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L'évaluation de l'impact de la variabilité climatique sur les ressources en eau souterraine dans les grands aquifères

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Abstract

Groundwater is the primary source of drinking water worldwide and plays a critical role in supporting food and energy production. It provides drinking water to at least 50% of the world's population (UNESCO-WWAP, 2009) and represents 43% of all of the water used for irrigation (Siebert et al. 2010). Groundwater plays an important role in society's adaptation to climate variability and change, especially because it is more resilient to the effects of climate change than surface water (Van der Gun, 2012, Taylor et al., 2013). Its unique buffer capacity provides a major strength to reduce the risk of temporary water shortage, and to create conditions for survival in areas were climate change is expected to cause water stress (e.g. semi-arid and arid regions). While groundwater abstraction has increased by more than 300% over the past 50 years with major socioeconomic benefits (Van der Gun, 2012), its development and use has often fallen outside of governance frameworks (FAO, 2015).

A challenging aspect to improve groundwater governance comes from the fact that a major share of world's groundwater volume in storage is located in a limited number of very large aquifer systems over 100 000 km² (Margat and Van der Gun, 2013), and may consequently add a transboundary dimension that needs to be taken into consideration in policy-making. Water decision-makers increasingly require innovative aquifer management tools that address the broad impacts of global change on aquifer storage and depletion trajectory management, land use, groundwater-dependent ecosystems, seawater intrusion, anthropogenic and geogenic contamination, supply vulnerability, and long-term sustainability. NASA's Gravity Recovery and Climate Experiment (GRACE) - first satellite mission able to monitor total water storage changes (including groundwater) remotely – has provided new insights of the dynamics of large aquifers (Scanlon et al., 2012, Richey et al., 2015) since 2002. However, given that the dynamics of groundwater are not solely a function of temporal patterns in pumping, but are also affected by internannual to multidecadal climate variability (Shamsudduha et al. 2012, Kuss and Gurdak, 2014), longer observation time than the one GRACE analyses currently permit is required to separate the respective impacts of anthropogenic activities (land use changes, abstraction) and climate on water resources. Thus, there is a need to extend storage information provided by GRACE to the "past" to better evaluate the current and future evolution of groundwater resources.

This study aims at paving the way to better water management decisions and policies in large aquifers by "reconstructing" past groundwater storage changes as a cornerstone to provide a first quantitative evaluation of the potential effects of anthropogenic activities and climatic oscillations cycles such as the El Niño Southern Oscillation (ENSO) (2–7 year cycle), Pacific Decadal 20 Oscillation (PDO) (10–25 year cycle), and Atlantic Multidecadal Oscillation (AMO) (50–70 year cycle) on large aquifers (area > 100 000 km²) located in arid/semi-arid and temperate regions. Validation is carried out by comparing obtained modeled results with GRACE groundwater storage changes, and ground-based measurements.

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1. Introduction

Groundwater is the primary source of drinking water worldwide and plays a critical role in supporting food and energy production. It provides drinking water to at least 50% of the world's population (UNESCO-WWAP, 2009) and represents 43% of all of the water used for irrigation (Siebert et al. 2010). Groundwater plays an important role in society's adaptation to climate variability and change, especially because it is more resilient to the effects of climate change than surface water (Van der Gun, 2012, Taylor et al., 2013). Its unique buffer capacity provides a major strength to reduce the risk of temporary water shortage, and to create conditions for survival in areas were climate change is expected to cause water stress (e.g. semi-arid and arid regions).

While groundwater abstraction has increased by more than 300% over the past 50 years with major socioeconomic benefits (Van der Gun, 2012), its development and use has often fallen outside of governance frameworks (FAO, 2015). As a result, unrestricted pumping and pollution have led to threats to the sustainability of some aquifers, and the allocation and use of groundwater have often been poorly aligned with society's goals for equity, sustainability and efficiency. Gleeson et al. (2012) estimated that the size of the global groundwater foorprint (the area required to sustain groundwater use and groundwater-dependent ecosystem services) is currently 3.5 times the actual area of aquifers and that about 1.7 billion people live in areas where groundwater resources and/or groundwater-dependent ecosystems are under threat. However, 80% of aquifers have a groundwater footprint that is less than their area, meaning that the net global value is driven by a few heavily exploited aquifers (e.g. Western Mexico, North Arabian, Upper Ganges, and Ogallala Aquifer). Hence, awareness has arisen to improve groundwater governance (FAO, 2015).

A challenging aspect to improve groundwater governance comes from the fact that a major share of world's groundwater volume in storage is located in a limited number of very large aquifer systems over 100 000 km² (Margat and Van der Gun, 2013), and may consequently add a transboundary dimension that needs to be taken into consideration in policy-making. Out of the 592 Transboundary Aquifers (TBAs) that have been identified by IGRAC & UNESCO-IHP (72 in Africa and the Middle East, 129 in Asia and Oceania, 73 in the Americas), only 6 are under a legal agreement for their sustainable management (4 in Africa and the Middle East, 1 in the Americas and 1 in Europe) (Eckstein and Sindico, 2014). The very little number of agreements on TBAs is mostly due to the fact that the dynamics of such aquifers are not yet fully monitored and understood because of data scarcity and accessibility (e.g. geography, conflicts).

Water decision-makers increasingly require innovative aquifer management tools that address the broad impacts of global change on aquifer storage and depletion trajectory management, land use, groundwater-dependent ecosystems, seawater intrusion, anthropogenic and geogenic contamination, supply vulnerability, and long-term sustainability. Therefore, it is particularly important to understand teleconnections in groundwater with interannual to multidecadal climate variability because of the tangible and near-term implications for water-resource

management and policy making (Kuss and Gurdak, 2014). NASA's Gravity Recovery and Climate Experiment (GRACE) - first satellite mission able to monitor total water storage changes (including groundwater) remotely – has provided new insights of the dynamics of large aquifers (Scanlon et al., 2012, Richey et al., 2015) since 2002. However, given that the dynamics of groundwater are not solely a function of temporal patterns in pumping, but are also affected by internannual to multidecadal climate variability (Shamsudduha et al. 2012, Kuss and Gurdak, 2014), longer observation time than the one GRACE analyses currently permit is required to separate the respective impacts of anthropogenic activities (land use changes, abstraction) and climate on water resources. Thus, there is a need to extend storage information provided by GRACE to the "past" to better evaluate the current and future evolution of groundwater resources.

This study aims at paving the way to better water management decisions and policies in large aquifers by "reconstructing" past groundwater storage changes fluctuations as a cornerstone to provide a first quantitative evaluation of the potential effects of anthropogenic activities and climatic oscillations cycles such as the El Niño Southern Oscillation (ENSO) (2–7 year cycle), Pacific Decadal 20 Oscillation (PDO) (10–25 year cycle), and Atlantic Multidecadal Oscillation (AMO) (50–70 year cycle) on large aquifers (area > 100 000 km²) located in arid/semi-arid and temperate regions. Validation is carried out by comparing obtained modeled results with GRACE groundwater storage changes, and ground-based measurements.

2. Background

2.1. Study areas

2.1.1. High Plains (Ogallala) Aquifer (USA)

The Ogallala Aquifer is an unconfined aquifer that underlies about 450,000 km² of the High Plains of the United States, extending northward from western Texas to South Dakota (Figure 1). Mean annual temperature ranges from about 6°C in the north to 17°C in the south. Mean annual precipitation ranges from 305 mm in the west to 840 mm in the east (USGS, 2015). The Ogallala is an unconfined aquifer, and virtually all recharge comes from rainwater and snowmelt. Recharge is considered to be minimal and is likely to be related to climate variability (Gurdak et al., 2007). Most of the water in the Ogallala Aquifer is derived from precipitation on the northern part of the aquifer (237 km^3, mean 1971–2000), A newly developed recharge map of the Ogallala Aquifer differentiates three areas. High recharge in the Northern High Plains (NHP) results in sustainable pumping, whereas lower recharge in the Central and Southern High Plains (CHP and SHP, respectively) has resulted in focused depletion of about 330 km^3 of fossil groundwater, mostly recharged during the past 13,000 (Scanlon et al., 2012). Values of porosity of about 30% were found for the southern part of the Ogallala Aquifer (Ashworth, 1980). For global analysis, the mean value is 15% (Kuniansky, 2001). In the Ogallala Aquifer, surface water resources are dominated by

internally drained ephemeral lakes or playas (~50,000 playas) because of the extremely flat topography.

The Ogallala Aquifer is ranked first among aquifers in the United States for total groundwater withdrawals (mainly for crop irrigation), and represents approximately 23 000 Mm^3/year, that is about 30% of all groundwater pumped in the USA (Maupin and Barber, 2005). This regional aquifer is one of the most intensively monitored aquifers globally. Groundwater level data based on water level monitoring in 3600 wells (1950s) to 9600 wells (2006) show generally monotonic declines with some recovery periods in mid-1980s, mid-1990s, and mid-2000s (Scanlon et al., 2012).

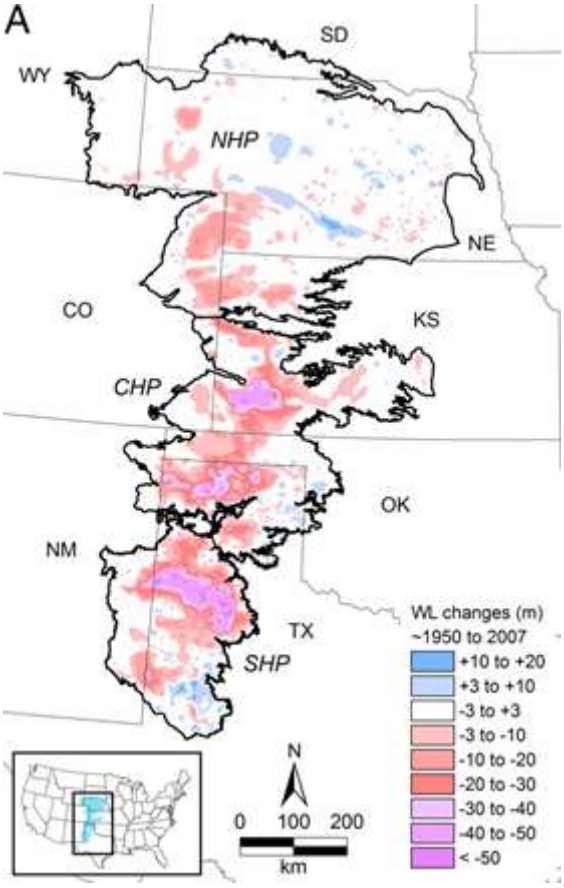

Figure 1 - Delineation of the Ogallala Aquifer (Scanlon et al., 2012)

2.1.2. Stampriet Transboundary Aquifer System (Botswana, Namibia and South Africa)

The Stampriet Transboundary Aquifer System (STAS) stretches from Central Namibia into Western Botswana and South Africa's Northern Cape Province, and lies within the Orange River Basin. The STAS covers a total area of about 100 000 km^2 (Figure 2). The STAS area is lightly populated (≈45 000) with population concentrated in small rural settlements. The STAS is made up of two deep confined transboundary aquifers in the Karoo sediments (Auob and Nossob aquifers), overlain by an unconfined non-transboundary aquifer system of

Kalahari and upper Karoo sediments (Kalahari aquifers). The STAS is located in an arid area with an annual mean temperature varying between 19 and 22°C. Temperature in summer can reach 50°C. Average rainfall in the STAS area is of 150 to 310 mm/yr. Recharge to the Kalahari aquifers during years with average rainfall is estimated at 0.5% of rainfall. Recharge to the Auob and Nossob aquifers in normal rainfall years is negligible but considerable recharge occurs during extreme rainfall events. Groundwater is the major source of water in the STAS, to provide portable water to the people, livestock and for irrigation (GGRETA, 2015). Surface water is scarce and unreliable which makes groundwater the major source of water in the STAS. It provides portable water to people, livestock and irrigation. There are neither industries nor mining activities taking place in the STAS area. Approximately 20 Mm^3/year are abstracted in the STAS, most of which occurs in Namibia (over 95%).The largest consumer of water is irrigation (~46%) followed by stock watering (~38%) and domestic use (~16%). ephemeral rivers only flow in periods of high rainfall. Values of porosity of about 25% were found for the Kalahari aquifers (Vogel et al., 1982).

