AGU PUBLICATIONS

Water Resources Research

RESEARCH ARTICLE

10.1002/2016WR018846

Key Points:

- The GRACE satellites' capability to monitor groundwater is evaluated in complex basement aquifers of East Africa
- Systematic data filtering reduces leakage uncertainties in the GRACE groundwater estimates in comparison to conventional scaling approaches
- Significant negative storage changes attributed to groundwater use and/or climate variability are revealed

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Citation:
Nanteza, J., C. R. de Linage, B. F. Thomas, and J. S. Famiglietti (2016), Monitoring groundwater storage changes in complex basement aquifers: An evaluation of the GRACE satellites over East Africa, *Water Resour. Res.*, *52*, 9542–9564, doi:10.1002/2016WR018846.

Received 26 FEB 2016
Accepted 22 OCT 2016
Accepted article online 1 NOV 2016
Published online 25 DEC 2016

Monitoring groundwater storage changes in complex basement aquifers: An evaluation of the GRACE satellites over East Africa

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Abstract Although the use of the Gravity Recovery and Climate Experiment (GRACE) satellites to monitor groundwater storage changes has become commonplace, our evaluation suggests that careful processing of the GRACE data is necessary to extract a representative signal especially in regions with significant surface water storage (i.e., lakes/reservoirs). In our study, we use cautiously processed data sets, including GRACE, lake altimetry, and model soil moisture, to reduce scaling factor bias and compare GRACE-derived groundwater storage changes to in situ groundwater observations over parts of East Africa. Over the period 2007–2010, a strong correlation between in situ groundwater storage changes and GRACE groundwater estimates (Spearman's $\rho = 0.6$) is found. Piecewise trend analyses for the GRACE groundwater estimates reveal significant negative storage changes that are attributed to groundwater use and climate variability. Further analysis comparing groundwater and satellite precipitation data sets permits identification of regional groundwater characterization. For example, our results identify potentially permeable and/or shallow groundwater systems underlying Tanzania and deep and/or less permeable groundwater systems underlying the Upper Nile basin. Regional groundwater behaviors in the semiarid regions of Northern Kenya are attributed to hydraulic connections to recharge zones outside the subbasin boundary. Our results prove the utility of applying GRACE in monitoring groundwater resources in hydrologically complex regions that are undersampled and where policies limit data accessibility.

1. Introduction

Accessibility to freshwater resources for drinking water and irrigation is an important issue that the global water management community strives to address amidst nonstationary hydro climatology [*Milly et al.*, 2005] and ever-increasing water demands as population grows [*Vörösmarty et al.*, 2000]. Over 1 billion people in the world [*Morris et al.*, 2003; *Alley*, 2006], a majority of whom reside in sub-Saharan Africa, currently have no access to safe drinking water [*UN-Water*, 2007]. The global reliance on groundwater to meet water demands has increased during the last few decades [*Giordano*, 2009; *Famiglietti and Rodell*, 2013; *Famiglietti*, 2014] due to factors including rising populations, agriculture, industry [*Gleick*, 2000; *Morris et al.*, 2003], and climatic stressors including drought [*Scanlon et al.*, 2012; *Castle et al.*, 2014]. These factors combined with limited to nonexisting water and land policies and poor governance [*Mogaka*, 2006; *Giordano*, 2009] render groundwater resources in many parts of the world susceptible to overexploitation and depletion. These factors have reportedly contributed to groundwater depletion at regional [e.g., *Dhawan*, 1995; *Changming et al.*, 2001; *Karami and Hayati*, 2005; *Rodell et al.*, 2009; *Scanlon et al.*, 2012; *Thomas and Famiglietti*, 2015] and global scales [*Wada et al.*, 2010; *Schwartz and Ibaraki*, 2011; *Gleeson et al.*, 2012; *Taylor et al.*, 2013; *Döll et al.*, 2014; *Richey et al.*, 2015a, 2015b].

Sustainable water management is emerging as a high priority issue to ensure national and international security, in both the developing and developed worlds, given the importance of water to food and energy production, to economic growth, and to human and environmental health. In many parts of sub-Saharan Africa including East Africa, the urge for economic development and the need for increased food security

[*Rosegrant and Cline*, 2003; *UN-Water*, 2007; *Sun et al*., 2010; *Sasson*, 2012] may greatly impact water availability, specifically groundwater [*Pavelic et al*., 2012]. In Africa, 60% of the population relies on groundwater to meet water demands including drinking water, especially in the rural regions without access to municipal water systems [*MacDonald et al*., 2009]. The widespread reliance on groundwater in sub-Saharan Africa [*Pavelic et al*., 2012] coupled with the limited knowledge and uncertainties about the impact of nonstationary climates and population growth on future groundwater availability pose new and as yet unquantified threats to water security in the region. To effectively and sustainably manage the resource, an understanding of regional groundwater dynamics is necessary.

Water resources monitoring is a limited practice in many parts of the world [*Giordano*, 2009; *Sophocleous*, 2010; *Scanlon et al*., 2012]. Groundwater in particular, unlike surface water, has received limited monitoring [*Robins et al*., 2006; *Pavelic et al*., 2012]. In aquifer systems that are monitored, hydrologists encounter sparse observation networks, discontinuities in data collection leading to data gaps and restricted institutional data sharing policies limiting access to essential observational data [*Famiglietti et al*., 2011]. This is true especially in Africa, where hydrologic data records are often impeded as a result of limited government funding and regional conflicts [*UNECA*, 2011; *Jacobsen et al*., 2012].

Since the 1970s, preexisting (e.g., altimetry satellites such as Jason-1 and 2, Topex/Poseidon, and ENVISAT) and newly launched satellite missions (e.g., the Gravity Recovery and Climate Experiment and the Soil Moisture and Ocean Salinity missions) have been (re-)focused on the monitoring of freshwater resources and provide global, time-variable observations of the terrestrial water storage [*Swenson and Wahr*, 2009; *Birkett et al*., 2011; *Crétaux et al*., 2011]. In particular, NASA's Gravity and Climate Recovery Experiment (GRACE) mission provides unprecedented, exceptionally clear observations of seasonal and year-to-year climate variations and human water use on water availability and stress [*Richey et al*., 2015a, 2015b]. The GRACE satellites provide monthly estimates of the Earth's gravity field at a 400 km spatial resolution at the equator [*Wahr et al*., 1998, 2004; *Tapley et al*., 2004] that can be converted over land into monthly estimates of vertically integrated storage, namely total water storage anomalies (TWSA) [*Syed et al*., 2008], including changes in snow water equivalent anomalies (SWEA), surface water anomalies (SWA) from lakes/reservoirs and rivers, soil moisture anomalies (SMA), vegetation canopy anomalies (VCA), and groundwater anomalies (GWA). Knowledge of these water budget components (SWEA, VCA, SMA, and SWA) can be utilized to deduce groundwater anomalies from TWSA using a simple water budget equation [*Rodell and Famiglietti*, 1999].

$$GWA_t = TWSA_t - SWA_t - SMA_t - SWEA_t - VCA_t. \tag{1}$$

Across East Africa (Figure 1), studies have applied GRACE together with other satellite observations, including the Tropical Rainfall Monitoring Mission (TRMM) and CHAllenging Minisatellite Payload (CHAMP), in the examination of the hydrologic behaviors of major lake basins in the last decade [*Awange et al*., 2007; *Swenson and Wahr*, 2009; *Becker et al*., 2010; *Hassan and Jin*, 2014]. *Awange et al*. [2007] examined the cause of the decline in Lake Victoria water levels during 2002–2006. The authors, assuming that changes in the lake levels were directly related to changes in its catchment water storage, deduced that the 80% reduction in the lake's catchment water storage was due to human management. Although lake levels may portray similar changes to basin total water storage changes over longer timescales, the case may not be true for highly managed basins like Victoria that are underlain by complex crystalline aquifer systems. *Swenson and Wahr* [2009] examined the cause of the declining levels in East African lakes (Victoria, Tanganyika, and Malawi) by comparing trends in a given lake's water storage to trends in its catchment water storage as observed from GRACE during 2003–2008. The study found that both human management and climatic variations contributed similarly to the water storage declines in Lake Victoria while changes in precipitation contributed to Lake Tanganyika and Malawi declines. Similar to *Awange et al*. [2007], *Swenson and Wahr* [2009] assumed that changes in the lake levels are related to changes in groundwater storage. *Becker et al*. [2010] and *Hassan and Jin* [2014] evaluated the behavior of East Africa's major lakes during the period 2003–2008 and 2003–2012, respectively. *Becker et al*. [2010] compared variations in GRACE TWSA, lake storage, and subsurface storage (the combination of groundwater and soil moisture) and found that individual lake storage governed total water storage (TWS) in lakes Victoria and Turkana basins while subsurface storage variations mainly governed Tanganyika storage. *Hassan and Jin* [2014] applied seasonal trend decomposition and correlation analyses to evaluate the relationship between lake storage, soil moisture, and GRACE TWSA. Attributing their results to the percentage of the lake's surface area relative to the basin area,

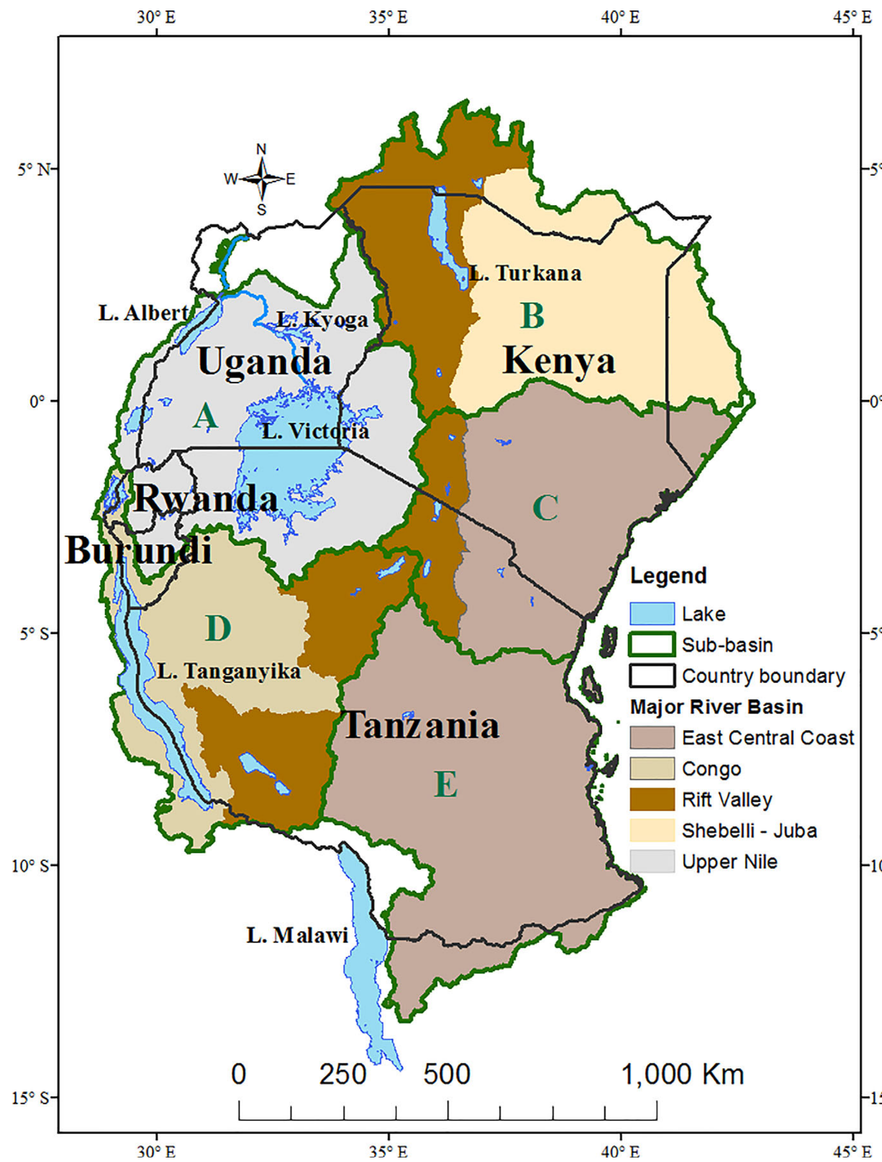

Figure 1. Map of East Africa showing the countries, regional Lakes, and alienated subbasins adapted from the Hydrological data and maps based on **S**huttle Elevation Derivatives at multiple Scales (HydroSHEDS) [*Lehner et al.*, 2008].

