 2.6 gm/cm\(^3\), porosity of 20% and average depth to water table of 35m (Ghodeif et al., 2002; Sultan et al., 2009). The missing masses due to the vadose zone (\( \Delta M_2 \)) and the saturated zone (\( \Delta M_1 \)) were estimated at -9.43 x 10\(^{12}\) kg and -5.64 x 10\(^{13}\) kg respectively.

Knowing the mass of the saturated zone the volume of water \( V_w = \frac{\Delta M_1}{\rho_w} \) in the basin can be calculated to be approximately 56 km\(^3\).

It is possible that part of the missing mass could be due to the presence of evaporites in the early syn-rift sedimentary section. Based on the reported occurrences of evaporites in the Gulf of Suez and surroundings, the evaporites if present, are likely to be halite (average density: 2.2 gm/cm\(^3\); Telford, et al., 1990) or anhydrites (average density: 2.85 gm/cm\(^3\); Telford, et al., 1990). In either case, the calculations that were conducted assuming that all the missing mass is solely related to water-filled pore
spaces would have to be corrected to account for contributions from the presence of evaporites. The presence of evaporites could mask or diminish the extracted volumes and the correction for anydrite will increase the calculated water volume, while the presence of halite would decrease the water contribution to the negative anomaly. Inspection of published geologic maps (Klitzsch et al., 1987) and available sparse borehole information did not reveal the presence of significant evaporite successions in El Qaa Plain basin. Nevertheless, the presences of 50 m thickness of halite will reduce the volume of water to 40 km$^3$ if I assumed the water table to be at 45 m and the vadose zone porosity to be 25%.

5.7 Aquifer Depletion Rate

Few studies provide reliable estimates for the temporal and spatial variations in natural and man-made discharge in the El Qaa Plain area due to the paucity of the required field data. I addressed the temporal and spatial variability in the groundwater depletion including the human extraction or the natural discharge across the El Qaa area using GRACE data. The natural discharge occurs in thermal spring at Ain Sokhna, Ayun Mussa, Hammam Faraoun and Hammam Mousa (inset; Fig. 19). Also, the El Qaa aquifers are considered to be the main fresh water source that supplies the Egyptian cities of El-Tor and Sharm El-Sheikh and sustains several agricultural and industrial projects and other activities.
Inspection of the time series (Fig. 25), constructed from spatially averaged GRACE solutions along the El Qaa Plain area (polygon c; Fig. 19; area: 3,500 km²), indicates a negative trend in the GRACE-derived Terrestrial Water Storage (TWS). The negative TWS trend is probably due to excessive water extraction combined with natural discharge. The negative TWS is estimated at $19 \times 10^6$ m³/yr (~5.5 mm/yr equivalent water layer). GRACE results are consistent with the field observations, where the groundwater extraction rates in the Quaternary alluvium aquifer are increasing with time and where the extraction rate was $0.18 \times 10^6$ m³/yr in 1930.
(Attia, 1930), and increased to $1.1 \times 10^6$ m$^3$/yr in 1972 (Gilboa, 1972), and became $9.5 \times 10^6$ m$^3$/yr in the year 2000 (Gorski and Ghodeif, 2000) and recently (2011) reached $11.0 \times 10^6$ m$^3$/yr (Ghodeif, personal communication, 2012).

5.8 Conclusions

I applied an integrated approach including ground gravity, aeromagnetic, and GRACE satellite techniques to identify the structural, hydrogeologic, the areal distribution and geometry, water volume, and water depletion rates in one of the Gulf of Suez rift-related basins, El Qaa Plain basin in Sinai. Analysis and interpretation of both gravity and magnetic data revealed a NW-SE trending basin, parallel to the Gulf of Suez, with an average length of 48 km and average width of 16 km. The 2-D modeling of both gravity and magnetic data indicates basin fill in the graben area with maximum thickness of 3.75 km along the graben axis. Using anomalous residual gravity techniques, the volume of water in storage was estimated to range from 40 (conservative limit) to 56 (upper limit) km$^3$ in the selected basin. Analysis of GRACE data reveals a depletion rate of $19 \times 10^6$ m$^3$/yr which is interpreted here to reflect losses due to natural discharge and extraction. The techniques applied here could be readily extended to investigate several similar aquifers in half-graben on the west side of the Gulf of Suez. Such studies could potentially assist in alleviating the increasing demands for fresh water supplies by the rapidly growing tourism industry in the region.
An integrated (space-borne and land-based gravity, aeromagnetic, GIS, and remote sensing) approach was applied to investigate the hydrologic, and geologic settings at two different spatial scales, namely, on the regional scale across the African continent and on the local scale over the El Qaa Plain, in the Sinai Peninsula.