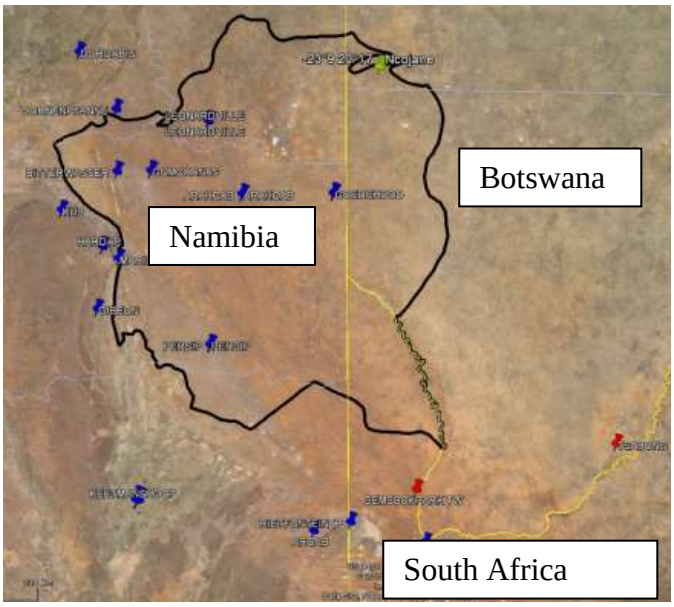

Figure 2 – Delineation of the Stampriet Transboundary Aquifer (in balck)

No long-term depletion has been observed in the STAS although a water level drop was observed in the late 1990s (JICA, 2002). Above-normal rainfall in the early 2000s has led to a water level rise (Matsheng, 2007).

2.1.3. Karoo Sedimentary Aquifer (Lesotho and South Africa)

The Karoo Sedimentary Aquifer covers an area of approximately 135 000km² in Lesotho and South Africa (Figure 3). Population and average rainfall in the area are 4 700 000 and 680mm/year, respectively. The Karoo Sedimentary Aquifer is a multi-layered system (5 layers within Lesotho and 4 layers within South Africa) that is mostly semi-confined, but some parts are unconfined. The average rest water level is between 20m and 33m and the

average depth to the top of the aquifer is 22m within Lesotho. The predominant source of recharge is through precipitation over the aquifer area. The predominant discharge mechanism is through springs within Lesotho. The mean annual recharge is 650 Mm^3/year. The size of the recharge area over the aquifer is 76 078 km^2. The predominant lithology is sedimentary sandstones that are characterized by a low to high primary porosity (≈25%) (Schmitz and Rooyani, 1987), with secondary porosity (fractures) and there is generally a low horizontal and vertical connectivity. The transmissivity values are low with an average value varying between 20 m^2/day (South Africa) and 43 m^2/day (Lesotho). Groundwater abstraction mainly occurs in Lesotho and accounts for 25 Mm^3/year (i.e. approximately 10% of total freshwater withdrawals in the aquifer area). No long-term depletion has been observed in the Karoo Sedimentary Aquifer.

Figure 3 - Delineation of the Karoo Sedimentary Aquifer (TWAP, 2015)

2.1.4. Irhazer-Illuemeden Basin Aquifer (Algeria, Benin, Mali, Niger, Nigeria)

The Irhazer-Illuemeden Basin Aquifer is located in central West Africa is a sedimentary basin shared over about 95% of its surface area between Niger, Mali and Nigeria, with minor non-connected sections in Algeria and Benin (Figure 4). The Irhazer-Illuemeden Basin Aquifer covers an area of approximately 510 000km^2. Population and average rainfall in the area are 18 000 000 and 310mm/year, respectively. The Irhazer-Illuemeden Basin Aquifer is an aquifer system composed by the cretaceous calcareous sandstone Continental Intercalaire (CI) and the tertiary sandstone Continental Terminal (CT) aquifers. The aquifers are mostly confined, but some parts are unconfined. The CT and CI aquifers outcrop 150,000 km and 250,000 km^2 over the basin, respectively. The predominant source of recharge is from precipitation over the aquifer area (Benin, Mali), and from runoff along river systems (Niger, Nigeria). The predominant discharge mechanism is through river base flow (Benin, Nigeria)

and through evapotranspiration (Mali). The predominant aquifer lithology consists of sedimentary rocks –sandstones (Benin, Mali), and sediments – gravel (Nigeria). The integranular aquifer is characterised by a low primary porosity (5%) (UNESCO, 2004).

Most of the population lives in small villages with a few hundreds of inhabitants in the southern part of the basin, where rainfed agriculture is dominant. To the north, livestock breeding is a main economic activity (Guengant and Banoin 2003). Groundwater abstraction has increased steadily from 50 to 250 Mm^3/year over the past 50 years (Dodo and Baba Sy, 2010). Favreau et al., (2009) made an extensive review of water table fluctuation in the Irhazer-Illuemeden Basin Aquifer since the early 1930s based on the little information available. It is suggested that a water table rise of up to 20m occurred in the area from the early 1930s to late 1950s (Jones, 1960 and Barber and Dousse 1965). During the drought of mid-1970s and 1980s throughout the Sahel, groundwater levels dropped an estimated 0.5 to 1 m/year (Reij 1983), and it many wells and boreholes went dry just after the end of the rainy season. Piezometric surveys in the Niger part of the CT aquifer were reviewed by Guéro (2003) and Favreau et al. (2009). Piezometric surveys performed to date showed a rise in the water table. Present day (2010) water table levels are the highest ever recorded, and measured rise intensities range from 0.1 m/year to up to 0.4 m/year (Leduc et al. 2001).

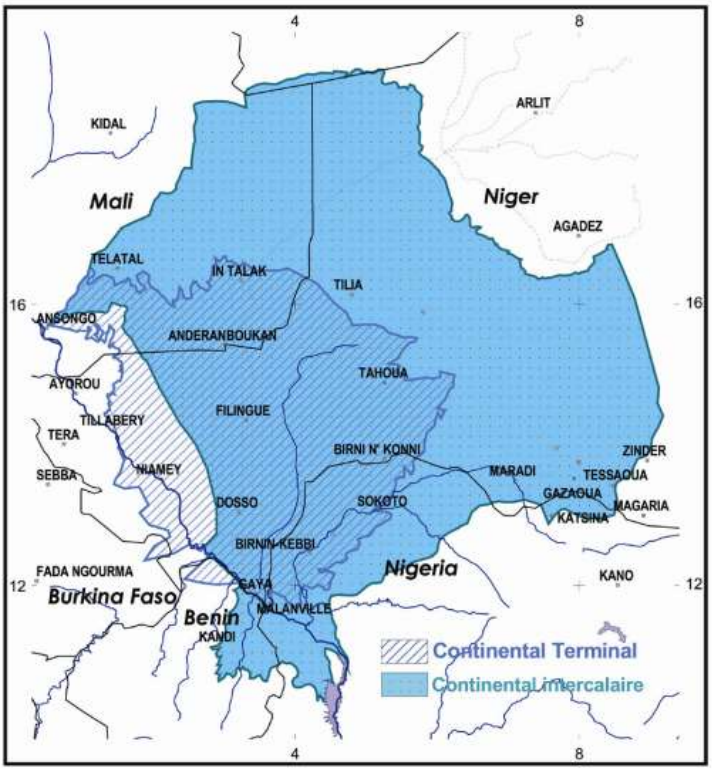

Figure 4 - Delineation of the Irhazer-Illuemeden Basin Aquifer (Dodo and Baba Sy, 2010)

Although not in the studied area, an observational record of groundwater levels (borehole depth = 20m) in the Sahel from 1978 to 2004 was found (Yameogo, 2008). A downward trend from 1978 to 1985 is observed. From 1985 to 1988, groundwater level stabilized, followed by a rise in the water table (Figure 5).

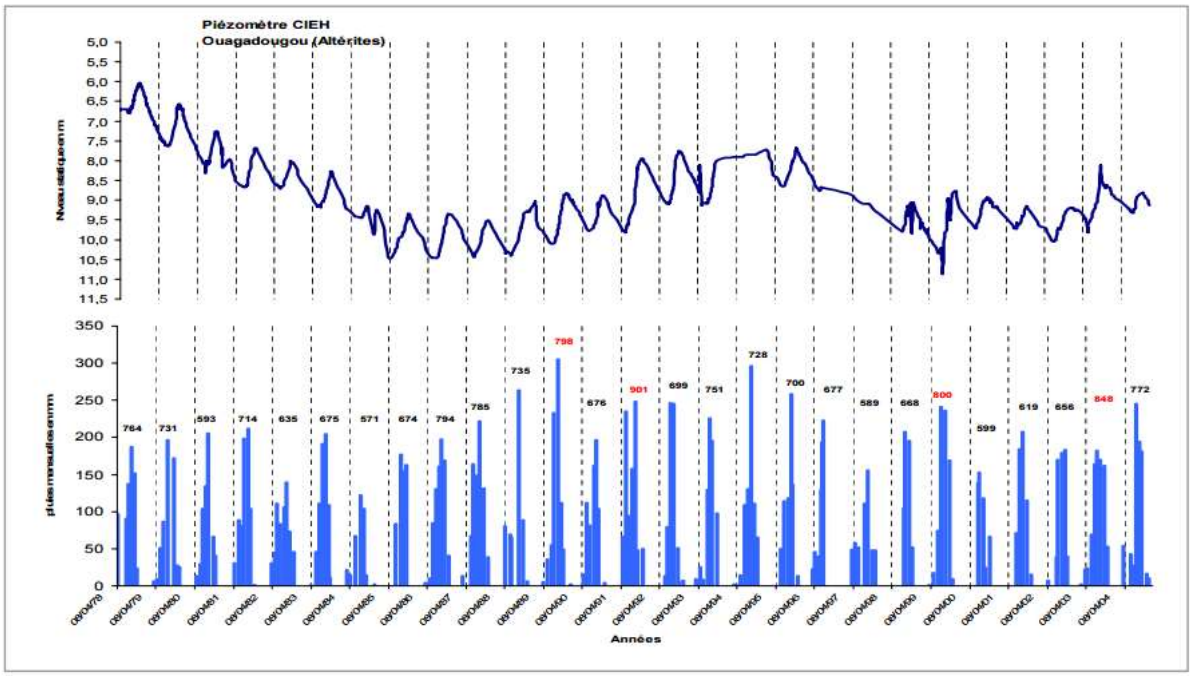

Figure 5 - Groundwater level and rainfall record in Ouagadougou (CIEH). Average rainfall (1978-2004): 709 mm (Yameogo, 2008)

2.1.5. Syr Darya Aquifer (Kazakhstan, Uzbekistan)

The Syr Darya Aquifer is located in Central Asia, and covers an area of 300 000km² across Kazakhstan and Uzbekistan (Figure 6). Total population and averga rainfall are 1 800 000 and 160mm/year, respectively. The aquifer is a multiple 3-layered hydraulically connected system that is mostly confined but some parts are unconfined. It is made of a confining layer of the Paleogene age (100 m in thickness) separating two hydrogeological levels: a top level, Pliocene-Quaternary complex - sedimentary aquifer mainly gravel, sand with high primary porosity (≈20%) and no secondary porosity, and a middle level Cretaceous complex - sedimentary aquifer mainly sand with high primary porosity and no secondary porosity. The average transmissivity is 3 300 m²/d. The annual recharge is estimated at 2 800 Mm³/year. The top aquifer is recharged by inflows of interstitial and karst waters from overlying Paleozoic rocks. Recharge also occurs by infiltration of rainfall, surface waters from rivers and streams, and groundwater that circulate through tectonic discontinuities. The regional direction of the groundwater flow is towards the local base level, the Aral Sea. The mean annual volume of groundwater abstraction in Kazakhstan is 120Mm³/year, largely for domestic use. This is less than 5% of the available recharge and no trends on water level depletion have been observed (TWAP, 2015).