the study found that TWS changes in lakes Malawi and Victoria basins were governed by the water stored in the individual lakes while that of Tanganyika was governed by both lake storage and soil moisture. *Awange et al*. [2014a] evaluated the impacts of global climate change on the TWS within the Nile basin using global climate forcing by the Indian Ocean Dipole (IOD) and El Nino Southern Oscillation (ENSO). Their results suggest that the Lake Victoria basin in East Africa experienced effects of anthropogenic and climate variability during 2002–2011 with ENSO and IOD both influencing TWS variability during the study period. Despite the well-recognized use of groundwater resources in the region, none of the above studies have explicitly evaluated groundwater storage behavior and associated trends, the main focus of our study.

Our study uses a combination of remote sensing (i.e., GRACE and altimetry products) and land surface model simulations to evaluate groundwater storage changes across East Africa. The use of GRACE for hydrologic studies necessitates application of a number of data postprocessing techniques aiming at filtering systematic and random errors [*Swenson and Wahr*, 2006]. Filtering of GRACE data along with their intrinsic low resolution (about 400 km at the equator) leads to signal attenuation and introduces bias in the mass changes as they get smoothed, and translated, an effect known as spatial leakage [*Swenson*, 2002; *Landerer and*

Swenson, 2012]. Spatial leakage from surface water bodies [*Longuevergne et al*., 2013] is anticipated over East Africa, especially within the Lake Victoria basin [*Awange et al*., 2014a]. A number of recent studies have applied several methods to correct for spatial leakage/bias from surface water storage including Independent Component Analysis [*Awange et al*., 2014a], kernel functions [*Moore and Williams*, 2014], and statistical decomposition [*Forootan et al*., 2014]. However, for several GRACE applications, the common method applies land surface model (LSM) output to derive scaling factors that are multiplied with the GRACE data [*Landerer and Swenson*, 2012] in order to restore the lost signal and improve their spatial resolution up to 100 km at the equator.

The simplified assumptions included in parameterization of the land surface processes, including a poor representation of the natural and human controls on groundwater storage behaviors and no dynamical representation of lake storage [*Zeng*, 1999; *Swenson and Milly*, 2006; *Han et al*., 2010; *Werth and Geuntner*, 2010], limit the use of LSM-based scaling techniques over regions with large and highly variable surface water resources like East Africa [*Kull*, 2006; *Awange et al*., 2007; *Long et al*., 2015]. Moreover, the storage constituents may have different spectral contents making the use of the same scaling factor for all the temporal frequencies inappropriate [*Landerer and Swenson*, 2012]. Also, scaling factors are not representative when hydrological signals are not significantly correlated in space (such as regions with surface water resources) whereby leakage from neighboring regions cannot be corrected using a simple scaling factor. Thus, if LSM-based approaches are used to scale GRACE data, associated biases are propagated to the groundwater estimates, limiting their accuracy.

In this study, we attempt to reduce the influence of lake leakage from Lake Victoria and other large water bodies in our GRACE groundwater storage estimates over East Africa by applying a methodical postprocessing of the GRACE data and judicious corrections for soil moisture and lake storage changes estimated from satellite altimetry observations (see section 3 for details). We specifically use an alternative approach to estimate groundwater storage anomalies, which consists of using the GRACE TWSA in their filtered, unscaled form [*Landerer and Swenson*, 2012] from which we subtract similarly filtered SMA and SWA (provided by an LSM and derived from satellite altimetry, respectively), using equation (1) [*Scanlon et al*., 2012; *Longuevergne et al*., 2013]. We did not include SWE and VCA (equation (1)) in the analysis as these two variables do not contribute significantly to total water storage over the region based on our analysis of model outputs from GLDAS (i.e., snow contributes 0% while canopy contributes less than 1% to TWSA). We hypothesize that using filtered water budget component data sets along with observed lake storage anomalies in isolating a groundwater signal from GRACE, more appropriately captures groundwater storage changes over East Africa, compared to the alternative method of scaling the raw data and subtracting high-resolution LSM estimates of known storage constituents, as it reduces scaling factor bias resulting from inaccurate assumptions underlying the scaling approaches. To validate our approach, we compare GRACE-derived groundwater storage changes to ground-based observations from monitoring wells located throughout Uganda. The sections that follow discuss the study area and its hydro-climatic characteristics (section 2), data sets and methods (section 3), results and discussions (section 4), and conclusions (section 5).

2. Study Area and Its Hydro-climatic Characteristics

The East Africa (EA) region encompasses five countries: Kenya, Uganda, Tanzania, Rwanda, and Burundi (Figure 1), covering an area of about 1,820,000 km^2 with a population of over 140 million [*UNPD*, 2010]. This region of sub-Sahara African hosts the most vulnerable populations to water stress [*Rekacewicz*, 2009], an alarming fact given that water scarcity is projected to increase [*Meigh et al*., 1999] due to climate change [*UNEP*, 2008], population growth and development [*Falkenmark*, 1990, 1997; *Fischer and Heilig*, 1997; *Wallace*, 2000]. East Africa encompasses a number of surface lakes (Figure 1) that significantly contribute to the productivity of the area by supporting a number of activities including fishing, agriculture, transport, and hydropower generation [*Molsa et al*., 1999; *Ntiba et al*., 2001; *Awange et al*., 2007; *UNEP*, 2008; *Swallow et al*., 2009; *Maitima et al*., 2010]. Lake Victoria, the largest regional lake, is known for generating great amounts of hydro-electricity for the entire region [*Awange and Ong'ang'a*, 2006], supplying over 30% of food to communities in its basin through fishing [*Song et al*., 2004] and supporting the livelihoods of over 70% of its basin communities through agriculture [*Kayombo and Jorgensen*, 2005]. Many regional surface lakes are undergoing both climate and human induced degradation [*Verschuren et al*., 2002; *Leiju*, 2012], posing a threat to the region's productivity.

East Africa is primarily underlain by shallow, low-permeable, and low-productive crystalline basement aquifers comprised of zones of weathered overburden (regolith), fractured rocks, and unweathered bedrocks [*Clark*, 1985; *Jones*, 1985; *Wright and Burgess*, 1992; *Chilton and Foster*, 1995; *Taylor and Howard*, 1996; *MacDonald and Davies*, 2000], with reported yields between 0.1 and 100 m^3/h [*Kashaigili*, 2010; *Pavelic et al*., 2012]. Groundwater within the weathered zones and bedrock fractures [*MacDonald and Davis*, 2000] is recharged by direct infiltration from precipitation and through preferential pathways [*Wright and Burgess*, 1992; *Aldous*, 2005; *Pavelic et al*., 2012]. Nonhomogeneous soil types cover the region, particularly dominated by ferrasols, acrisols, and nitisols (Uganda), cambisols and acrisols (Tanzania), and solonetz and fluvisols along river systems (Kenya) [*Dewitte et al*., 2013].

The average rainfall across the study area varies from >2000 mm/yr around Lake Victoria and the mountainous areas to <200 mm/yr in the arid and semiarid lands [*Nicholson*, 1996]. Rainfall is controlled by a number of atmospheric phenomena including the Inter Tropical Convergence Zone (ITCZ), subtropical anticyclones, El Niño Southern Oscillation (ENSO), monsoonal winds, Indian Ocean Dipole (IOD), Congo air mass among other factors [*Ogallo*, 1988; *Janowiak*, 1988; *Basalirwa*, 1995; *Nicholson*, 1996; *Saji et al*., 1999; *Indeje et al*., 2000; *Black et al*., 2003; *Behera et al*., 2005; *Marchant et al*., 2007; *Tierney et al*., 2013]. The region typically experiences bimodal rainfall [*Nicholson*, 1996; *Conway et al*., 2005; *Nikulin et al*., 2012], with peaks around April and November. East Africa's rainfall has been shown to portray decadal climate variability [*Omondi et al*., 2012, 2013; *Yang et al*., 2014] with an intensifying dipole (positive anomalous rainfall in the north and opposite conditions in the south) associated with warming climates [*Schreck and Semazzi*, 2004]. Warm ENSO (El Nino) events are associated with abundant rains over the region and the conditions are reversed during cold ENSO (La Nina) episodes [*Janowiak*, 1988; *Ogallo*, 1988; *Nicholson*, 1996; *Schreck and Semazzi*, 2004]. The IOD is associated with excess rainfall especially during the short rainfall season of October–December [*Black et al*., 2003; *Clark et al*., 2003; *Bergonzini et al*., 2004; *Behera et al*., 2005; *Hastenrath*, 2012]. Both ENSO and IOD are associated with significant influence on water storage variability in the region [*Awange et al*., 2013, 2014a; *Ahmed et al*., 2014].

Groundwater recharge is a function of rainfall and over parts of basement aquifers in East Africa, significant recharge has been shown to follow heavy rainfall events [*Owor et al*., 2009; *Bonsor and Macdonald*, 2010; *Taylor et al*., 2013]. Groundwater recharge over the region is also spatially distributed with wet and humid areas receiving more recharge than the semiarid regions [*Gavigan et al*., 2009]. If global model predictions of rainfall are accurate, groundwater recharge and storage in East Africa may increase in the future [*Gavigan et al*., 2009] as global climate models predict an increase in mean precipitation rates and intensity over East Africa during the 21st century [*IPCC*, 2007; *Solomon*, 2007; *Shongwe et al*., 2011]. However, little is known about current groundwater use rates and storage capacity. The increasing demand and use of groundwater from growing populations [*Carter and Parker*, 2009] could potentially offset the impacts of increasing recharge thus leading to declines in storage as already reported in some parts of this region [*Mogaka*, 2006; *Ministry of Water and Enviornment (MWE)*, 2007; *Tenywa*, 2007; *Wanzala*, 2013; *Lwanga*, 2015]. Such declines have great social and economic implications since groundwater is the major supplier of drinking water for rural communities [*Tindimugaya*, 2008] and municipal and industrial water [*Wijnen et al*., 2012] for urban centers such as Nairobi, Dar es Salaam, Arusha, Kampala, Dodoma, and Nakuru [*Mogaka*, 2006; *Adelana et al*., 2009; *Kashaigili*, 2010].