On the regional scale (Africa), the Gravity Recovery and Climate Experiment (GRACE) monthly (04/2002 to 08/2011) data were used, with others extracted from a wide range of readily available global remote sensing (i.e., Tropical Rainfall Measuring Mission [TRMM], Landsat TM, DEM, stream networks, slope, and water bodies), geologic, and climatic datasets, to address the following: (1) the nature of, and the factors controlling, the observed temporal GRACE mass variations over the African watersheds on sub-basin scales; (2) the capability of GRACE data for monitoring elements of hydrologic systems (i.e., runoff, recharge, groundwater flow) on the sub-basin level; (3) the spatial and temporal distribution of the long term inter-annual trends in the GRACE-derived terrestrial water storage (TWS); and (4) the nature and the controlling factors (e.g., climatic and/or human pressure-related) affecting these TWS trends over a much larger suite of hydrologic systems (e.g., watersheds, lakes, rivers, marshes) and domains (e.g., source areas, lowlands) within watersheds, sub-basins, and aquifers across Africa.
On the local scale, the ground-based gravity, aeromagnetic as well as GRACE data were used to identify or quantify the following: (5) structural, geologic and hydrologic settings of the El Qaa Plain aquifer; (6) characteristics (geometry, area, saturated volume, volume of water) of the selected aquifer; and (7) aquifer depletion rates.

Findings and results regarding the first and the second objectives include the following: (1) large persistent anomalies (standard deviation [SD] > 10 cm) on SD images over periods of 2 to 10 years; (2) anomalous areas originate at mountainous source areas that receive high precipitation, extend down slope toward mountain foothills, and often continue along main channels, wetlands, or lakes that drain these areas; (3) time series analyses over anomalous areas showed that seasonal mass variation lags behind seasonal precipitation; (4) seasonal mass variations showed progressive shift in phase and decrease in amplitude with distance from the mountainous source areas; (5) the observed temporal mass variations are largely controlled by elements of the hydrologic cycle (e.g., runoff, infiltration, groundwater flow) and have not been obscured by noise as was previously thought, (6) it is possible to use GRACE to investigate the temporal responses of a much larger suite of hydrologic systems (watersheds, lakes, rivers, marshes, etc.) and domains (e.g., source areas, lowlands) within watersheds and sub-basins world-wide.

Findings regarding the third and the fourth objectives, include the following: (1) large sectors of Africa are undergoing statistically significant variations (+36 mm/yr to –16 mm/yr) due to natural and man-made causes; (2) warming of the
tropical Atlantic ocean apparently intensified Atlantic monsoons and increased precipitation and TWS over western and central Africa’s coastal plains, proximal mountainous source areas, and inland areas as far as central Chad; (3) warming in the central Indian Ocean decreased precipitation and TWS over eastern and southern Africa; (4) the high frequency of negative phases of the North Atlantic Oscillation (NAO) increased precipitation and TWS over northwest Africa; (5) deforestation in the Congo Basin and southern Tanzania decreased TWS; (6) the construction of dams throughout the GRACE period increased TWS in upstream Nile Valley countries; and (7) given the 10-year monthly GRACE record of water availability data (represented by GRACE-derived TWS) acquired on the sub-basin scale across the globe, and the plans underway for deployment of a GRACE follow-up (2016–2026), consideration should be given to using GRACE TWS data as an alternative, viable index for measuring temporal and spatial variations in aridity.

Regarding the fifth, sixth, and seventh objectives, examination of the complete Bouguer anomaly and the residual total magnetic intensity maps shows two connected sub-basins separated by a narrow saddle with an average basin length of 48 km and an average width of 16 km. The 2D modeling of both gravity and magnetic data indicates basin fill in the graben area with maximum thickness of 3.75 km along the graben axis. Using anomalous residual gravity, the volume of water in storage was estimated at 40 to 56 km$^3$ in the selected basin. Analysis of GRACE time series revealed a depletion rate of $19 \times 10^6$ m$^3$/yr that is probably related to natural discharge and extraction. Similar geophysical exploration campaigns are here recommended over
the entire coastal plains of the Gulf of Suez in the Sinai Peninsula and in the Eastern Desert. Such studies are important for the development of sound and sustainable management schemes for the fresh water resources in these areas.
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APPENDIX

Scaling of GRACE TWS
1 GRACE Errors:

Errors in GRACE-derived TWS fall into two main categories (Wahr et al., 2006): (1) Errors in monthly GRACE solutions. These include measurement and processing errors, aliasing errors related to the short-period (sub-monthly) variations and errors in the European Centre for Medium-Range Weather Forecasts (ECMWF) models used for atmospheric corrections of GRACE data; (2) Errors due to changes caused by factors other than continental water storage (i.e., leakage errors). The leakage errors can come from time variable mass anomalies either vertically above or below, or from mass anomalies off to the side of the area of interest. The time variable gravity below the area of interest might include the gravity signal related to un-modeled mass variations in the Earth’s interior while that above the area of interest might be related to errors in the atmospheric models (ECMWF). Leakage errors can also come from continental water storage signals where it comes from water storage outside of the region of interest.