Figure 6 - Delineation of the Syr Darya Aquifer (TWAP, 2015)

2.1.6. East Ganges River Plain Aquifer (Bangladesh, India)

The East Ganges River Plain Aquifer is shared by Bangladesh and India and covers an area of approximately 180 000 km² (Figure 7). Population and average rainfall are 230 000 000 and 1900mm/year. The aquifer is a multiple 3-layered hydraulically connected system that is mostly confined but some parts unconfined. The average depth to the water table varies between <5 m (Bangladesh) and 10m (India). The average depth to the top of the aquifer varies from <5 m (Bangladesh) to 7 m (India) while the average thickness of the aquifer system is between 400 m (Bangladesh) and 600m (India). The predominant source of natural recharge is through monsoon precipitation over the aquifer area and through recharge from river flood plains. The major discharge mechanism is through river base flow and through groundwater flow into another aquifer. There are extreme recharge events but no data was found for average extreme amounts. A significant portion of the recharge is not through natural causes but is through return flows from irrigated lands. The predominant aquifer lithology is sediment – sand that has a high primary porosity ≈20 (BGS and DPHE, 2001) with a high horizontal and a low vertical connectivity. The average transmissivity value varies between 1 500 m^2/d and 4 500 m^2/d.

The East Ganges River Plain Aquifer is extensively used for irrigation. Agriculture in Bangladesh was entirely dependent on surface water and monsoon rainfall prior to the 1970s (UNDP, 1982). Irrigated agriculture using groundwater through power operated pumps was introduced in the late 1970s. Since then, groundwater-fed irrigation area has steadily increased and nowadays account an area of 32 000km² in Bangladesh. Statistics reveal that

about 75 percent of total cultivated land is irrigated by groundwater and 25 percent by surface water. From 1979 to 2003, groundwater-fed irrigation for dry season rice cultivation in Bangladesh increased by approximately 875 Mm³/year (BADC 2003) elevating annual rice production from 11.9 megatonnes (Mt) in 1975 to 27.3 Mt in 2006-2007 (Bangladesh Bureau of Statistics 2008). Over the last 50 years, groundwater abstraction on the Indian subcontinent increased from about 10-20 000 Mm³/year to approximately 260 000 Mm³/year (Shah et al. 2003; Giordano 2009). Shamsudduha et al., 2012 depicted declining trends in most parts of Bangladesh, although the magnitudes of these trends vary spatially. Long-term (1985 to 2005) trends in groundwater levels of shallow aquifers across Bangladesh are shown in Figure 8 show contours of linear trends (cm/year) during the dry season (5th percentile), wet season (95th percentile), and in overall (annual mean) time series. Strong declining trends (0.5 to 1 m/year) in dry-period groundwater levels are observed in the central part of the country surrounding the Dhaka city. Moderately declining trends (0.1 to 0.5 m/year) occur in western, northwestern, and northeastern areas. In the northern piedmont areas and floodplains of the major rivers, magnitudes of declining trends are low (0.01 to 0.05 m/year). Stable or slightly rising trends (0 to 0.1 m/year) are generally observed from the Meghna estuary to the southern coastal areas in the country. A similar overall pattern is seen during wet periods except in the northern piedmont areas, southwestern delta plains and southern coastal areas where wet period trends are slightly rising or stable. Similar to long-term trends during dry and wet periods, declining trends in annual mean groundwater levels are observed in the central, northwestern, and northeastern parts. Relatively stable to rising mean groundwater levels are detected in the northern piedmont, floodplains of major rivers, and deltaic plains.

Figure 7 - Delineation of the East Ganges River Plain Aquifer (TWAP, 2015)

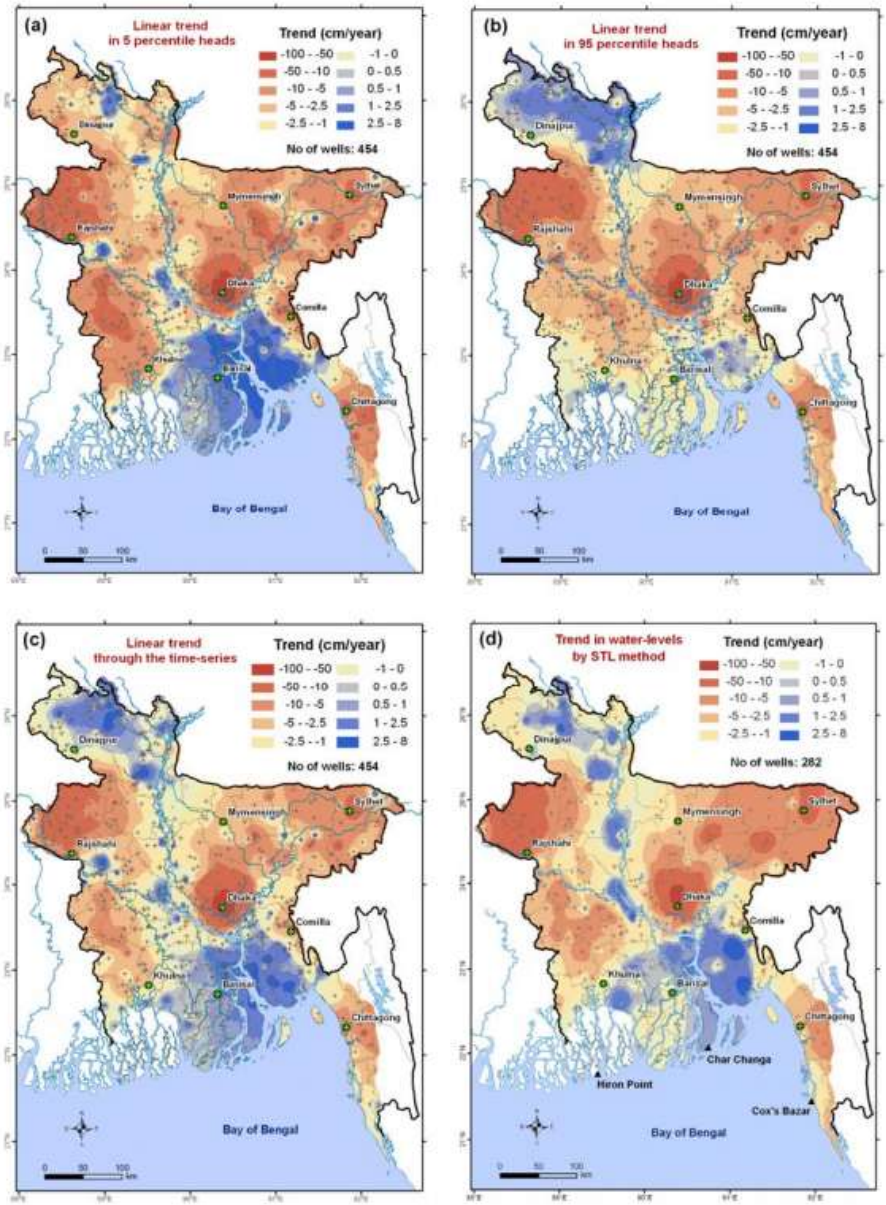

Figure 8 - Trends in groundwater levels for the period of 1985 to 2005. Linear trends in the dry period groundwater levels (5th percentiles of observations in each year) are shown in (a), trends in the wet-period groundwater levels (95th percentiles) are shown in (b), linear trends in annual means are shown in (c), and nonparametric trends calculated from the long-term trend component derived from an STL decomposition are shown in (d). Three locations in coastal regions of Bangladesh are shown in (d) where linear trends in sea levels were calculated by Singh (2002) (Shamsudduha et al., 2012)

Available groundwater levels fluctuation in the studied aquifers is given in Annex 1.

2.2. Climate variability: El Niño Southern Oscillation (ENSO), The Pacific Decadal Oscillation (PDO) and the Atlantic Multidecadal Oscillation (AMO)

The El Niño Southern Oscillation (ENSO) is a coupled ocean-atmospheric phenomenon that has interannual variability with irregular 1- to 6-year cycles between the warm (El Nino) and cold (La Nina) phases that has been occurring for the past 300 years. An El Niño is characterized by stronger than average sea surface temperatures in the central and eastern equatorial Pacific Ocean, reduced strength of the easterly trade winds in the Tropical Pacific, and an eastward shift in the region of intense tropical rainfall. A La Niña is characterized by the opposite – cooler than average sea surface temperatures, stronger than normal easterly trade winds, and a westward shift in the region of intense tropical rainfall. Although ENSO is centered in the tropics, the changes associated with El Niño and La Niña events affect climate around the world. For instance, Maidment et al, (2015) suggest that the trend to more La Niña-like conditions since 2000 is a likely contributing factor driving the increase in Southern Africa rainfall between 1983 and 2008. Typical ENSO effects are shown in Figure 9:

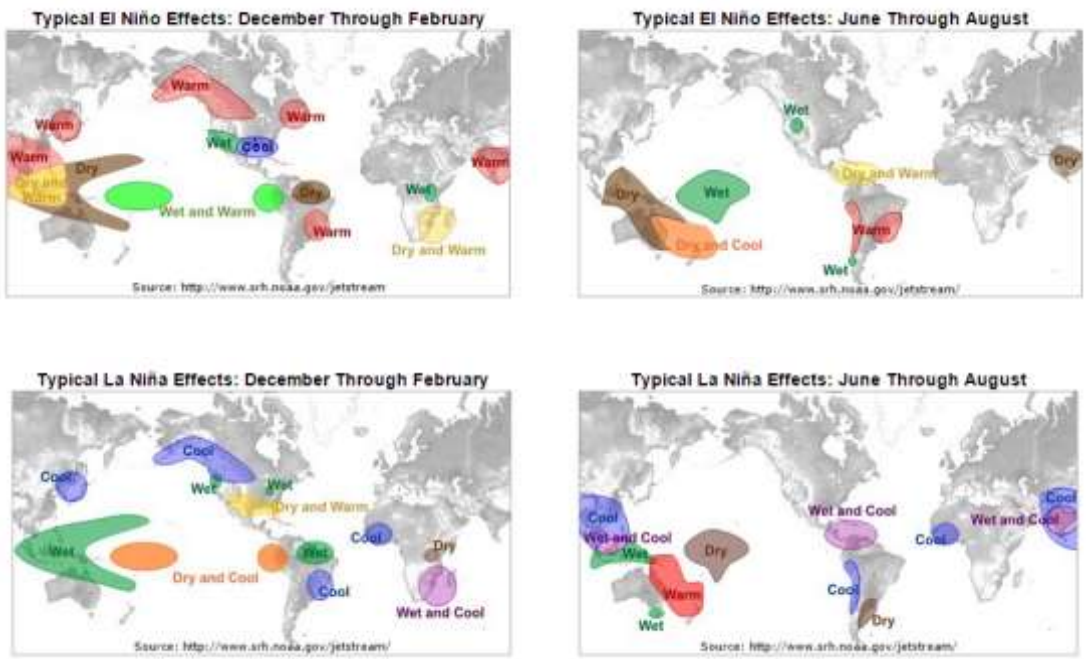

Figure 9 - Typical ENSO effects around the world (www.srh.noaa.gov/jetstream/)

The Pacific Decadal Oscillation (PDO) is a climate index based upon patterns of variation in sea surface temperature of the North Pacific Ocean from 1900 to the present (Mantua et al. 1997) with warm and cold phases that can persist for 20-30 years. Unlike ENSO, the PDO is not a single physical mode of ocean variability, but rather the sum of several processes with different dynamic origins. Although PDO has an ENSO-like pattern of climate variability, it is distinct from ENSO in three ways:

1) *Location:* The strongest signature of the PDO is in the North Pacific, instead of the tropical Pacific.

2) *Duration:* PDO phases last much longer – typically 20 to 30 years for a single warm or cool phase – than ENSO events – 6 to 18 months for a single warm (El Niño) or cold (La Niña) phase.

3) *Cause and Predictability:* The dynamics of the PDO remains very complex and climate models can't predict the future evolution of the PDO, especially the shift from one PDO phase to another. Even in the absence of a theoretical understanding, the PDO signal improves the climate forecasts combined with ENSO for different regions of the world due to its strong tendency for multi-decadal persistence.