A complex geologic history of the region is apparent in Figure 1 whereby the East African Rift has limited hydrologic connectivity to regional basins extracted from the Hydrological data and maps based on Shuttle Elevation Derivatives at multiple Scales (HydroSHEDS) [*Lehner et al*., 2008] data set (http://hydrosheds.cr.usgs.gov/index.php, accessed on 11/13/2012). Our analysis of the groundwater behavior using GRACE was done at a hydrologic subbasin scale and assumes that groundwater boundaries are coincident with surface water boundaries. The study area was divided into five subbasins A, B, C, D, and E representing the Upper Nile, Shebelli-Juba, Upper East Central Coast, Congo, and Lower East Central Coast basins, respectively (Figure 1), to permit evaluation of GRACE TWSA. Hydromorphological features including precipitation [*UNEP*, 2008; *Omondi et al*., 2012] and geologic characteristics [*Foster et al*., 2008] were assessed to combine rift valley basins with neighboring basins. GRACE groundwater storage anomalies were estimated for the five subbasins with area

Table 1. Subbasins and Their Respective Areas

Subbasin	Size (km^2)
A	494,280
B	456,460
C	308,710
D	393,760
E	428,570

ranges of 300,000–494,280 km^2 (Table 1), larger than spatial requirements of 200,000 km^2 deemed suitable for GRACE analysis [*Rodell and Famiglietti*, 1999].

3. Data Sets and Methods

3.1. Surface Water Storage and Soil Moisture

Time series of lake heights (for the first six lakes in Table 2), volume (for Lake Edward), and lake areas measured by satellite altimeters were obtained from the Hydroweb-GOHS (Geodesy, Oceanography, and Hydrology from Space) website (http://www.legos.obs-mip.fr/soa/hydrologie/hydroweb/). The Hydroweb-GOHS data are a merged product of measurements from the Topex/Poseidon, Jason-1, ENVISAT, and GFO satellites [*Crétaux et al*., 2011]. The temporal resolution of the satellites is 10 days for Topex/Poseidon and Jason-1, 17 days for GFO, and 35 days for ENVISAT [*Birkett et al*., 2011; *Crétaux et al*., 2011]. This means that for a given lake, sampling frequencies vary depending on the satellite used. The data samples are also at higher temporal resolutions for months when two satellite missions overlap. We resampled the lake height altimetry data at the same dates as the GRACE TWSA data after applying a low-pass filter (i.e., a Finite Impulse Response (FIR) filter whose order corresponded to the minimum sampling rate in days for each Lake height satellite data set) to remove high frequencies. Validation of altimetry data over in situ lake height observations for a number of global lakes including Lake Victoria has revealed successful results [*Birkett et al*., 2011] thus providing a reliable data set for use in hydrologic studies. The lakes (Table 2 and Figures 1 and 2) used in the analysis included those within and outside the study area for which we could obtain a minimum of a 1° × 1° grid mask. For all the lakes but Edward, the lake height was converted to anomalies by deducting each month's data from a long term mean (i.e., January 2004 to December 2009, similar to the GRACE product), multiplying by the reference lake area and dividing by the lake mask area to conserve mass. By doing so, we assume a constant lake area through time. Volumes of storage for Lake Edward were divided by the lake's mask area to estimate water equivalent depth anomalies. Smaller lakes data from the Hydroweb-GOHS were available for the period 2002–2010, and had not been updated by the time of this study; we thus focused on the period 2003–2010.

In situ soil moisture (SM) monitoring is limited over many parts of the world including the East African region. This study, similar to other GRACE groundwater estimation studies [*Rodell et al*., 2006, 2009; *Swenson et al*., 2006; *Strassberg et al*., 2007; *Syed et al*., 2008; *Moiwo et al*., 2012; *Shamsudduha et al*., 2012; *Feng et al*., 2013; *Voss et al*., 2013; *Joodaki et al*., 2014], utilizes gridded soil moisture output from global land surface models (LSMs). Monthly soil moisture estimates were extracted from the CLM (v.4) model (1° × 1° spatial resolution, 10 soil layers, 3.4 m total soil depth) [*Oleson et al*., 2010], the WaterGAP (WG) version 2.1 model (0.5° × 0.5° spatial resolution) [*Alcamo et al*., 2003; *Döll et al*., 2003], and three 1° × 1° spatial resolution models from the Global Land Data Assimilation System (GLDAS), i.e., NOAH (four soil layers, 2.0 m total soil depth), VIC (three soil layers, 1.9 m total soil depth), and MOSAIC (MOS) (three soil layers, 3.5 m total soil depth) [*Rodell et al*., 2004a] accessible via NASA's GES DISC (http://disc.gsfc.nasa.gov/hydrology). GLDAS soil moisture output were assumed to be the mean of the three model products, for which no uncertainty was assumed. CLM (v.4) includes a rudimentary groundwater scheme which represents an unconfined aquifer hydraulically connected to soil layers [*Oleson et al*., 2010]. The WaterGAP model, unlike all the other models used in the study, is calibrated against measured global discharge [*Döll et al*., 2003]. Soil moisture data from WaterGAP were spatially interpolated to a 1° × 1° resolution to match the other model data sets and GRACE.

Table 2. East African Lakes Used in the Study

Lake	Area (km^2)	Standard Deviation (2003–2010) (km^3)
Victoria	68,800	21.06
Tanganyika	32,000	10.18
Malawi	29,600	12.53
Turkana	6,750	4.72
Albert	5,300	3.16
Mweru	5,120	2.93
Edward	2,325	0.38

Soil moisture for each model was combined with lake height from altimetry to produce a combined product of surface and subsurface storage. The lake altimetry data helped compensate for the poor performance of the land surface models over lakes by providing a more realistic estimate of lake storage. The lake height data were first projected to a 1° × 1° global grid before being added to the 1° × 1° soil moisture grid. The combined soil moisture and lake

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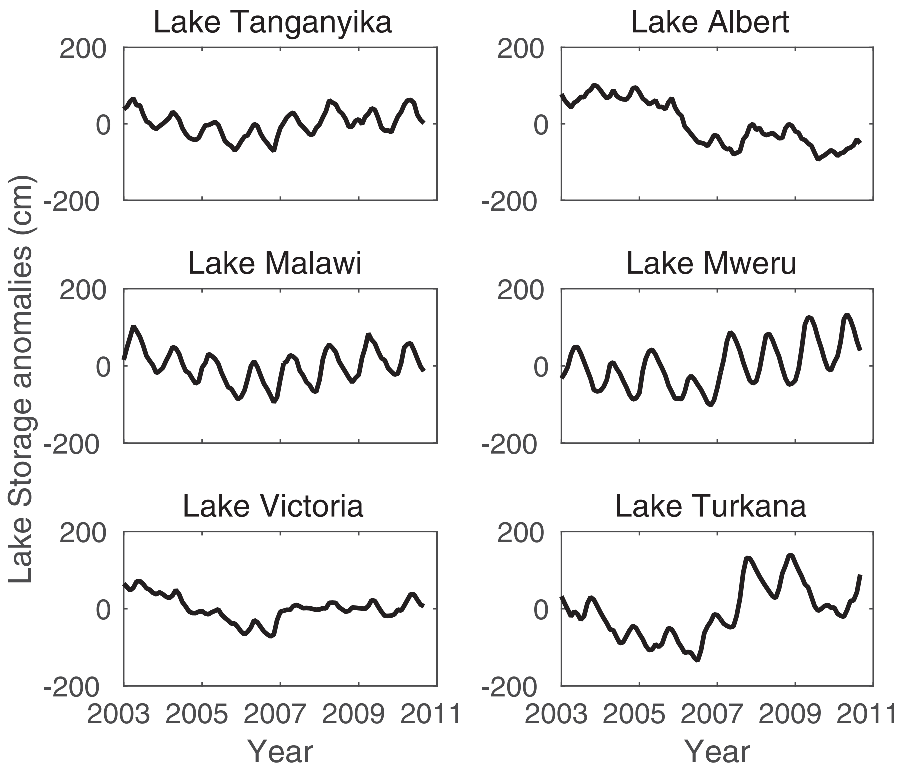

Figure 2. Storage variations for the lakes used in the study.

altimetry product was then processed similarly (i.e., truncated at harmonic degree 60, destriped using a decorrelation filter following *Swenson and Wahr* [2006], and smoothed with a 200 km radius Gaussian filter) to the GRACE postprocessed products we used (as highlighted in section 3.3) to produce a filtered product hereafter referred to as SM_{model} + ALT (for model = CLM, WG, and GLDAS). Processing storage changes using true lake locations allows for more accurate spatial distribution of water mass to correctly remove the influence of lake storage in the GRACE TWSA signal, which if not well captured may lead to a bias in the groundwater estimates [*Longuevergne et al*., 2013]. The error in average SM_{model} + ALT was taken as the standard deviation for the three model data sets.

3.2. Tropical Rainfall Measuring Mission (TRMM) Precipitation Product

The study uses the 3B43 version 7 monthly gridded (0.25° × 0.25° spatial resolution) rainfall estimates (released as 3 hourly precipitation rates) sourced from the TRMM Multisatellite Precipitation Analysis product [*Huffman et al*., 2007] available at the NASA website (http://trmm.gsfc.nasa.gov/data_dir/data.html). It is a merged product of data from different satellite sensors (e.g., precipitation radar and Special Sensor Microwave Imager) and rain-gauge observations from the Global Precipitation Climatology Center. The TRRM precipitation rates (in cm/d) were converted to a cumulative monthly product and Gaussian smoothed with a 200 km smoothing radius similar to GRACE, for use in this study. Several studies have used TRMM over different parts of Africa [*Awange et al*., 2007, 2013, 2014a, 2014b, 2016; *Ahmed et al*., 2014; *Dinku et al*., 2008; *Forootan et al*., 2014; *Naumann et al*., 2012] and it has been shown to provide the best spatial and temporal rainfall distribution [*Beighley et al*., 2011], rainfall intensity and frequency [*Sylla et al*., 2013] and closest agreement when compared with in situ rain gauge observations [*Adeyewa and Nakamura*, 2003], and GPCC precipitation product [*Awange et al*., 2016]. In this study, TRMM is used in the water balance evaluation (section 3.3) and to evaluate the estimated GRACE groundwater storage behaviors relative to precipitation changes.