The measurement errors in GRACE data (type 1) exhibits a zonally banded pattern and varies with the Gaussian smoothing function used and with the latitude. At lower latitudes, a maximum error of ~36 mm (equivalent water thickness) was obtained with a Gaussian radius of 300 km whereas an error of ~28 mm was obtained when using a Gaussian radius of 750 km was adopted. Poleward, the error decreases to <15 mm (Landerer and Swenson, 2012; Wahr et al., 2006).

Estimates of TWS variations suffer from signal degradation due to noise. The noise is manifested as: (1) random errors that increase as a function of spherical
and (2) systematic errors that are correlated within a particular spectral order (Swenson and Wahr, 2006).

Several filtering techniques are used to damp or isolate and remove the GRACE-derived TWS errors. The problem with most of those techniques is that the filters also modify the true geophysical signal that the researchers are interested in. The following section explains the methods used to scale GRACE TWS data to account for the effect of the filter on the GRACE signal.

2 Scaling GRACE TWS Data

The scaling process (Landerer and Swenson, 2012) is supposed to take care of the GRACE TWS errors that result from applying the following filters and/or truncations: (1) destriping filter designed to remove systematic errors that are characterized by correlations between certain spherical harmonic coefficients (Swenson and Wahr, 2006); (2) Gaussian (radius = 250 km) smoothing filter when applied, it smoothes, but reduces the spatial resolution of GRACE observations by damping the higher degree coefficients (Wahr et al., 1998); and (3) truncation of the monthly GRACE solutions where GRACE gravity solutions are typically truncated at a spectral degree (lmax ≤ 60; wavelength of ~330 km). This truncation means that GRACE can not resolve signals with spatial variability finer than 330 km (Wahr et al., 1998; Wahr et al., 2006).

The filtering effects on the actual GRACE signal were quantified based on realistic TWS models (Swenson et al., 2003; Seo and Wilson, 2005). Monthly TWS
anomalies simulated by the NOAH land model, running within the Global Land-Data Assimilation System (GLDAS-NOAH (Rodell et al., 2004b); available at: ftp://podaac-ftp.jpl.nasa.gov/allData/tellus/L3/gldas_monthly/) were used in this study. To generate the scaling factor, the monthly (April 2002 - August 2011) TWS estimates from GLDAS-NOAH were processed as follows: (1) the TWS estimates from GLDAS-NOAH were converted to Spherical Harmonic Coefficients truncated to a degree and order 60; (2) to match the processing steps applied to GRACE data, the GLDAS-NOAH fields were then destriped using the Swenson and Wahr (2006) approach; (3) the Gaussian smoothing filter of radius 250 km was then used to convert the GLDAS solutions to 0.5 x 0.5° grids; (4) the gain factor (k) was then extracted for each grid point by applying the root-mean square difference method. For each grid point, the selected gain factor minimized the misfit between the unfiltered, and filtered TWS estimates through a simple least square regression (Landerer and Swenson, 2012).
Figure A: The Scaling factors for GLDAS-NOAH monthly TWS variations derived by least square fitting each filtered grid-point time series to the unfiltered time series.
3 The Scaling Factor

Figure (A) shows that the scale factor is larger than 1 for areas at and close to the coastline. These values are required due to signal interference with the much weaker ocean signal (Landerer and Swenson, 2012). The figure also shows that most of Northern Africa, with low TWS variability, has a scale factor less than 1 which indicates leakage errors from larger signals of surrounding regions. These scale factor values are used to reduce the signal amplification along those regions.

To address the effect of using the scaling factor to restore the original signal we applied scaling factor (Fig. A) to the trend image generated from the monthly (04/2002 to 08/2011) GLDAS-NOAH data (Fig. B). The resultant scaled trend image (Fig. C) shows that the scaling factor does a reasonable job of reproducing the correct answer and it doesn’t capture the short-scale variability.
Figure B: Linear trend image generated from original (1° x 1°) monthly (04/2002 – 08/2011) GLDAS-NOAH TWS data.
Figure C: Linear trend image generated from processed (destriped, Gaussian 250 km smoothed, then scaled) monthly (04/2002 – 08/2011) GLDAS-NOAH TWS data.