The scientific community is reasonably agreed on the factors contributing to ENSO events, making it possible to provide skilled forecasts of ENSO events several seasons in advance of the event's onset (for a sample of forecast centers, see Seasonal to Interannual Forecasts). The causes of the PDO, on the other hand, are not understood. Part of the difficulty in understanding what triggers PDO phase shifts is the persistence of PDO events. Accurate instrumental records for the North Pacific begin around 1900; because of the persistence of the PDO phases, we have seen only two complete PDO cycles in that time, making it difficult to determine the cause for – and therefore the predictability of the PDO. However, even in the absence of a theoretical understanding, PDO climate information improves season-to-season and year-to-year climate forecasts for North America because of its strong tendency for multi-season and multi-year persistence. Simply assuming persistence of observed PDO-related North Pacific sea surface temperature (SST) anomalies in the fall in any given year provides some skill in predicting PDO-related winter climate anomalies. However, this persistence based forecast will always fail to predict the relatively infrequent switches from one PDO phase to another.

The PDO was in a warm phase continued from 1925 to 1946 and 1977 to 1998, and in a cold phase from 1947 to 1976. However, these decadal cycles have recently broken down as the PDO entered a cold phase that lasted only 4 years followed by a warm phase of 3 years, from 2002 to 2005, neutral until August 2007 and abruptly changing again to a cold until 2013. Climatic fingerprints of the PDO are most visible in the North Pacific/North American sector. More recently, Maidment et al., (2015) suggest that rainfall increase in Southern Africa is found to be associated with an unprecedented strengthening of Walker circulation (e.g., L'Heureux et al., 2013) and linked to SST patterns related to the PDO. PDO may therefore determine low-frequency rainfall variability over Southern Africa.

The interannual relationship between ENSO and the global climate is not stationary and can be modulated by the PDO. Wang et al. (2014) reported that when ENSO and PDO are in phase, the El Niño/La Niña-induced dry/wet anomalies are not only intensified over the canonical regions influenced by a typical ENSO event but also expand poleward. If ENSO and the PDO are out of phase, then the associated dry/wet anomaly is dampened or disappears. Generally, during the warm phase of the PDO, El Niño induces much broader and more severe droughts over land compared with the cold PDO phase. For example, the arid and

semi-arid climate over the Sahel and southern Africa worsens. Correspondingly, during the cold phase of the PDO, more rain occurs over land in La Niña winters than during the warm PDO phase. However, there are a few exceptions. The amplitude of the dry–wet variation is larger over northern Europe and the Mediterranean during the out-of-phase condition, while the variation over the Horn of Africa tends to be stronger in the cold phase of the PDO.

The Atlantic Multidecadal Oscillation (AMO) is an index of SST over the North Atlantic Ocean (quasi-cycles of roughly 70 years) with cool and warm phases that may last for 20-40 years each and lead to differences of about 15°C between extremes. Paleoclimatologic studies have confirmed that these changes have been occurring over the past 3000 years. The AMO was in warm phases from 1860 to 1880 and 1930 to 1960, and in cool phases from 1905 to 1925 and 1970 to 1990 (Figure 10). Since the mid-1990s we have been in a warm phase whose peak is expected to occur around 2020 (Curry, 2008). The AMO index is correlated to air temperatures and rainfall over much of North America and Europe (Enfield et al. 2001; Sutton and Hodson 2005), North East Brazilian and African Sahel (Folland et al. 2001, Knight et al. 2006), Indian and East Asian summer monsoons (Goswami et al. 2006, Lu et al. 2006, Zhang and Delworth 2007, Li et al. 2008, Song and Hu 2008). Climate models suggest that a warm phase of the AMO strengthens the summer rainfall over India, and Sahel and the North Atlantic tropical zone (Zhang and Delworth, 2006). It is also associated with changes in the frequency of droughts in North America and hurricanes in the Atlantic. The above observational evidences suggest a possible link between the North Atlantic ocean state and Asian summer monsoon intensity on multidecadal and millennial time scales. It has also been reported that AMO modulates the ENSO variability (Dong et al. 2006, Dong and Sutton 2007, Timmermann et al. 2007) as well as the ENSO-Monsoon interaction (Chen et al 2010). When the North Atlantic was anomalous warmer, the main remote features are enhanced Indian summer monsoon and enhanced east Asian summer monsoon.

McCabe et al. (2004) showed that the PDO and the AMO strongly influence multidecadal droughts pattern in the United States. If the PDO is associated with a warm AMO phase, drought frequency is enhanced over much of the Northern United States during warm PDO phase and over the Southwest United States during the cold PDO phase. The Asian Monsoon is also affected, increased rainfall and decreased summer temperature is observed over the Indian subcontinent during the cold PDO phase.

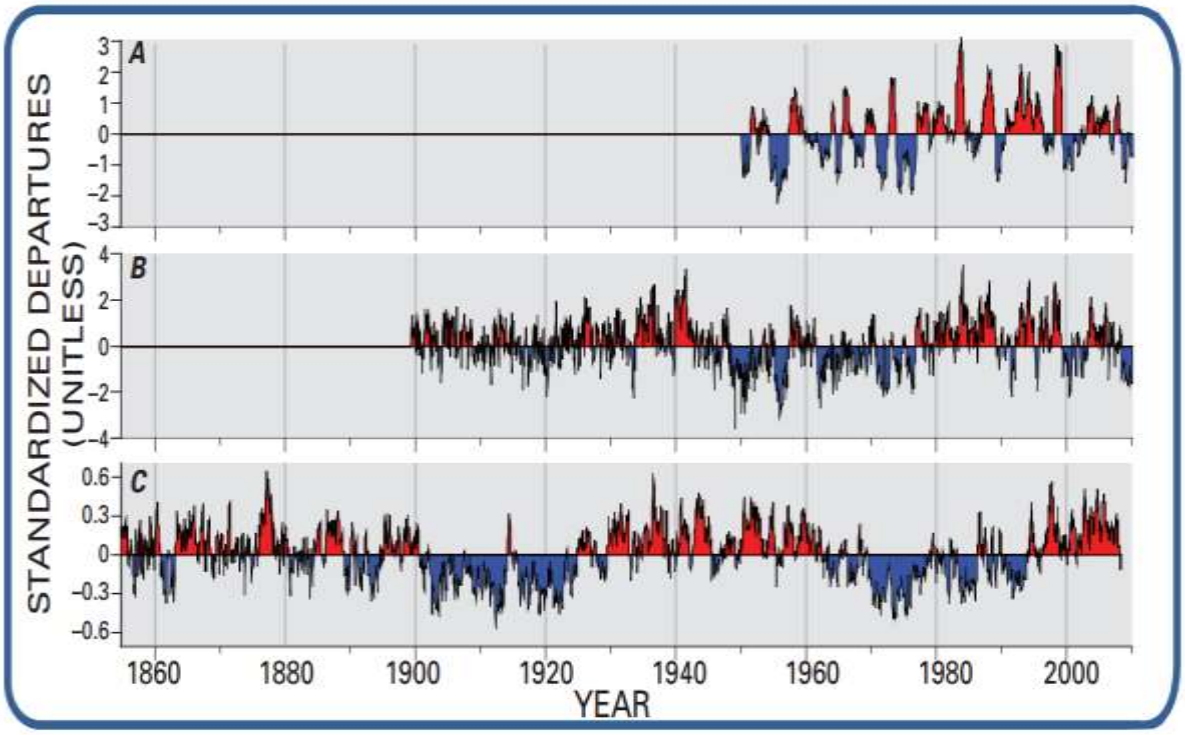

Figure 10 - The monthly El Niño Southern Oscillation (ENSO) index, Pacific Decadal Oscillation (PDO) index, and Atlantic Multidecadal Oscillation (AMO) index (Kuss and Gurdak, 2014)

3. Materials and Methods

3.1 GRACE observations

Launched in March 2002, NASA's Gravity Recovery and Climate Experiment (GRACE) has revolutionized the way large mass changes can be detected on Earth (Tapley et al., 2004). By monitoring the temporal variations of Earth's gravity field with an unprecedented temporal and spatial resolution", GRACE has provided new insights in mass redistribution processes of the atmosphere, the oceans, terrestrial water, and the cryosphere (Ramillien et al., 2008). Consisting of two twin satellites flying in a polar orbit at about 450 km altitude and about 200 km apart, GRACE infers on Earth's gravity variations by constantly monitoring the distance between the two satellites at the micrometer level. Two types of products have been developed from GRACE range-rate data: The first translates satellite range-rate data directly into a set of localized surface mass concentrations, so-called "mascons", e.g. (Rowlands and Luthcke, 2005); the second is a global spherical harmonic (SH) expansion of the gravity field, where a set of Stokes coefficients is the standard GRACE Level 2 product. Note that mascon- and SH-derived mass changes are equivalent (Klees et al., 2008). In this study, the SH formulation is used. A description on how surface mass changes can be derived from SH coefficients is given in Wahr et al., 1998.

3.1.1. GRACE observations processing

GRACE responds to all sources of mass redistribution near the Earth's surface. Thus, it is important to correct measured range-rate variations for well-recognized influences, which include atmospheric mass redistribution, ocean mass redistribution due to currents and winds, ocean and solid Earth tides, and others. When recovering surface mass variations from GRACE products, two major difficulties have to be dealt with. The first is linked to the limited spectral content (i.e. the sensitivity to large spatial scales) of GRACE when focusing on a space-limited area (Simons and Dahlen, 2006). The second is given by the fact that GRACE (and more generally, gravity) provides information about vertically integrated mass changes only, which makes a separation of individual sources challenging. The first point leads to so-called leakage effects (Klees and Zapreeva, 2007), i.e. to a loss in signal amplitude when concentrating GRACE on a region of interest, and a partial compensation from mass changes outside that region. Several methods have been developed to overcome this issue, including the use of a-priori information on the spatial distribution of the expected mass changes (e.g. Longuevergne et al., 2010; Swenson and Wahr, 2002). The second point is inherently difficult, and is generally tackled by using models to discern between individual sources.

The processes that potentially contribute to gravity variations can be subdivided into two categories, i.e. (1) processes related to near-surface mass transport, such as water storage in lakes, the unsaturated zone, glaciers, the seasonal snow cover, and erosion, and (2) internal processes, including glacial isostatic adjustment since the Last Glacial Maximum and vertical crustal movements related to tectonic processes. Previous studies focusing on GRACE and groundwater showed that the largest uncertainties stem from the hydrological contribution. Recently, (Scanlon et al., 2012) showed that forward modelling (i.e. application of a spatial filter to the modelled hydrological contribution in order to mimic the large-scale sensitivity of the GRACE signal) can be used for reducing that uncertainty. Given that the processing method used for the derivation of the GRACE data is known, the required mathematical process is straightforward. Forward modelling the impact of all known contributions to the GRACE signal was previously shown to be the most suited method for extracting a specific storage compartment. This general strategy is adopted in this work. In principle, the impact of all known mass-change contributions derived from models and/or independent estimates are spatially filtered to match the GRACE resolution, subtracted from the total mass change derived from the GRACE data, and the residuals interpreted as glacier mass changes. The so obtained glacier mass changes, which still refer to the GRACE resolution, are re-focused on the region using regular approaches (Longuevergne et al., 2010; Swenson and Wahr, 2002). Mass change rates are then obtained by computing time-series trends through a 4th order low-pass Butterworth in order to remove seasonal variations. In order to account for a wide range of possible error sources, the work is based on a set-like approach that accounts for uncertainties in the (1) GRACE-derived gravity change products, (2) applied spatial filtering and processing strategy, (3) mass-change contributions other than groundwater subtracted from the total signal.

Several processing centers propose GRACE data which is available at the official data repository online at http://podaac.jpl.nasa.gov/grace/ (CSR, GFZ, JPL), and more recently from additional sources such as DEOS (http://www. lr.tudelft.nl/live/pagina.jsp?id=6062b504-715e-4a22-9e87-ab2231914a4b&lang=en), ITG (http://www.geod.uni-bonn. de/itg-grace03.html) or CNES/Groupement de Recherche en Géodesie Spatiale (GRGS) (http://bgi.cnes.fr:8110/geoid- variations/README.html). However, several research centers also propose GRACE products, with, most of the time, alternative strategies. This diversity is important to improve GRACE products and prepare next generation satellite to be launched beginning 2017. Official products are generally provided up to degree 90 (spatial resolution ~220 km), but post-processing is required to increase signal to noise ratio, leading to actual resolution ~400 km at best. A few processing centers propose improved mathematical strategies to invert GRACE data, offering improved spatial resolution at the cost of more constrains on the solutions. GRGS, for example, offer now a dataset at ~300 km resolution which do not any post-processing (Bruinsma et al., 2010). CSR proposes a regularized product (Save et al., 2012), which is provided up to degree 120 (~170 km resolution) without further post-processing required. Figure 11 compares regular CSR products and next generation regularized products from CSR over the Tienshan region (Farinotti et al., 2015). While both give the same total mass over the region as a whole, the regularized CSR product offers a much more detailed description of actual mass changes over sub-regions.