3.3. GRACE Total Water Storage Anomalies

The GRACE satellite provides monthly estimates of global gravity fields since 2002 which, once processed, produce vertically integrated Total Water Storage Anomalies (TWSA) [*Syed et al*., 2008] that can be used to evaluate the behavior of a range of hydrologic variables including lake and reservoir storage [*Awange et al*., 2007; *Swenson and Wahr*, 2009; *Becker et al*., 2010; *Wang et al*., 2011; *Hassan and Jin*, 2014] and groundwater storage [*Rodell and Famiglietti*, 2002; *Rodell et al*., 2006, 2009; *Yeh et al*., 2006; *Famiglietti et al*., 2011; *Voss et al*., 2013; *Castle et al*., 2014]. We use release-5 [*Swenson*, 2012], $1° \times 1°$ gridded GRACE TWSA data produced by the Center for Space Research (CSR) at the University of Texas and provided by the Jet Propulsion Laboratory (JPL) through NASA's TELLUS website (http://grace.jpl.nasa.gov/). This GRACE data set corresponds to monthly spherical harmonic coefficients that have been postprocessed (i.e., truncated at harmonic degree 60, destriped using a decorrelation filter following *Swenson and Wahr* [2006], and smoothed with a 200 km radius Gaussian filter) [*Landerer and Swenson*, 2012] to remove systematic and random errors [*Swenson and Wahr*, 2006].

Due diligence of using the GRACE TWSA data over the region necessitates a water balance analysis to assess the accuracy of the GRACE data to capture basin total water storage. Monthly changes in total water storage from GRACE dTWS′/dt (the time derivative of TWSA, generated using a forward difference derivative of TWSA), were compared in a water balance analysis with monthly water storage variations (dS/dt) estimated from observed precipitation (P) from the Tropical Rainfall Monitoring Mission (TRMM) product (i.e., the 3B43 monthly product, section 3.2), satellite-model-observation-based evapotranspiration (ET) data [*Mueller et al*., 2013] and streamflow (Q) measurements at the outlet of the Upper Nile basin (red dot, Figure 1) according to equation (2)

$$\frac{dS}{dt} = P - ET - Q. \tag{2}$$

The long-term mean similar to the GRACE product (i.e., January 2004 to December 2009) was removed from the three products P, ET, and Q. It is worth noting that quantifying errors in satellite precipitation products is challenging given poor distribution of rain-gauge networks [*Awange et al*., 2016] across regions. For this analysis, the error in TRMM precipitation is taken as 0.3 mm/d [*Pipunic et al*., 2013]. The error value on P considered here lies within the bias and RMSE estimates generated when TRMM is compared to GPCC [*Awange et al*., 2016]. The error on ET, reported as a standard deviation value of ~150 mm/yr (equivalent to ~4.33 cm/month) [*Mueller et al*., 2013], is used in this study. The ET data set is a merged synthesis product based on evapotranspiration estimates from satellite and/or in situ observations and LSMs outputs [*Mueller et al*., 2013], thus the error on E, combines uncertainties from the different data sets. A relative error of 5% was considered for Q [*Rodell et al*., 2004b; *Famiglietti et al*., 2011; *Long et al*., 2014], while errors on GRACE TWSA were computed as basin averages of the measurement errors [*Swenson and Wahr*, 2006; *Landerer and Swenson*, 2012] provided as a gridded product at NASA's TELLUS website. The GRACE gridded measurement errors are spatially correlated [*Landerer and Swenson*, 2012] for nearby grid cells and therefore to derive a representative basin-wide total measurement error from GRACE, we estimate the error covariance following methods in [*Landerer and Swenson*, 2012].

3.3.1. Estimating Groundwater Storage Variations From GRACE

The filtered SM_{model} + ALT storage anomalies (section 3.1) were subtracted from the filtered GRACE storage anomalies to estimate filtered groundwater anomalies (GWA) using equation (1). Water storage from snow and canopy were ignored in equation (1) as these contribute little (<1%) to the TWSA and are accounted for in model residuals. Scaling of the resulting groundwater residuals would still be required, especially if the region under study consists of many distributed local aquifers. However, a poor knowledge of aquifer spatial distribution may make the estimate of such a scaling factor highly uncertain. We correct for bias in the groundwater residual amplitudes using the ratio (estimated for this study as ~1.16) between filtered and unfiltered groundwater storage from the WaterGAP model [*Döll et al*., 2003] similar to methods in *Longuevergne et al*. [2013]. We assume that the study area is underlain by a homogeneous basement rock aquifer system [*Foster et al*., 2008; *Adelana et al*., 2009; *MacDonald et al*., 2012] and leakage from aquifers in neighboring subbasins is minimal. The errors in the groundwater estimate are propagated from the GRACE measurement errors provided at the TELLUS website and SM + ALT error terms.

3.4. In Situ Groundwater Data From Monitoring Wells

In situ groundwater monitoring and measurement over East Africa has been minimal in the past and, where data do exist, policies for data sharing limit access to groundwater level observations. To assess GRACE groundwater estimates, we used monthly groundwater records from a network of 29 monitoring wells over

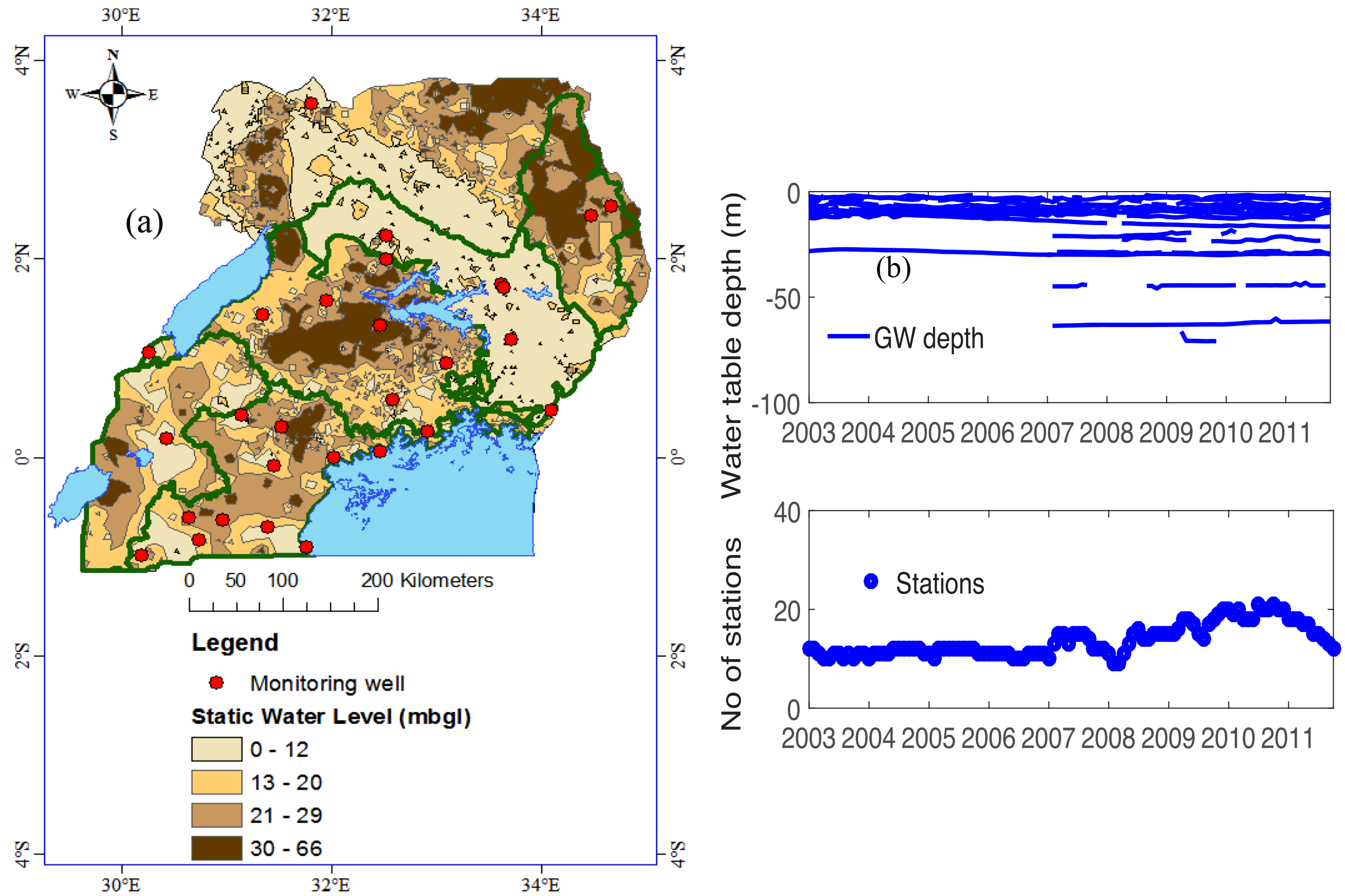

Figure 3. Spatial distribution of depth to static water level over Uganda, groundwater monitoring well locations (red dots), and subbasin parts of the Upper Nile basin in Uganda (green) (a) and the groundwater monitoring network since 2003 (b).

Uganda (Figure 3a), available in form of depths to the water table (Figure 3b). Depths to the water table measurements were converted to water table elevations using topographic values extracted from a GTOPO30 global digital elevation model (DEM), with a horizontal grid spacing of 30 arc sec (https://lta.cr.usgs.gov/GTOPO30). The data set was subjected to quality control to remove ambiguous values, especially those that were deemed to result from active groundwater pumping. Gaps in the groundwater elevation time series were filled using spatial-temporal interpolation approaches [*Zeng and Levy*, 1995, equations (1) and (2)], shown as equations (3) and (4) below, with a 1 month time step T and a 200 km (similar to the radius used for the GRACE Gaussian smoothing) spatial range D. The variables x_k, y_k, and t_k (x_0, y_0, and t_0) are the location and time of the nonmissing (missing) groundwater observation gw_k, (gw_{est}) within the distance D and time range T, while w_k is the weight assigned to each nonmissing value based on its temporal and spatial gap from the missing value. The interpolator uses both temporal and spatial information and substitutes spatial information where temporal information is missing and vice versa.

$$gw_{est}=\frac{\sum_{k=1}^{n} w_k gw_k}{\sum_{k=1}^{n} w_k}, \tag{3}$$

where

$$w_k=\frac{2-\left[\frac{(x_k-x_o)^2+(y_k-y_o)^2}{D^2}+\frac{(t_k-t_o)^2}{T^2}\right]}{2+\left[\frac{(x_k-x_o)^2+(y_k-y_o)^2}{D^2}+\frac{(t_k-t_o)^2}{T^2}\right]}. \tag{4}$$

The groundwater elevations were further converted to anomalies by subtracting a long-term mean. Our initial analysis involved computing an anomaly based on the 2004–2009 mean similar to the GRACE data.