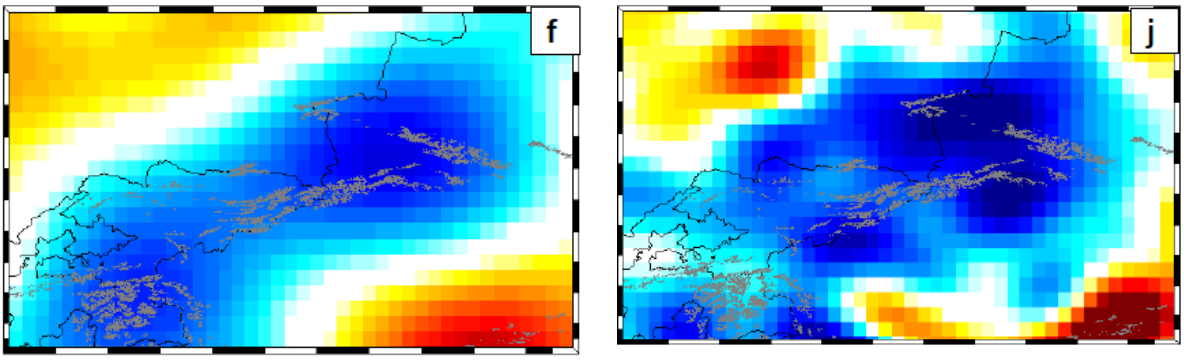

Figure 11 – "Official" CSR products, filtered with DDK4 (Kusche et al., 2009) (left) and Regularized CSR product (Save et al., 2012) without processing (right)

GRACE solutions are continuously improving with growing experience on the satellite behavior and update of so-called "dealiasing products" describing mass variations in the atmosphere and the ocean.

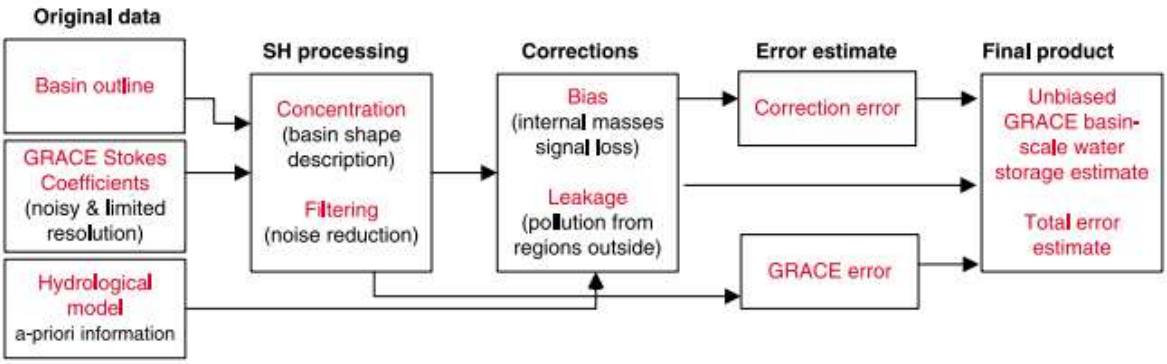

Figure 12 - Diagram synthetizing the Gravity Recovery and Climate Experiment (GRACE) satellite processing for basin-scale water storage estimate on the basis of level 2 spherical harmonics products (Longuevergne et al., 2010)

3.1.2. GRACE observations processing of groundwater storage changes

There is considerable interest in using GRACE satellites to monitor changes in groundwater storage at basin scales because they provide continuous coverage globally and complement long-term water-level monitoring and regional hydrologic modeling. GRACE measures changes in total water storage (S), which are used to estimate changes in groundwater storage (GWS) by substracting all know contribution in storage (snow, soil moisture, surface water). When aiming at recovering glacier-mass changes, water storage variations that need to be subtracted from the GRACE signal include variations in (1) ice storage (ICE) (2) snow water equivalent in seasonal snow pack (SWE), surface water storage (SWS), (2) soil moisture storage (SMS). The combination of the 5 components is referred to as total water storage (S), which is measured by GRACE.

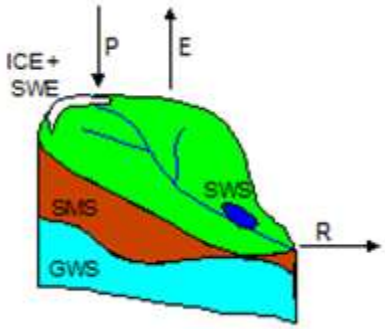

Figure 13 - Water storage components of GRACE signal: ice storage (*ICE*), snow water equivalent in seasonal snow pack (*SWE*), surface water storage (*SWS*), soil moisture storage (*SMS*), groundwater storage (*GWS*)

$$S = ICE + SWE + SWS + SMS + GWS \qquad \text{(Eq. 1)}$$

In this work, we have used a set of 10 global hydrological models and land surface models to describe storage changes in SMS and SWE . As surface water and glacier storage changes are generally not modeled, groundwater systems located close to large lakes (e.g. Caspian) or glaciers can be affected by leakage from these zones. Using a priori monitoring or model-

based estimates of ICE, SWE, SWS, SMS, changes in GWS can be calculated as residual from the disaggregation equation:

$$\Delta GWS = \Delta S - \Delta ICE - \Delta SWE - \Delta SWS - \Delta SMS \qquad \text{(Eq. 2)}$$

GRACE processing was uptaded in this study to provide more reliable estimates of GWS changes with optimal use of available information. The new processing approach differs from the regular approach in calculating GWS from S using filtered data at GRACE resolution before any rescaling is applied (Figure 14). In this updated approach GRACE data were recombined and filtered to provide filtered S as previsouly described. The various water balance components (ICE, SWE, SWS, SMS) were then filtered in the same way as GRACE data, i.e. projection of model grids on spherical harmonics, recombination to maximum degree 50 for comparison with GRGS data or degree 60 for comparison with CSR data and application of a 300km Gaussian filter for comparison with CSR data. Gridded ICE, SWE, SMS data and point SWS data were used, allowing spatial variability in these different storage components to be incorporated in the processing approach, in contrast to the regular processing approach, which uses basin means. Restoring the amplitude of the filtered GWS signal only requires bias correction (simple rescaling) and no leakage correction (no external groundwater masses leaking into the area of interest) because GWS changes are assumed to be concentrated inside the aquifer; therefore, errors associated with leakage corrections should be minimized. Bias correction was done using a multiplicative factor that was calculated from the ratio of unfiltered to filtered GWS changes from output from the hydrologic model. This updated processing approach minimizes reliance on a priori information and allows GRACE to be used as independent observational data as much as possible. However, this updated approach requires knowledge of changes in ICE, SWS, SWE, SMS inside and outside the basin and the quality of the models for these water balance components. Computation of GWS is independent of the S calculation at basin scale. The derived method has been validated on GW level data over several aquifer systems including California Central valley (Scanlon et al., 2012) and for glacier applications over Tienshan region (Farinotti et al., 2015).

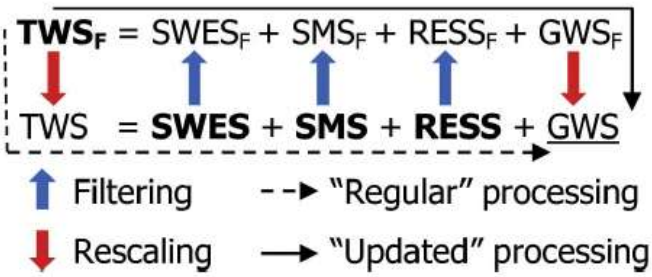

Figure 14 - Synthesis of regular and updated methods for processing GRACE data to extract changes in GWS. Subscript represents spatial filtering, applied equivalently to GRACE and water budget data (ICE, SWS, SWE, SMS). ICE and SWS refer to $RESS$, and SWE to $SWES$ (Longuevergne et al., 2010)

3.1.3. GRACE observations in large aquifers

Considering the spatial resolution of regularized CSR products, large aquifers having an area larger over 100 000 km² based on aquifers delineation provided in the UNESCO-IHP and IGRAC 2015 Transboundary Aquifers Map. Overall, since 2002, most Transboundary Aquifers (TBAs) have had low depletion rates (< 5mm/y). However, there are hotspots in the Middle East and Central Asia with depletion rates higher than 20mm/y (Figure 15). Mass changes over TBAs range from -10 to +15 km3/year and -25 to 25 mm/year. Linear long-term trends, errors, and R^2 and P-value of TBAs are given in Annex 2.

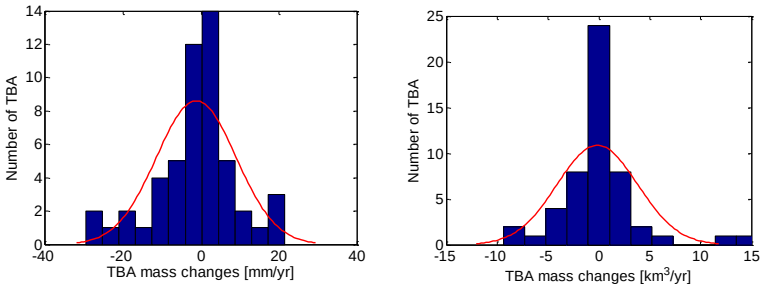

Figure 15 - GRACE data linear depletion rates in Transboundary Aquifers (TBAs)

GRACE observations of the Stampriet Transboundary Aquifer System show a slight mass increase (although uncertainties are quite high) and homogeneous storage changes over a wide region (Figure 16). On the other hand, GRACE observations of the Syr Darya aquifer show small storage changes, which have been affected by large droughts from 2006 to 2008. Storage changes are highly heterogeneous over the region as a whole. For instance, general mass decrease is observed in plains, while general mass increase is observed in mountainous region (Figure 16). The Hindus River Plain Aquifer and the East Ganges River Palin Aquifer show significant and continuous mass decrease. Most storage changes are concentrated in the upper part of the basin, which might be linked to snow pack and significant groundwater pumping in northern India (Figure 16).

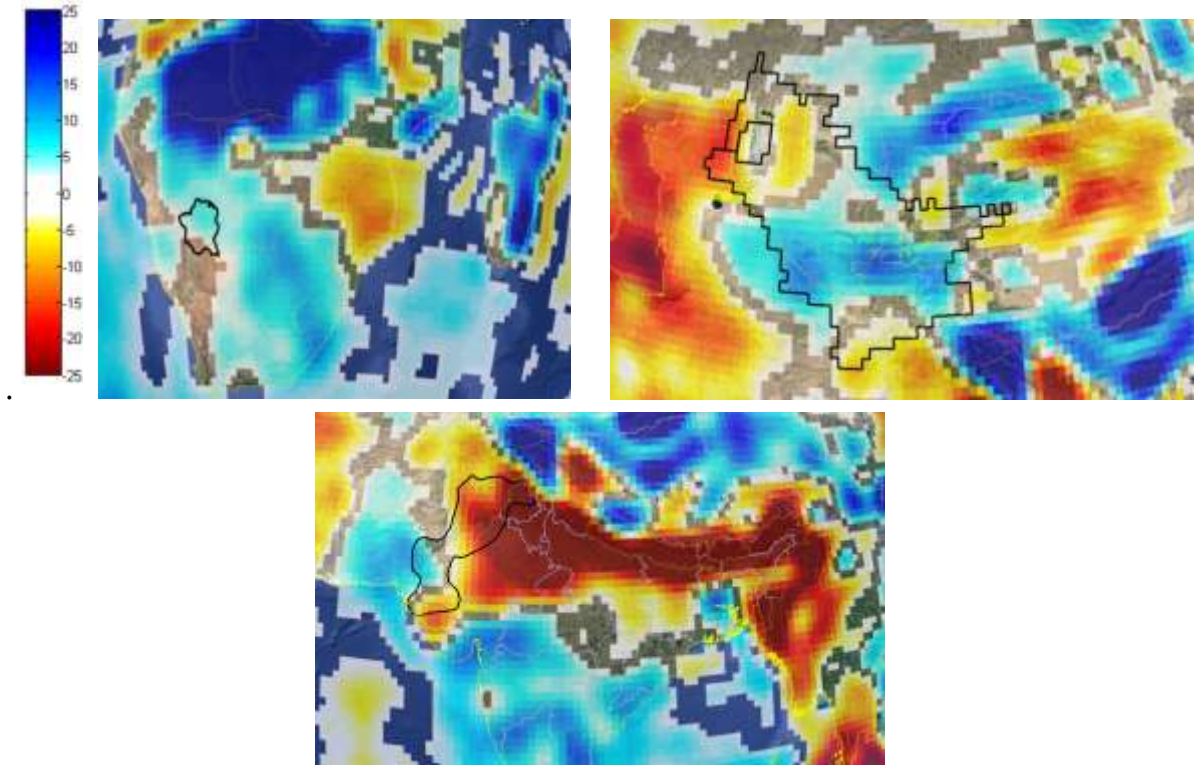

Figure 16 - GRACE observations of the Stampriet Transboundary Aquifer System (top left hand), Syr Darya Aquifer (top right hand), and Hindus River Plain Aquifer and East Ganges River Palin Aquifer (bottom center). Units in mm.

3.2 Limitations of GRACE observations for groundwater storage estimates

GRACE has provided useful information about global groundwater depletion. In particular, the GRACE results have been highly effective in getting large numbers of people to start thinking about groundwater and the sustainability of its use. GRACE's color-coded maps show at a glance where groundwater is being rapidly depleted around the globe. GRACE has proven to be a powerful tool to evaluate groundwater resources. However, as all hydrologic "tools", it has strengths and limitations. This section aims at addressing the main limitations of GRACE for assessing groundwater storage estimates.

GRACE cumulates errors and uncertainties arising from the use of models in the disaggregation process. Therefore, it is likely that GRACE results might promulgate some misperceptions about groundwater resources. GRACE *GWS* changes are likely to result from unconfined aquifers storage changes because storage coefficients in confined aquifers are typically a couple of orders of magnitude less than those in unconfined aquifers (Scanlon et al., 2010a; Scanlon et al., 2010b). Therefore, GRACE is mainly applicable to aquifers with renewable groundwater. Furthermore, leaking of errors and uncertainties from models used in GRACE observations processing leads to difficulties to address GRACE trends for aquifers located close to lakes (Victoria, Tchad, Malawi, Tanganyika, Baikal, Aral Sea) and glaciers.

Alley and Konikow (2015) have recently pinpointed GRACE's main practical limitations for water management policy-making as it provides a one-dimensional (vertical) indicator of the status of a large three-dimensional groundwater body. It does provide a "big picture" view that might be appropriate input for setting broad regional or national policies, and compelling evidence of the need for better groundwater management. However, given that many aquifers that play a critical role in meeting human needs, however, occur at scales of 100s or 1000s of km², much smaller than the GRACE footprint; which substantially hinders GRACE utilization as a widespread local management tool. Additionally, GRACE does not address issues such as saltwater intrusion, land subsidence, how streams and other surface water bodies are being affected by groundwater pumping, and how water quality is changing. GRACE also does not address relevant information to understand the dynamics of aquifers such as it is unable to decipher flow directions or velocities, their changes over time, or actual drawdowns. Additionally, it is worth mentioning that not all groundwater depletion arises from pumping. In some aquifers, long-term drainage and water table declines resulting from climate change—perhaps thousands of years ago—can cause depletion under predevelopment conditions.

3.3 A modelling approach to extend GRACE time frame

While GRACE is the only observation method able to measure groundwater storage changes; its timescale is limited to 10 years, which is below most of internannual to multidecadal climatic oscillations (e.g. ENSO, NAO, PDO, AMO). Consequently, GRACE data analyses would be affected. Additionally, ground-based measurements are limited and when available they are scattered in different timescales that usually fall out of GRACE timescale. Therefore there is a need to go back to the "past" through a model in order to reconstruct groundwater storage changes evolution. Springer et al. (2014) proposed a method to estimate bias in precipitation and evapotranspiration fluxes data using GRACE, allowing a confident estimation of GRACE storage changes over a longer time period. A water mass balance model to estimate long-term groundwater storage dynamics, which independent from GRACE data, was developed and applied at aquifer scale (Figure 17):

$$\frac{dS}{dt} = P - E - R \qquad \text{(Eq. 3)}$$

where P is the precipitation, E the actual evapotranspiration, R the runoff (or the discharge at the outlet of the basin), and S the total water storage (sum of water stored in the vegetation, snow, lakes and rivers, soil moisture, and groundwater). All the components are given in millimeters per month. Mean precipitation and actual evapotranspiration are computed by averaging data available from global datasets with spatial resolutions of 0.5° x 0.5° at aquifer scale (Climate Research Unit database (CRU) and the Global Precipitation Climatology Centre (GPCC) dataset for precipitation, and Max Planck Institute (MPI) dataset for actual evapotranspiration). Runoff data was obtained from the Global Runoff Data Centre (GRDC).

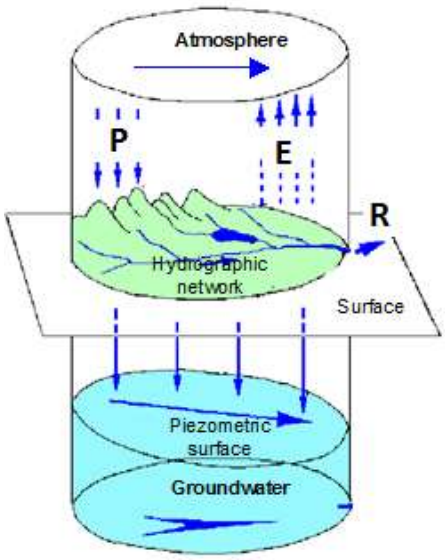

Figure 17 – Aquifers are considered at basin-scale

Integration of Eq. 3 gives:

$$S_{model} = \int P\,dt - \int R dt - \int E\,dt \qquad \text{(Eq. 4)}$$

where S_{model} is the total water storage changes.

Integration over time of P, R, and E will generate a long-term trend in S_{model} which is attributed to the integration of systematic errors of the latter variables. Unbiased precipitation and actual evapotranspiration datasets are therefore required. Estimation of R is challenging because aquifer systems are generally not superimposed with the basins where runoff data would be available. In unconfined aquifers, there is often strong interaction between surface water and groundwater, as groundwater contributes to streams in most physiographic settings (Figure 18). Overlap between surface water and groundwater is considered as the part of the renewable water resources which is common to both surface water and groundwater. It is equal to groundwater drainage into rivers (typically, base flow of rivers) minus seepage from rivers into aquifers. The proportion of stream water that is derived from groundwater inflow varies across physiographic and climatic settings. However, in large aquifers, groundwater contribution to stream flow could be considered as very high (Winter et al., 1998).

In a first step, it is assumed that runoff is constant over time, i.e. contribution to storage as a linear trend, in an equivalent manner as a potential bias in precipitation and evapotranspiration. Therefore, when estimating storage changes by integration of P, R, and E, a long-term trend is removed. While this hypothesis is problematic in overexploited groundwater systems where depletion is known, it is not problematic in semi-arid systems where runoff is generally negligible. Moreover, this work's main objective is to highlight the sensitivity of groundwater storage to interannual climate variability. Thus, Eq. 4 becomes:

$$GWS_{model} \approx S_{model} = \int P\,dt - \int R dt - \int E\,dt \approx \text{detrend}\left(\int P\,dt - \int E\,dt\right) \qquad \text{(Eq. 5)}$$

The simplicity of the proposed model leads to limitations as anthropogenic contribution (e.g. land use change) is indirectly taken into account through actual evapotranspiration. Additionally, abstraction is not considered in the model. Despite the significance, few quantitative assessments of large aquifers exist. A recent study by Richey et al., (2015) revealed a considerable amount of groundwater depletion in large aquifers. However, there is still a need to analyze if groundwater depletion should be mainly attributed to aquifers' overexploitation or other phenomena. Wada et al. (2012) and TWAP (2015) have revealed that many large aquifers over Europe, Asia and Africa are currently not overexploited, although groundwater development stress resulting from the combination of high human dependency, low renewable groundwater per capita, and high abstraction/recharge ratios has been increasing at an alarming rate for the past fifty years. It is projected that new hotspots largely driven by population pressure will develop in the Middle East, Northern and Southern Africa, South and Central Asia.

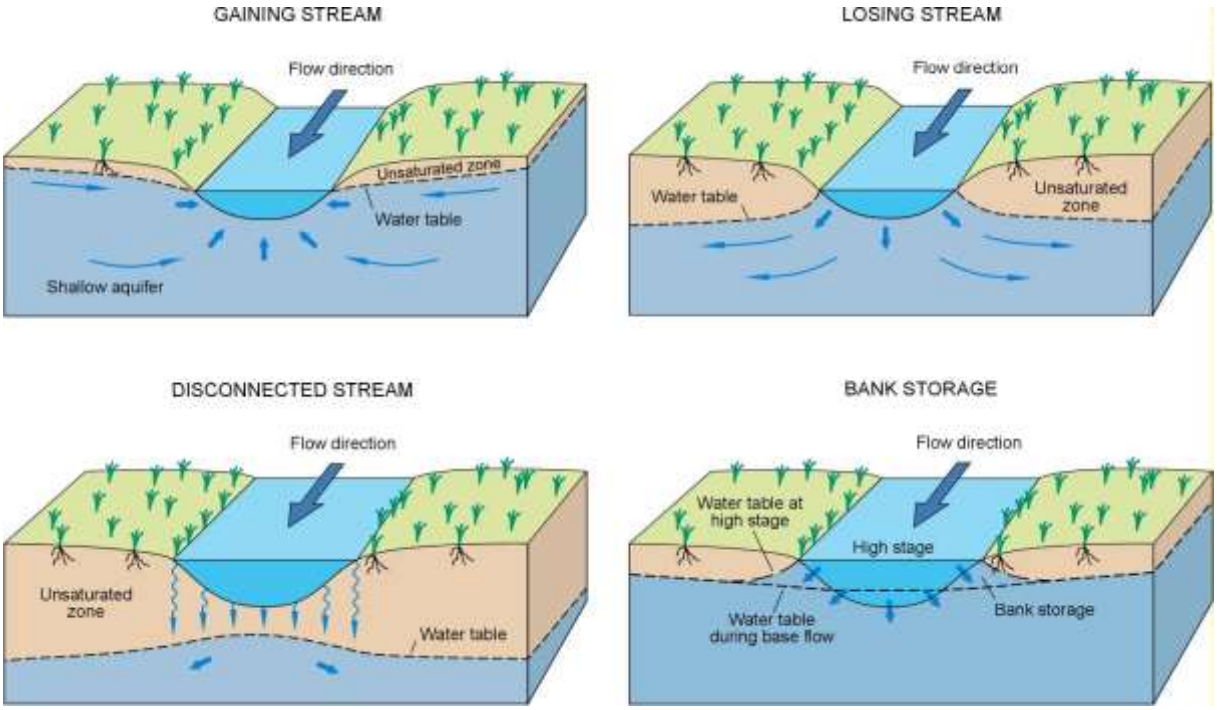

Figure 18 - Surface water and groundwater interaction in unconfined aquifers (Winter et al., 1998)

3.3.1. Dataset model sensitivity analysis and validation

This section aims at analyzing the global datasets used in the proposed model.