However, the under sampling of in situ observations during the 2003–2007 (Figure 3b) period affected the long-term mean and the results. Therefore to minimize large uncertainties due to under sampling, groundwater anomalies (GWLA) were generated from the 2007–2010 mean. The GWLA were then multiplied by specific yield values to estimate groundwater storage anomalies (GWSA). The static water level map of Uganda (Figure 3a) shows distinct characteristics of the underlying aquifers of Uganda, with shallow ($<$12 m below ground level (mbgl)) unconfined aquifers in the middle of the country stretching from southwestern to northeastern Uganda and relatively deep aquifers ($>$13 mbgl) covering the rest of the country. A specific yield (S_y) of 0.11 was applied for the shallow aquifer region (i.e., with static level depths 0–12 mbgl, Figure 3a) while $S_y = 0.004$ was used to represent storativity in the relatively deeper groundwater regions (i.e., with static level depths 13–66 mbgl, Figure 3a). These values are within reported specific yield or storativity ranges of 0.1–0.23 for unconfined [*Taylor and Howard*, 2000; *Tindimugaya*, 2008; *Pavelic et al*., 2012] and 0.001–0.02 for fractured and crystalline basement rock aquifers [*Chilton and Foster*, 1995; *Healy and Cook*, 2002; *Nyagwambo*, 2006; *Taylor et al*., 2010a] in East Africa. An area-weighted average monthly time series of groundwater storage anomalies was estimated for the subbasins delineated in Figure 3a. This data, hereafter referred to as $GWSA_{OBS}$, is compared with groundwater storage changes as estimated from GRACE (averaged over areas in Figure 3a) using goodness of fit criteria including Spearman's correlation coefficients (ρ) and root mean square error (RMSE), for purposes of evaluating the satellite's capacity to monitor regional groundwater storage behaviors. It is worth noting that using a constant specific yield across parts of the study area that is characterized by mixed hydrogeologic conditions (unconfined, confined, bedrock aquifers) potentially introduces uncertainties in the in situ groundwater results [*Rodell et al*., 2006; *Shamsudduha et al*., 2012]. Spatially distributed constant specific yield values have been shown to impact the magnitude of groundwater depletion (e.g., in Bengal basin [*Shamsudduha et al*., 2012]) as compared to using a single value, although the direction of such an impact is nontrivial [*Taylor et al*., 2010b].

4. Results and Discussion

4.1. Comparison of GRACE's Changes in Total Water Storage With Observations

Over the Upper Nile basin (subbasin A, Figure 1), the results of the water balance comparison for the period 2003–2006 (Figure 4b) show that the observed water balance and the GRACE time derivative are fairly well correlated ($\rho = 0.6$ and RMSE = 3.3 cm), thus illustrating the potential use of GRACE to estimate individual variables of the water balance including groundwater storage variations. The time series in Figure 4a shows similar semiannual variations in both precipitation and total water storage (with peaks in April and November) (Figure 4b), an indication that changes in total water storage (dTWS′/dt) are related to changes in precipitation (P) (i.e., $\rho = 0.63$). The contribution of P and E to the average subbasin changes in total water storage over the 2003–2006 period is about 58% and 37.5%, respectively. Evapotranspiration shows less interannual variability and remains relatively stable throughout the period of analysis. Subbasin discharge is 1 order of magnitude less than P and E and contributes much less ($\sim$4.5%) to the observed changes in total water storage. Discharge (Q) is generally reduced at the beginning of the year and elevated at the end, a behavior that depicts management practices. The interannual variations in Q show that the period 2003–2005 registered higher Q values compared to 2005–2008 (Figure 4a). These high Q values are a manifestation of the reported increase in discharge/outflows from upstream Lake Victoria during the period 2001–2005 [*Kull*, 2006; *Mubiru*, 2006; *Sutcliffe and Petersen*, 2007], which also led to the drop in lake's water levels [*Awange et al*., 2007; *Swenson and Wahr*, 2009] observed between 2003 and 2007 (Figure 2).

Our correlation result (i.e., $\rho = 0.6$) between the water balance (P-ET-Q) and GRACE's dTWS′/dt is slightly lower that values reported elsewhere [e.g., *Famiglietti et al*., 2011; *Long et al*., 2014, 2015]. This could be attributed to uncertainties in the data sets used. The study is over the equatorial region where measurement errors in GRACE are higher [*Landerer and Swenson*, 2012], given the satellite's intrinsic low resolution (about 400 km at the equator). Uncertainties in the merged ET product used in this study [*Mueller et al*., 2013] that result from the different data sets (i.e., satellite, in situ, model, and reanalysis) and their associated derivation assumptions [*Long et al*., 2014] could also affect the results. For instance, remotely sensed ET products are shown to be less responsive to climate variations (compared to LSM-based ET products) [*Long et al*., 2014, 2015], and thus fail to capture soil moisture constraints on evapotranspiration [*Long et al*., 2014, 2015; *Yang et al*., 2015], resulting in dampened seasonal cycles [*Long et al*., 2014] and unreasonable spatial

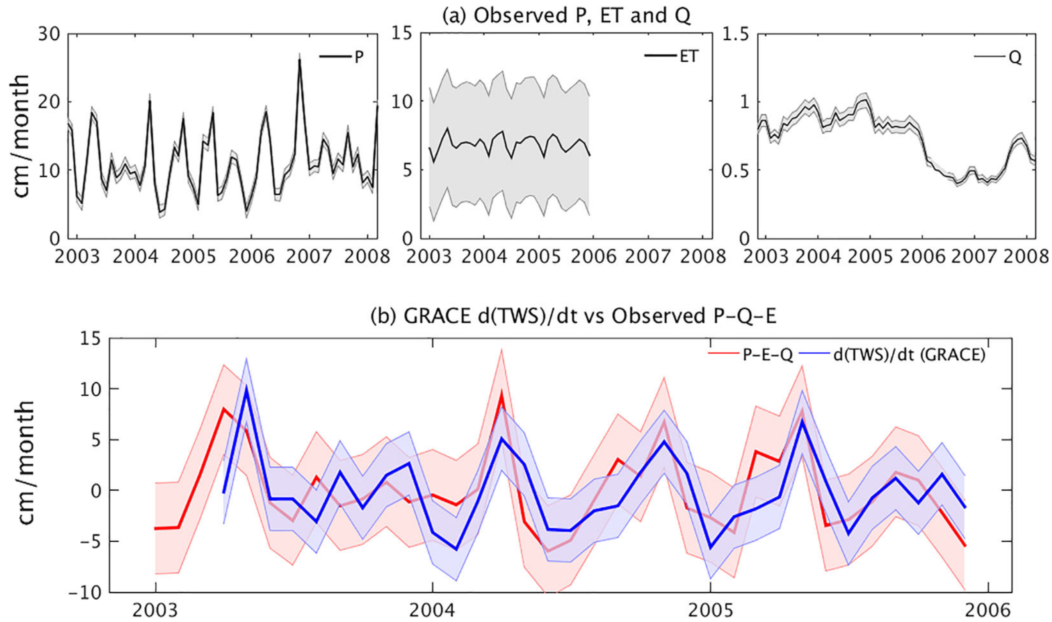

Figure 4. Precipitation, evapotranspiration, and discharge time series (a) and the Upper Nile basin water balance based on GRACE and observations (b).

patterns in ET [*Yang et al*., 2015]. Such uncertainties are transferable to the water balance (P-ET-Q) estimates, affecting their correlation with GRACE's dTWS/dt [e.g., *Long et al*., 2014].

4.2. Comparison of GRACE Groundwater Estimates With In Situ Observations

Results from the comparison of monthly groundwater observations from monitoring wells (Figure 3) with those estimated from filtered GRACE (Figure 5b) are promising, given that we compared low spatial resolution but homogeneous (in space and time) GRACE data with high-resolution but scattered (in space and time) monitoring wells data. The two time series portray a similar annual signal that generally peak in November/December, although the seasonal amplitudes differ in most of the months, with the GRACE estimates having larger peaks/amplitudes than the in situ data. A correlation coefficient $\rho = 0.41$ and a RMSE = 4.36 cm are estimated between the two data sets for the entire period 2003–2010. Considering piecewise relationships, a better agreement between GRACE groundwater estimates and in situ observations from monitoring wells is apparent during the period 2007–2010 ($\rho = 0.6$ and RMSE = 3.1 cm), when compared to the period 2003–2006 ($\rho = 0.4$ and RMSE = 5.3 cm). These results show that GRACE can accurately observe changes in groundwater storage and thus can be used to monitor groundwater changes over the poorly sampled aquifers in East Africa. The improved relationship during 2007–2010 is likely attributed to an increase in groundwater sampling including deeper parts of the aquifer (Figure 3b). In a similar study, *Feng et al*. [2013] show that for a region where the contribution of deep aquifer groundwater depletion is considered, GRACE groundwater depletion rates were consistent with those from in situ observations. Such results are intuitive since GRACE monitors changes in the entire groundwater resource (from top to bottom) and will agree best compared with a representative groundwater data set when those observations are well sampled in space and depth.

An evaluation of the GRACE groundwater estimation approach used in this study (i.e., filtered data sets in equation (1)) against common/regular approaches (i.e., GRACE data scaled using model outputs [e.g., *Landerer and Swenson*, 2012]) show that the former approach (Figure 5a) produces better groundwater estimates than the latter (GRACE data scaled using CLM4-based gain factors provided at http://disc.gsfc.nasa.gov/hydrology (Figure 5b)), relative to in situ groundwater observations, particularly after 2007, the time period during which observations are most representative. The 2007–2010 statistics for the regular scaling approach are 0.2 and 3.5 cm for the correlation coefficient and RMSE, respectively. Thus, the approach used in this study improves the relationship between GRACE groundwater estimates and in situ observations by over 75% and reduces the root mean square error by ~14%. The differences between the two methods are attributable to a combined effect of unresolved leakage from changes in the lakes within the basin that

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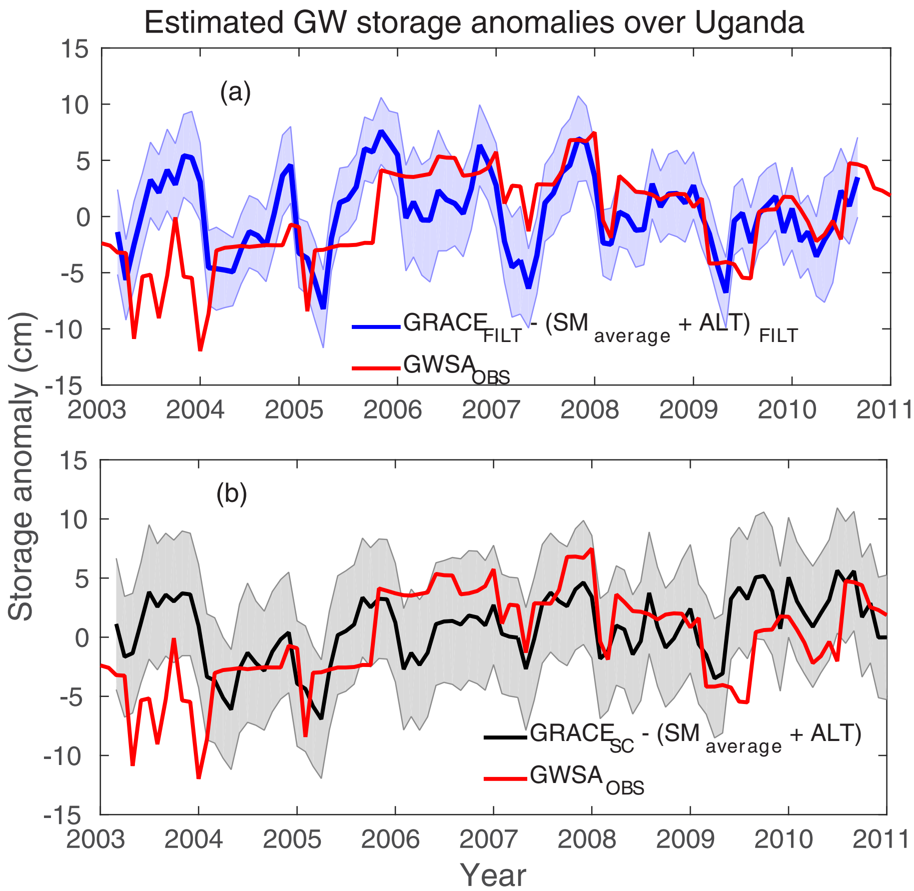

Figure 5. Comparison of GRACE groundwater estimates based on the approach in this study (5a) and on conventional scaling methods (based on scaling factors provided at the TELLUS website) (5b) with in situ observations over Uganda.

result from limitations in model-based GRACE scaling approaches (especially over areas with large and highly variable surface water storage) and to soil moisture uncertainties. It is also worth noting that differences in the removal of the mean in the generation of anomalies (e.g., 2004–2009 mean was removed from GRACE and SM + ALT but not in situ groundwater estimates as highlighted in section 3.4) may affect specific time periods, further creating uncertainties in the end result.