Actual evapotranspiration

The Max Planck Institute (MPI) dataset is used. MPI dataset provides up-scaled monthly evapotranspiration in spatial resolutions of 0.5° x 0.5° from 1982 to 2011. It is based on

empirical up-scaling of FLUXNET data from the LaThuile synthesis effort (www.fluxdata.org). The global FLUXNET network provides continuous in situ measurements of land-atmosphere exchanges (including water vapour), and this data can be used to estimate global land actual evapotranspiration (*E*) dynamics. Jung et al., 2010 have designed an approach to assessing the temporal behavior and global spatial distribution of *E* since 1982. It integrates point-wise *E* measurements at the FLUXNET observing sites with geospatial information from satellite remote sensing and surface metereological data in a machine-learning algorithm, the Model Tree Ensemble (MTE). *E* data was validated by internal cross-validation at FLUXNET sites, corroboration against independent ET estimates from 112 catchment water balances, and the simulations of 16 landsurface models participating in the Global Soil Wetness Project 2 (GSWP-2). The station based and annually resolved runoff data originate from the Global Runoff Data Centre (GRDC).

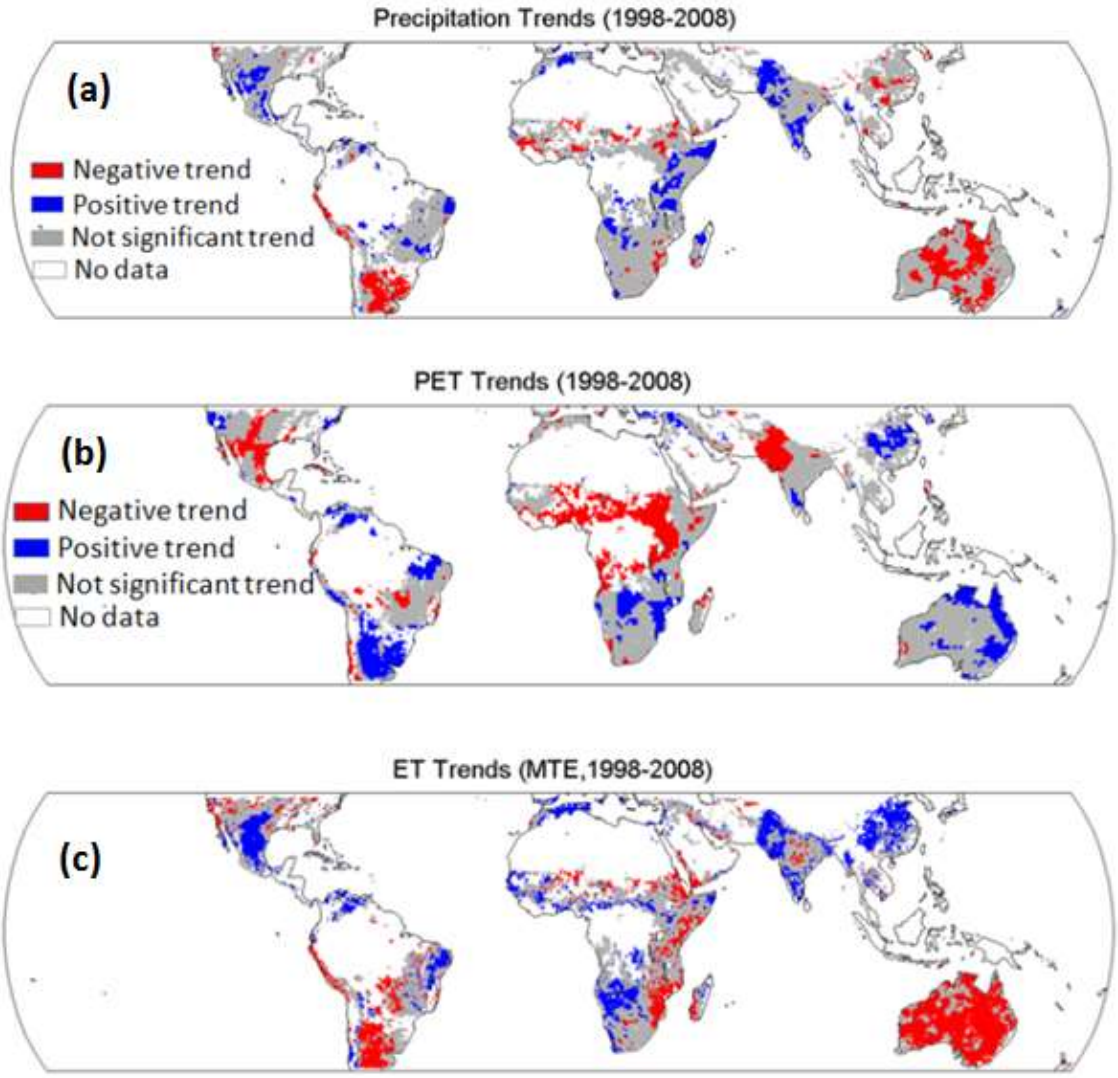

Figure 19 - Trends from 1998 to 2008 of (a) precipitation (P), (b) potential evapotranspiration (PET) and (c) actual evapotranspiration (ET – referred as *E* in this work) (Jung et al., 2010)

Overall, trends of precipitation are also consistent with trends in actual evapotranspiration (*E*). According to Jung et al., 2010, global annual evapotranspiration increased on average by 7.16 ± 1.0 millimeters per year per decade from 1982 to 1997. After that, coincident with the last major El Nino event in 1998, the global evapotranspiration increase seems to have ceased until 2008. This change was driven primarily by moisture limitation in the Southern Hemisphere, particularly Africa and Australia. In these regions, microwave satellite observations indicate that soil moisture decreased from 1998 to 2008.Trends in atmospheric demand assessed with potential evapotranspiration are in the opposite direction to trends in actual evapotranspiration, except in China and southern India, where potential and actual evapotranspiration both exhibit positive trends.

Precipitation

Two global precipitation databases are used, namely the Climate Research Unit database (CRU) and the Global Precipitation Climatology Centre (GPCC) database. Both datasets provide gauge-based gridded monthly precipitation for the global land surface from 1901 to 2011, and are available in spatial resolutions of 0.5° x 0.5° based on analyses of over 4000 weather stations for CRU (CRU, 2015) and about 67,200 stations for GPCC (Rudolf et al., 2011).

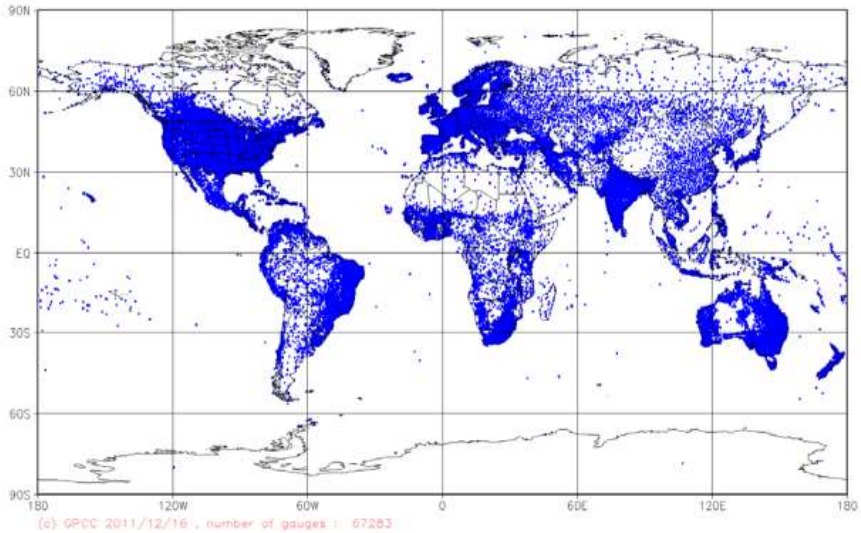

Figure 20 - Spatial distribution of monthly in-situ station with a climatological precipitation normal, based on at least 10 years of data in GPCC database (number of station in July 2011: 67 283)

A recent study by Hwang et al., 2013 has shown that GPCC and CRU datasets agree with the drying rend in Sahel, Venezuela over northern edge of South America, and the moistening trend over eastern Brazil. However, some regional discrepancies between both datasets might be found. For example, GPCC reports a significant decrease in rainfall over Indonesia and western Brazil, which are less apparent in CRU, whereas CRU reports a significant increase in rainfall over northern Australia, which is less apparent in GPCC.

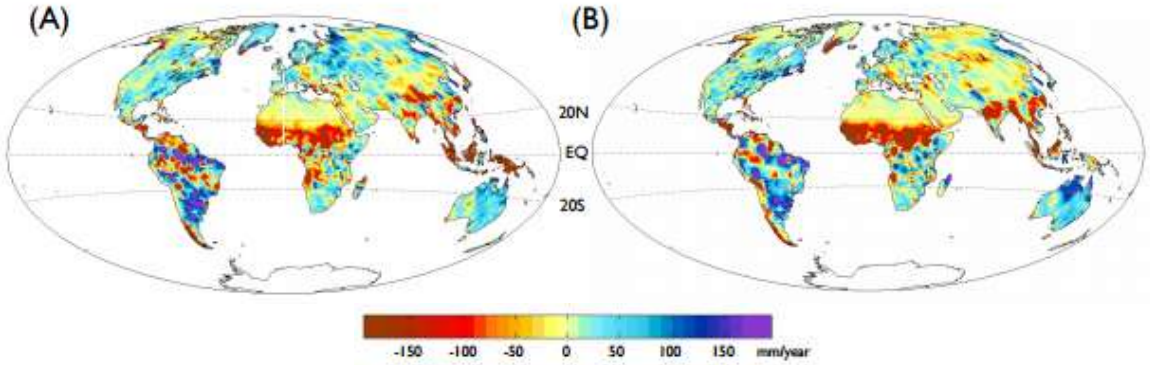

Figure 21 - Changes in precipitation from 1931~1950 to 1971~1990 based on (a) GPCC and (b) CRU (Hwang et al., 2013)

3.3.2. Comparison at regional scale

The Stampriet Transboundary Aquifer System (STAS) was chosen to analyze and compare GPCC and CRU data with precipitation ground measurements[1]. GPCC and CRU datasets use the same stations and ground measurements are available for four stations (Ncojane, Tshane, Tsabong, and Gemsbok Park – Twee Rivieren). It is found that global datasets data and ground observation generally have an identical long-term annual mean (within 2%), however, the rainfall sequence is generally much better represented by GPCC dataset (r^2=0.97) compared to CRU data (r^2=0.57) (Figure 23). Therefore, we choose GPCC rainfall product in the modeling exercise. Similar results have been found over mountainous regions of Africa (Dinku et al., 2007). It is to note that GPCC generally provides larger interannual variability in precipitation as compared to CRU. As a consequence, long-term modeled storage contains higher long-term variability when using GPCC dataset. Over the Stampriet Transboundary Aquifer system, obtained results show that GPCC dataset with a larger interannual variability in precipitation leads to a better prediction of interannual groundwater storage changes (Figure 24). This shows the importance of rainfall data for modeling long-term variations.

[1] It is worth mentioning that the computation of the Stampriet Transboundary Aquifer System (STAS) mean precipitation is based on the datasets spatial resolution of 0.5° x 0.5°, and consequently also takes into account stations that are located outside the aquifer boundary.

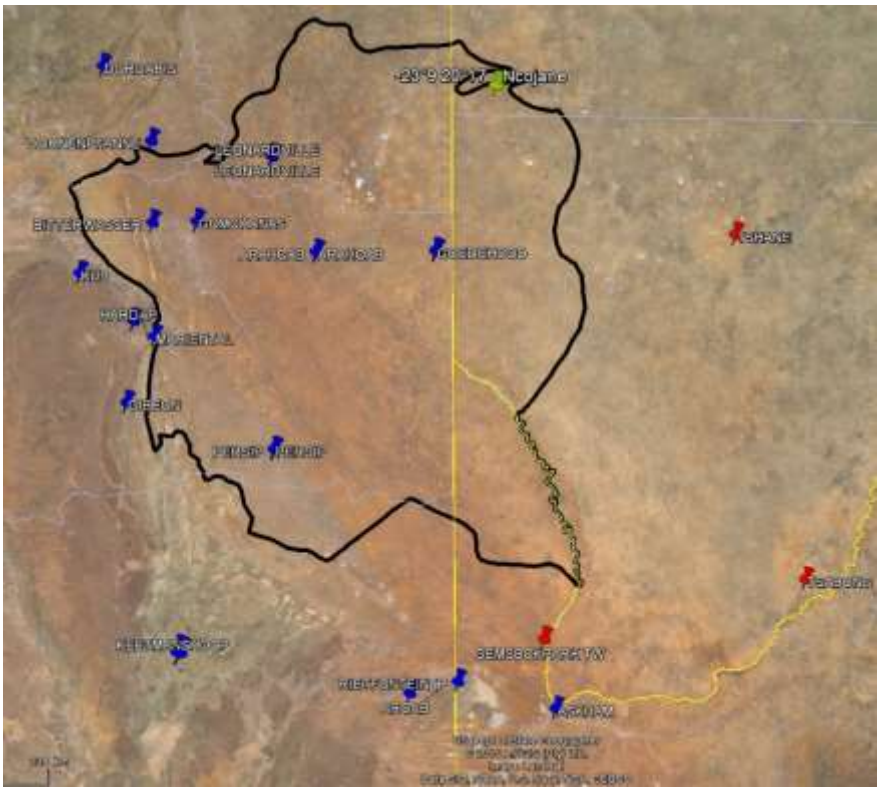

Figure 22 - GPCC and CRU stations considered to compute the Stampriet Transboundary Aquifer System (STAS) mean precipitation.