4.3. Estimated GRACE Groundwater Variations Over East Africa

Figures 6–10 show plots of monthly time series for filtered GRACE TWSA (Figure 6a), SM + ALT (Figure 6b), and estimated GWSA (Figure 6c), for subbasins A, B, C, D, and E. Also included in Figure 6a for each plot is the TRMM mean monthly precipitation estimates. Figure 6d in each plot represents the monthly climatology of TWSA, SM + ALT, GWSA, and precipitation for the period March 2003 to September 2010. Table 3 and Figure 11 show trends in each storage component over the 2003–2006, 2007–2010, and the 2003–2010 periods. Trends in storage were estimated while accounting for seasonality in an error-weighted generalized linear regression model [*Barnett et al.*, 2012] adjusted with an annual wave, given error propagation in equation (1) of groundwater calculation. A Student's *t* test was applied to evaluate trend significance with $\alpha = 0.05$.

The results show that total water storage generally follows changes in precipitation, with a peak time lag of about 1 month for all subbasins (Figures 6, 10a, and 10d). Higher peaks in TWSA occur after extreme precipitation events like those experienced during the September–October–November–December (SOND) season in 2006 over the entire region (Figures 6–10a).

The interannual variations in precipitation and TWSA show that the SOND season of 2006 was the wettest within the entire 2003–2011 period of study for all the subbasins, and is attributed to a strong 2006 Indian

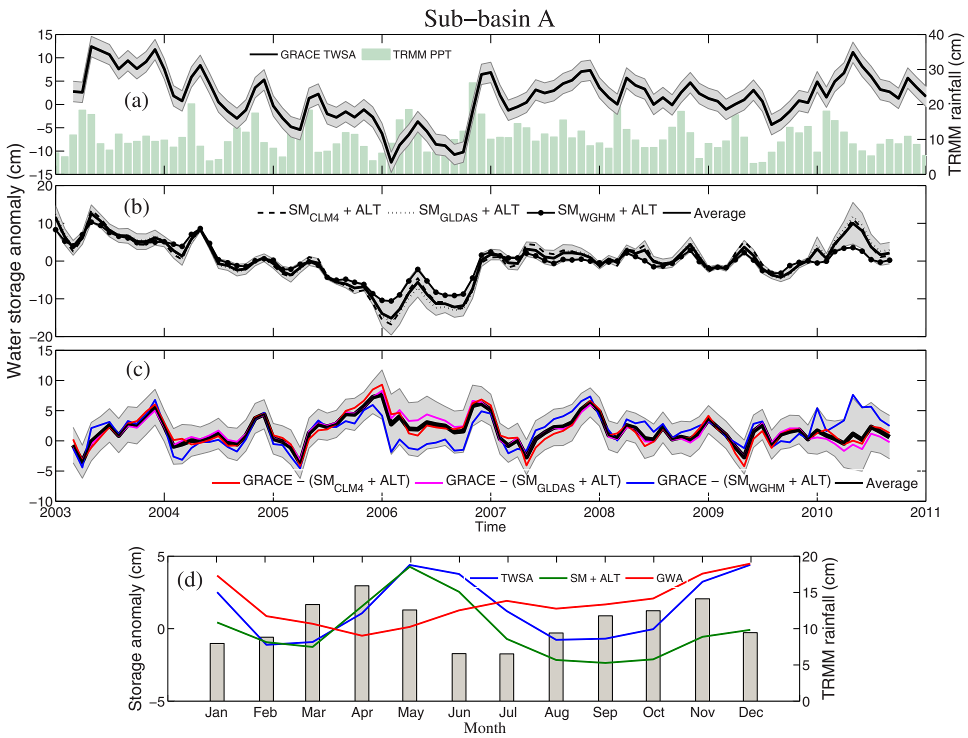

Figure 6. Plots of monthly time series for (a) GRACE Total Water Storage Anomalies (TWSA), (b) Soil Moisture and Surface Water (SM+ALT) storage anomalies, and (c) estimated Groundwater Storage Anomalies (GWSA), for subbasin A. Also included in subplot a is a bar graph of the mean monthly precipitation from the Tropical Rainfall Monitoring Mission (TRMM). Subplot d represents the monthly climatology of TWSA, SM + ALT, GWSA and TRMM for the period March 2003 to September 2010.

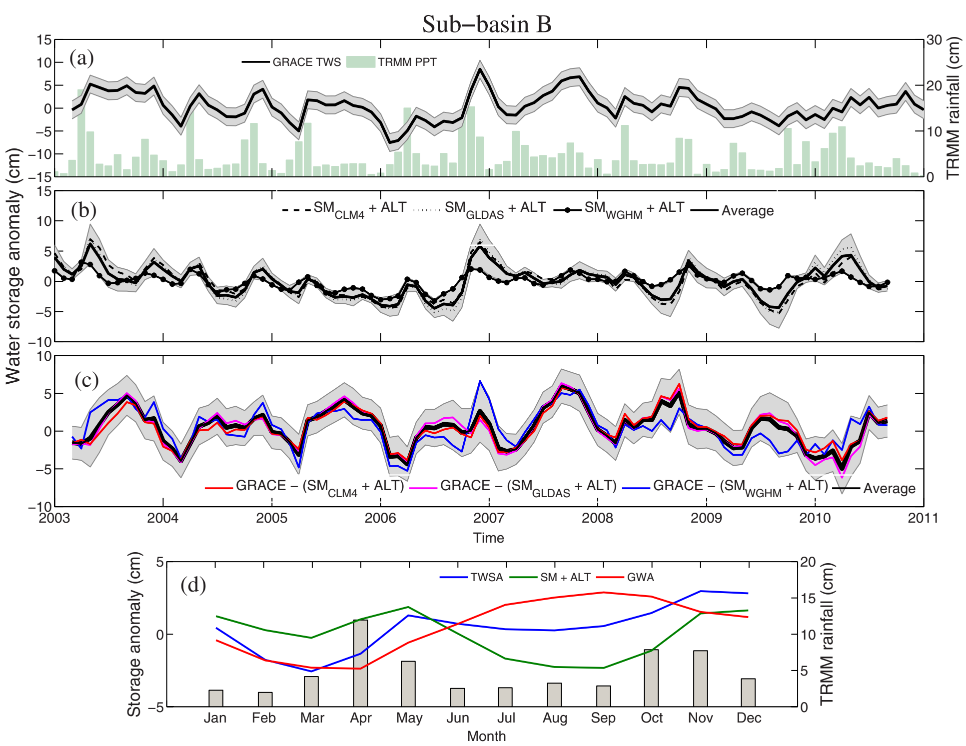

Figure 7. Same as Figure 6 for subbasin B.

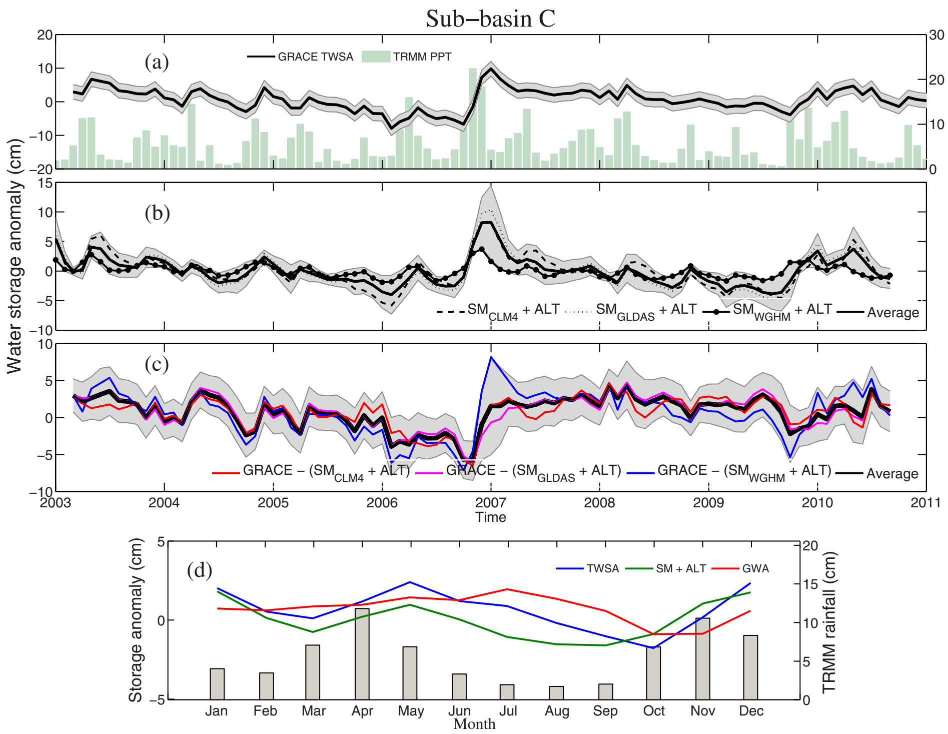

Figure 8. Same as Figure 6 for subbasin C.

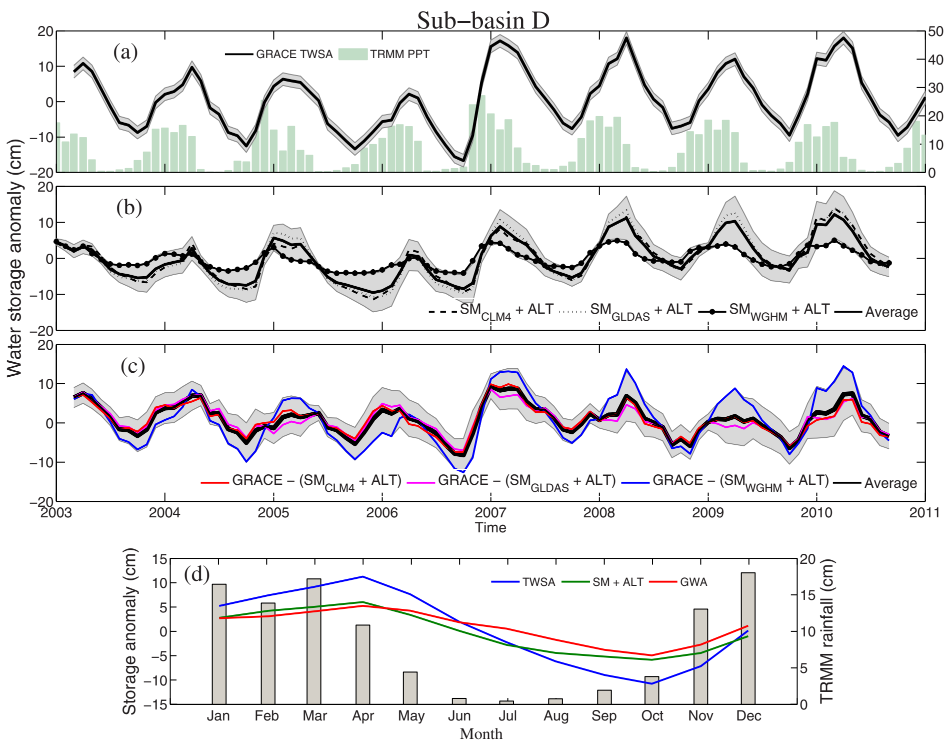

Figure 9. Same as Figure 6 for subbasin D.

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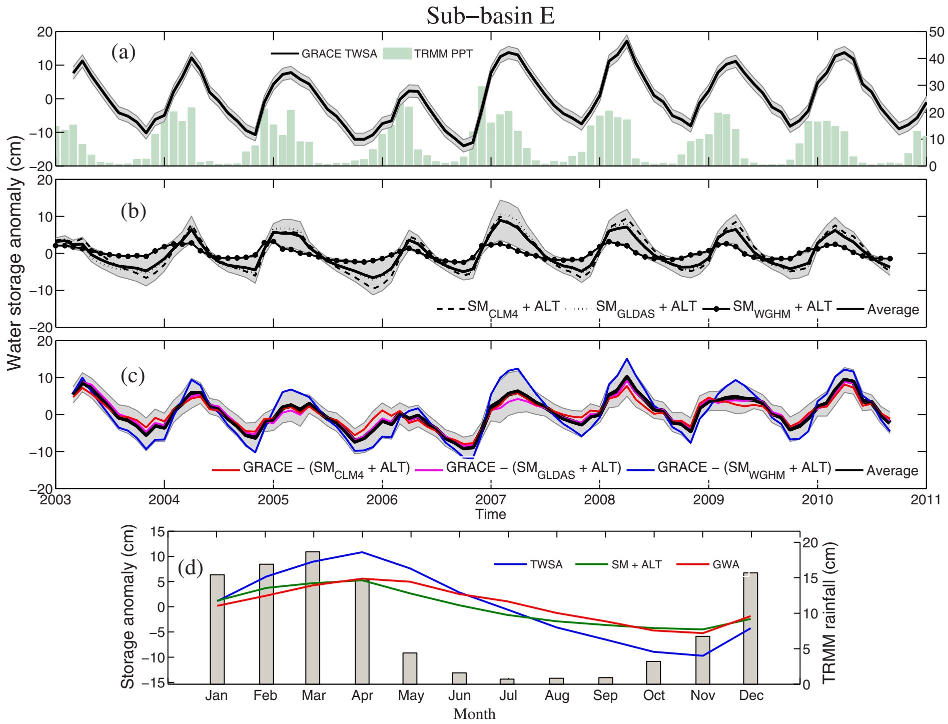

Figure 10. Same as Figure 6 for subbasin E.

Ocean dipole (IOD) forcing on the regional rainfall [*Becker et al*., 2010] as well as ENSO influences [*Awange et al*., 2014a]. In their study, *Awange et al*. [2014a] show that ENSO related rainfall increased total water storage in the Lake Victoria basin by 14.52 cm/yr during 2006–2007. Prior to this high rainfall period, a declining behavior in total water storage, that reaches a minimum in February 2006, is apparent over all the subbasins. The water storage declines that accumulated during the period 2003–2006 were counteracted by the SOND 2006 high precipitation events for all subbasins, causing a rebound in TWS (Figures 6–10a) that ultimately affected the overall TWS trend over the entire period of study. Thus, no significant trends were detected in total water storage (TWS) for the entire period 2003–2010 for all subbasins (Figures 6–10 and Table 3). However, significant piecewise negative trends in TWS were detected during the period 2003–2006 (Table 3). Subbasin A, encompassing Lake Victoria, experienced the biggest decline in total water storage (4 cm/yr (20.4 km^3), $p < 0.001$; Table 3 and Figures 6 and 11) and SM + ALT (6 cm/yr (29.3 km^3/yr), $p < 0.001$; Table 3 and Figures 6 and 11) during 2003–2006, while the other subbasins experienced total water storage declines between 1 and 1.8 cm/yr and SM + ALT declines between 0.5 and 2 cm/yr during 2003–2006. The large decline in SM + ALT over subbasin A during the period 2003–2006 is relatively higher (about 80%) than the declines experienced in the other subbasins, and largely influenced the negative TWS trend. Changes in soil moisture and lake storage are often argued to be highly related to changes in precipitation [*Findell and Eltahir*, 1997; *Sheffield and Wood*, 2008].

Our analysis of TRMM data shows no decline in precipitation (Table 3) over the entire region during the 2003–2006 period, a result consistent with other studies [e.g., *Awange et al*., 2007; *Becker et al*., 2010]. This implies that the observed negative trends in SM + ALT during 2003–2006 are unlikely to be related to changes in precipitation but should be attributed to changes in Lake Victoria storage due to human activities. In related studies [*Swenson and Wahr*, 2009; *Becker et al*., 2010; *Hassan and Jin*, 2014; *Moore and Williams*, 2014], similar results were obtained where changes in TWSA over the Lake Victoria basin during 2003–2006 were found to be highly governed by changes in the surface water stored in the lake itself, that were largely driven by management practices (i.e., increased discharge) during that period [*Kull*, 2006; *Mubiru*, 2006; *Sutcliffe and Petersen*, 2007].

Groundwater storage portrays no significant trends over the region during the entire 2003–2010 period except for subbasin D which largely covers the western parts of Tanzania that recorded a significant

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Table 3. Trends in Each Storage Component Over the 2003–2006, 2007–2010, and 2003–2010 Periods

		Period					
		2003.01–2006.12		2007.01–2010.09		2003.01–2010.09	
Subbasin	Variable	Trend (cm/yr)	p Value	Trend (cm/yr)	p Value	Trend (cm/yr)	p Value
A	GRACE	−4.12 ± 0.29[a] (−20.4 ± 1.43)	<0.001	0.39 ± 0.32	0.881	−0.07 ± 0.11	0.254
	SM + ALT	−5.92 ± 0.27[a] (−29.3 ± 1.33)	<0.001	0.55 ± 0.22	0.992	0.43 ± 0.07	1.000
	GW	0.99 ± 0.43	0.987	−0.21 ± 0.42	0.310	−0.06 ± 0.14	0.666
	TRMM	0.37		−0.61		−0.11	
B	GRACE	−1.17 ± 0.26[a] (−5.34 ± 1.18)	<0.001	−1.01 ± 0.29	0.310	−0.16 ± 0.09	0.054
	SM + ALT	−1.80 ± 0.18[a] (−8.22 ± 0.82)	<0.001	−0.46 ± 0.21[a] (−2.09 ± 0.96)	0.019	0.01 ± 0.07	0.582
	GW	−0.28 ± 0.34	0.208	−1.10 ± 0.39[a] (−5.02 ± 1.78)	0.004	−0.14 ± 0.13	0.133
	TRMM	0.05		−0.02		−0.15	
C	GRACE	−1.76 ± 0.29[a] (−5.43 ± 0.89)	<0.001	−0.76 ± 0.32[a] (−2.35 ± 0.98)	0.011	−0.02 ± 0.11	0.432
	SM + ALT	−0.61 ± 0.24[a] (−1.88 ± 0.74)	0.008	−0.12 ± 0.29	0.3456	−0.28 ± 0.08[a] (−0.86 ± 0.25)	<0.001
	GW	−1.39 ± 0.39[a] (−4.29 ± 1.20)	0.001	−0.63 ± 0.47	0.095	0.09 ± 0.14	0.736
	TRMM	0.68		−0.13		−0.04	
D	GRACE	−1.44 ± 0.24[a] (−5.67 ± 0.95)	<0.001	0.06 ± 0.26	0.095	0.99 ± 0.09	1.000
	SM + ALT	−0.74 ± 0.32[a] (−2.91 ± 1.26)	0.013	1.07 ± 0.53	0.976	1.08 ± 0.12	1.000
	GW	−0.87 ± 0.44, (−3.73 ± 1.89)	0.025	−1.49 ± 0.63[a] (−5.87 ± 2.48)	0.012	−0.37 ± 0.17[a] (−1.46 ± 0.67)	0.018
	TRMM	0.95		−0.322		0.06	
E	GRACE	−1.52 ± 0.25[a] (−6.51 ± 1.07)	<0.001	−0.30 ± 0.27	0.136	0.69 ± 0.09	1.000
	SM + ALT	0.21 ± 0.26	0.783	−0.66 ± 0.35	0.032	0.29 ± 0.09	0.999
	GW	−1.61 ± 0.38[a] (−6.89 ± 1.63)	<0.001	0.04 ± 0.48	0.536	0.33 ± 0.14	0.999
	TRMM	1.01		−0.50		0.02	

[a]Significant at 0.05%; ( ) equivalent volume lost/gained km^3/yr.

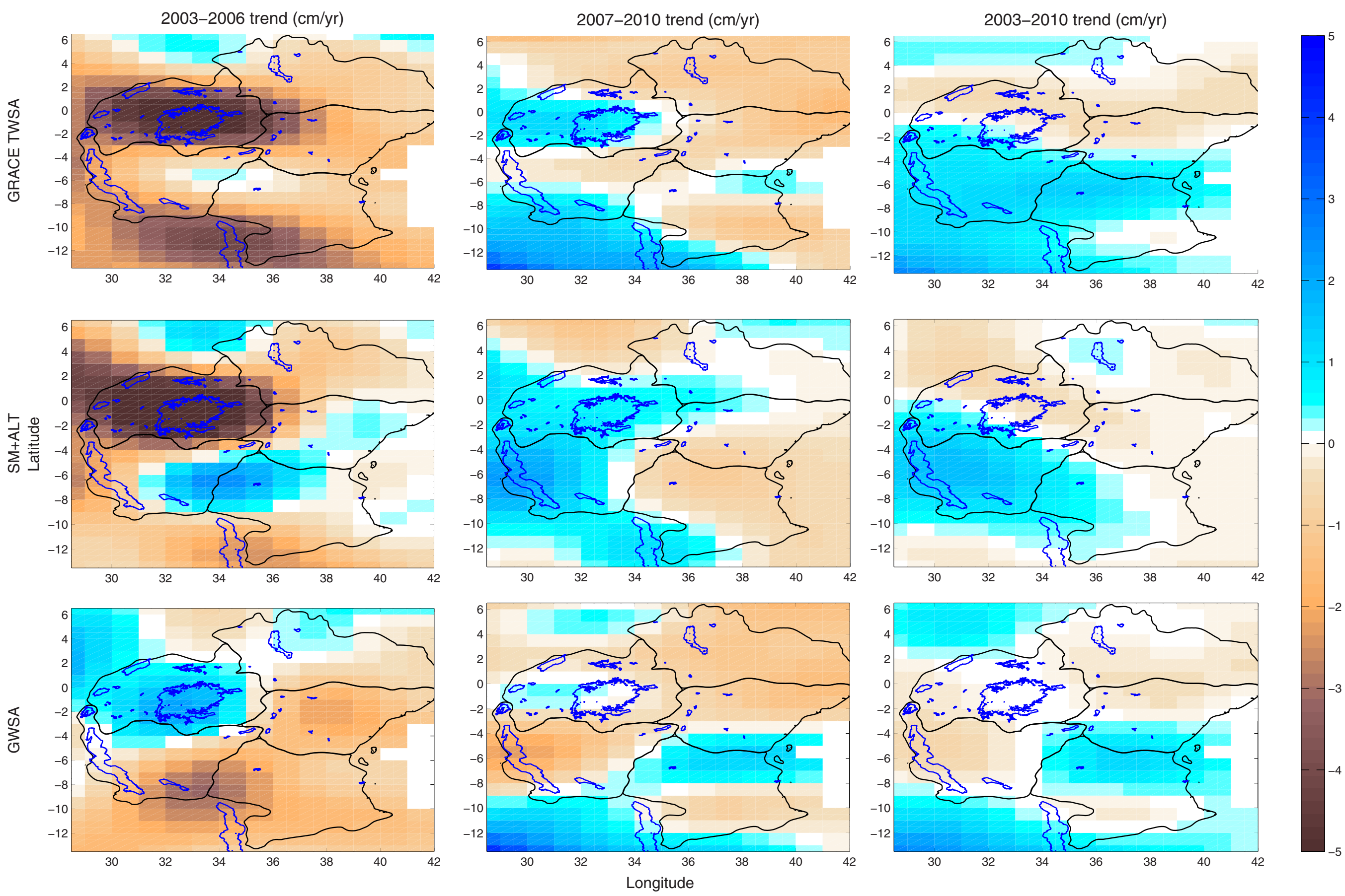

Figure 11. Spatial trends in TWSA, SM + ALT and GWSA over East Africa.

negative trend (-0.37 cm/yr (-1.46 km^3/yr), $p = 0.018$) (Figure 9 and Table 3). However, during 2003–2006, significant negative trends in groundwater storage are observed in subbasins C (-1.4 cm/yr (-4.3 km^3/yr), $p = 0.001$), D (-0.87 cm/yr (-3.73 km^3/yr), $p = 0.025$), and E (-1.6 cm/yr (-6.88 km^3/yr), $p < 0.001$) and each of these trends is largely correlated to the observed negative trend in TWSA in the respective subbasins. The behavior of TWS, GW, as well as SM + ALT, closely follow rainfall patterns for these three subbasins (Figures 8d, 9d, and 10d), an indication that the underlying aquifers are shallow and geologic characteristics permit rapid infiltration. These three subbasins (i.e., C, D, and E) have limited surface water resources thus likely rely on groundwater resources to meet water demands. We attribute the 2003–2006 negative changes in groundwater storage over subbasin C (covering South-Kenya), D, and E (covering parts of Western Tanzania and Southern Tanzania, respectively) to groundwater use rather than climate variability (since no significant trends in TRMM are reported in 2003–2006). For the period 2007–2010, subbasins B and D registered significant negative trends in groundwater storage (-1.1 cm/yr (-5.02 km^3/yr), $p = 0.004$; -1.5 cm/yr (-5.87 km^3/yr), $p = 0.012$, respectively) (Table 3). TRMM data show that precipitation slightly decreased by about 0.32 cm/yr over the entire region during the period 2007–2010, consistent with other regional studies [*Becker et al.*, 2010; *Awange et al.*, 2014a]. Given the relationship between groundwater storage and infiltration, we attribute the observed 2007–2010 declines in groundwater storage in subbasins B and D to a combination of groundwater use and reduced infiltration. The negative trend in GWS was larger (-1.4 cm/yr) during 2007–2010 (slightly decreased overall precipitation) than it was (-0.87 cm/yr) in 2003–2006 (relatively stable overall precipitation) for subbasin D, a result that shows that the combined contribution of groundwater use and variations in precipitation exposes the groundwater system to additional stress. Elsewhere [*Scanlon et al.*, 2015], groundwater storage declines during reduced precipitation periods have been attributed to reduced recharge and increased groundwater discharge (e.g., for irrigation).

5. Conclusions

The East African region is largely characterized by highly variable surface water resources and low-productive crystalline aquifers, with limited hydro-geologic observations. Our study shows that the GRACE satellite captures the changes in total water storage well, based on our analysis of the water balance ($\rho = 0.6$ and RMSE = 3.3 cm between the GRACE-based dTWS′/dt and observed P-E-Q) and groundwater storage ($\rho = 0.6$ and RMSE = 3.1 cm between GRACE groundwater estimates and in situ groundwater observations), allowing for evaluations of groundwater behaviors over the entire East African region. Central to our analysis is the use of filtered data sets in the disaggregation of the groundwater component from the GRACE total water storage signal. When compared to conventional scaling methods, our approach, which accounts for surface water storage, performs better. The correlation coefficient ρ between GRACE groundwater estimates and in situ observations improves by $\sim$75% while the RMSE reduces by 14% for our approach as compared to the commonly used scaling-based methods. The goal of conventional scaling of GRACE products is to restore signal loss that results from postprocessing. However, scaling should be done only when one is sure that the GRACE residuals are not smeared with unwanted signals, e.g., from surface water bodies. Our results demonstrate that a careful postprocessing of the GRACE data is required to account for the surface water storage component correction, especially over regions with large and highly variable lakes and reservoirs to accurately assess regional groundwater behaviors.

Precipitation is a significant driver in the observed total water storage behavior over the entire region, contributing about 60% to TWS changes in parts of the region, e.g., the Upper Nile basin. We find that the semi-annual precipitation behavior is similar to that of total water storage (with a month time lag between peak P and peak TWS) and that at interannual timescales; high precipitation events (e.g., SOND 2006) which are associated with large-scale circulations (e.g., ENSO and IOD [*Awange et al.*, 2014a]) play a role in eliminating accumulated declines in total water storage from previous years, if any. The combined impact of human groundwater use and precipitation variability, also demonstrated by other regional studies [e.g., *Awange et al.*, 2007, 2014a; *Swenson and Wahr*, 2009; *Becker et al.*, 2010], leads to additional stress in some of the regional aquifers. For example, subbasin D reported a negative trend in groundwater storage of -0.87 cm/yr during the relatively stable precipitation period 2003–2006 (attributed to groundwater use) and -1.4 cm/yr during the low precipitation period 2007–2010 (attributed to both groundwater use and reduced infiltration). With increasing variability in regional precipitation [*Yang et al.*, 2014], and increased groundwater demands from growing populations and industry, groundwater sustainability is questionable.

This study has provided the first regional view of aquifer behavior within the East African region. Southern Kenya (i.e., subbasin C) and Tanzania (i.e., subbasins D and E) exhibit permeable and/or shallow aquifer systems that are recharged by rapid infiltration from precipitation within the basins themselves. Groundwater behavior and precipitation are in phase within these basins, peaking 1 month after the peak rains. Our observations indicate that groundwater behaviors in subbasins C and E are largely driven by groundwater use while subbasins B and D are driven by a combination of reduced natural infiltration and groundwater use. Such results suggest that changes in the ratio of groundwater use to precipitation over these subbasins may result in short-term groundwater exploitation with the potential for rapid rebound in groundwater storage after subsequent heavy rainfall events, similar to "Variable Stress" aquifers identified in *Richey et al.* [2015a]. The groundwater system underlying the Upper Nile (subbasin A) seems to be generally deep and/or less permeable, resisting faster infiltration, and thus is highly susceptible to depletion if groundwater use exceeds recharge. Over the semiarid parts of Northern-Kenya (subbasin B) the groundwater system seems to be hydraulically connected to recharge zones outside the basin boundary that need to be identified and characterized for sustainable groundwater management in this dry-land region.

Uncertainties in the results are attributed to limitations in data sets used. It is worth noting that uncertainties in model soil moisture, which result from poor representation of the vadose zone (including soil type, depth, and thickness) in land surface models, propagate into the GRACE groundwater estimates. A recent study [*Swenson and Lawrence*, 2015] while comparing GRACE TWSA and CLM model TWSA shows that the interannual variability in total water storage depends on the thickness of the soil column, which is currently poorly represented in LSMs. For instance, LSM simulations with thin soil columns promote faster depletion of soil moisture from the surface layers while thick soil columns sustain near surface soil moisture while depleting deeper layer soil moisture (through prolonged evapotranspiration). The study [*Swenson and Lawrence*, 2015] also showed that with a soil column depth of $\sim$2–3 m over semiarid regions (e.g., lower Colorado river basin), all of the observed GRACE TWS decline could be explained by model soil moisture variability (i.e., due to climate variability and not human groundwater use), signaling that care should be taken when making conclusions about observed trends in GRACE groundwater estimates. On the other hand, the use of constant yield values across the basins also creates bias in the groundwater results [*Rodell et al*., 2006; *Shamsudduha et al*., 2012]. The results also demonstrate the importance of having more in situ groundwater samples including those for deeper aquifer layers (i.e., GRACE groundwater estimates compare better with in situ observations for periods with more groundwater samples), portraying a need for more representative hydro-geologic data sets over the region.

For a region that is under-sampled and where policies limit data accessibility, this study has demonstrated the utility of applying GRACE in monitoring groundwater resources over East Africa. Our results clearly identify patterns of hydro-climatology and groundwater use that has impacted changes in groundwater storage in the data-sparse region. Such results can be readily applied to regional water resources modeling to interpret local-scale water resources availability. The study was limited in timeframe given short period data availability especially for small lakes and in situ groundwater. We thus recommend that future works consider extending the analysis with inclusion of more surface lakes, to further isolate long term anthropogenic and climate related controls on regional groundwater and associated trends.

Acknowledgments
This study was supported through fellowships and research grants awarded to the first author including the Fulbright Science and Technology fellowship, National Aeronautics and Space Administration (NASA) Earth and Space Science graduate student fellowship, Faculty for the Future Schlumberger fellowship, the Intergovernmental Panel on Climate Change (IPCC) research grant and the Department of Earth System Science, University of California, Irvine. The support of the NASA GRACE Science Team, the University of California office of the President Multi-campus research and programs initiative, the University of California School of Physical Sciences and Makerere University, Department of Geography, Geo-Informatics and Climatic Sciences are gratefully acknowledged. A portion of the research was carried out at the Jet Propulsion Laboratory, California Institute of Technology, under a contract with the National Aeronautics and Space Administration. This study was also made possible using freely available data from the Global Land Data Assimilation System (http://disc.sci.gsfc.nasa.gov/hydrology/data-holdings) while the GRACE data are available at http://grace.jpl.nasa.gov, supported by the NASA MEaSUREs Program. We acknowledge Andreas Guentner from GFZ for the WaterGAP model outputs. We also acknowledge the Uganda Department of Water Resources Monitoring (DWRM) for availing us with ground-based observations data sets. We wish to thank the Associate Editor and anonymous reviewers for their constructive comments which improved our manuscript. Additional data used in t/his study are available from the first author upon request (nantezajmt8@gmail.com).

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