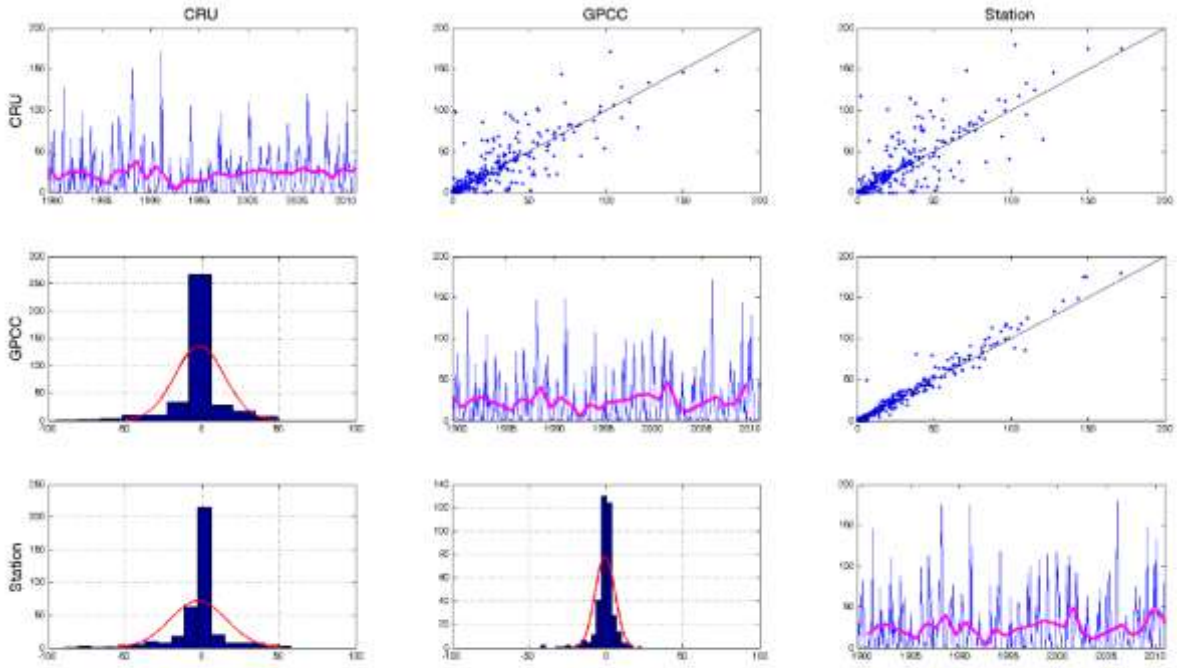

Figure 23 - Analysis of precipitation datasets (GPCC and CRU) and ground measurements for the Tsabong station (Botswana)

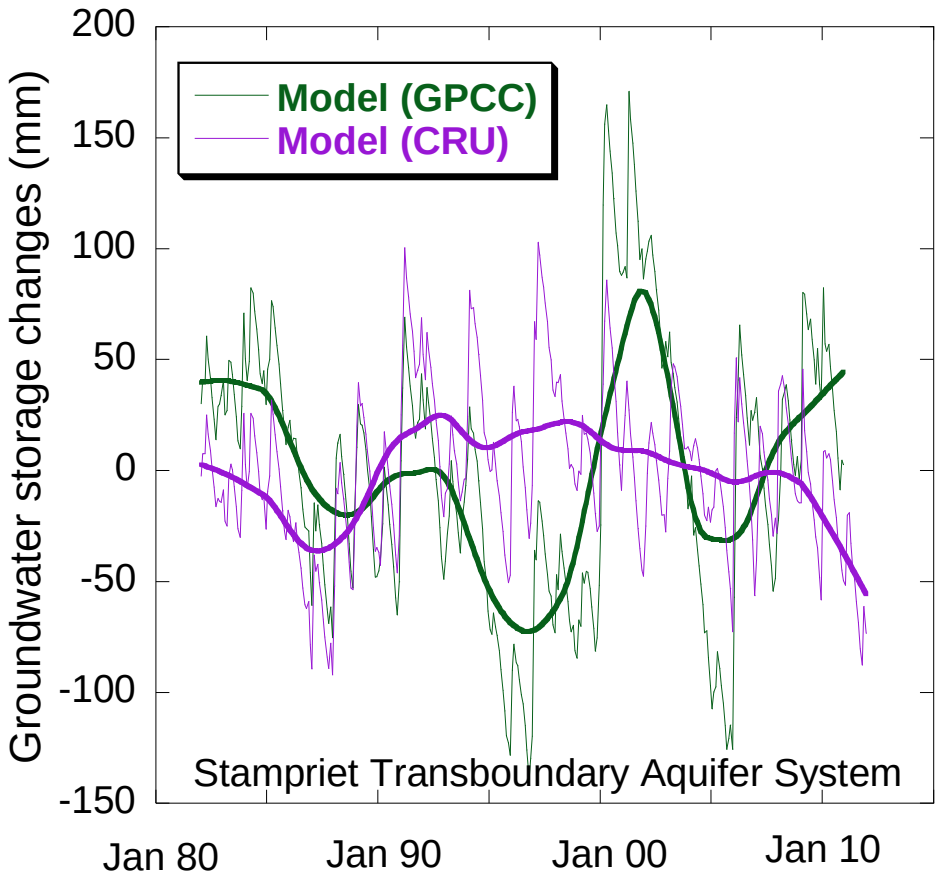

Figure 24 - Comparison of experimental and simulated groundwater storage changes in the Stampriet Transboundary Aquifer System (STAS) using GPCC and CRU databases. Dark lines were obtained using the locally weighted Least Squared error (Lowess) method.

4. Results and discussion

GRACE data show that *GWS* depletion rates from 2002 to 2011 are negative in Northern America (-16 mm/y in the Ogallala Aquifer), Central (-2.2 mm/y in the Syr Darya Aquifer) and Southern Asia (-6 mm/y in the East Ganges River Plain Aquifer) aquifers, and positive in Southern Africa (8 mm/y in the Karoo Sedimentary Aquifer and 5.3 mm/y in the Stampriet Transboundary Aquifer System) and the Sahel (3.5 mm/y in the Irhazer-Illuemeden Basin Aquifer) aquifers. However the latter findings could be misleading to draw any exhaustive conclusion about *GWS* depletion trends because of the limited 10-year time scale, and the effects of climate variability.

Thus, this section presents an analysis of GRACE, modeled and experimental groundwater storage changes in the studied aquifers. Experimental *GWS* changes were calculated from well data by converting water level changes to water volumes using porosity. Findings presented here indicate that *GWS* changes from GRACE data, model and detailed groundwater level monitoring are in satisfactory agreement.

Studies indicate that the ocean-atmosphere oscillation patterns of ENSO, PDO, and AMO and associated hydroclimatic variability may affect aquifers recharge rates and groundwater level variability (Gurdak et al., 2007 and Taylor et al., 2013). Therefore, GRACE data should only be used as a mean to monitor basin-scale changes in *GWS*, and not as a mean to draw any exhaustive conclusion about long-term depletion rates trends and the impact of climatic oscillations cycles on groundwater resources. In order to circumvent GRACE's timescale limitation, groundwater storage changes have been reconstructed by the model developed in this study. Obtained results partially indicate that the ocean-atmosphere oscillation patterns of ENSO, PDO, and AMO and associated hydroclimatic variability affect groundwater storage changes in all the studied aquifers. We find that such changes are highly correlated with rainfall-fed recharge and might suggest that groundwater level variability on interannual to multidecadal timescales might be 1) a response to ENSO, PDO, and AMO, 2) and not solely influenced by temporal trends in groundwater pumping (Gurdak et al., 2007; Taylor et al., 2012). Results indicate that ENSO and PDO have a greater control than AMO on variability in groundwater storage changes for the studied aquifers, except for aquifers located in the Sahel region (e.g. Irhazer-Illuemeden Basin Aquifer).

In Southern African aquifers (Karoo Sedimentary and Stampriet Transboundary Aquifer System), obtained results indicate that groundwater storage changes might be highly correlated to ENSO Index. When El Niño (La Niña) events occur, more intense droughts (floods) and below (above) normal recharge is likely to occur in Southern Africa aquifers. For instance, a steady increase of groundwater level was observed during the early 2000s La Niña years which lead to above normal precipitation and recharge (Figure 25 and Figure 26). Such results concur with Maidment et al., (2015) results which suggest that the trend to more La Niña-like conditions since 2000 is a likely contributing factor driving the increase in Southern Africa rainfall[2]. Additionally, it is assumed that obtained groundwater storage changes in the two studied Southern African aquifers are not driven by temporal trends in groundwater pumping because human groundwater use in both aquifers is considered as very low to low[3] (TWAP, 2015).

[2] Taylor et al., (2013) have shown strong evidence of the dependence of groundwater resources on extreme rainfall in East Africa, as five out of the seven largest recharge events coincided with El Niño events

[3] According to TWAP (2015), a hotspot is considered when groundwater development stress (equal to total groundwater abstraction divided by long-term mean annual groundwater recharge) is higher than 40%

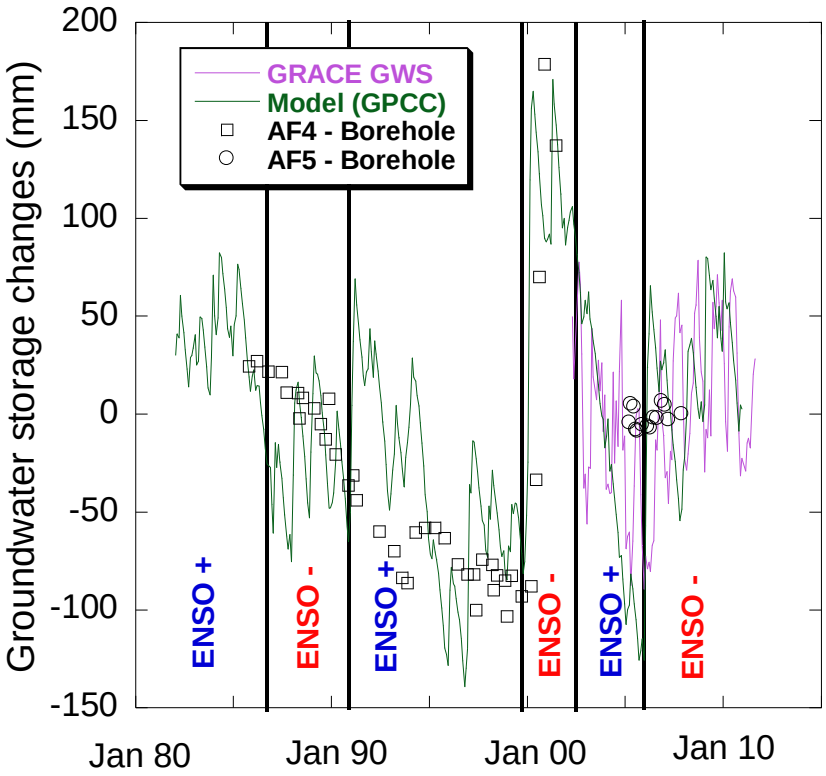

Figure 25 - Comparison of GRACE, modeled and experimental groundwater storage changes in the Stampriet Transboundary Aquifer System

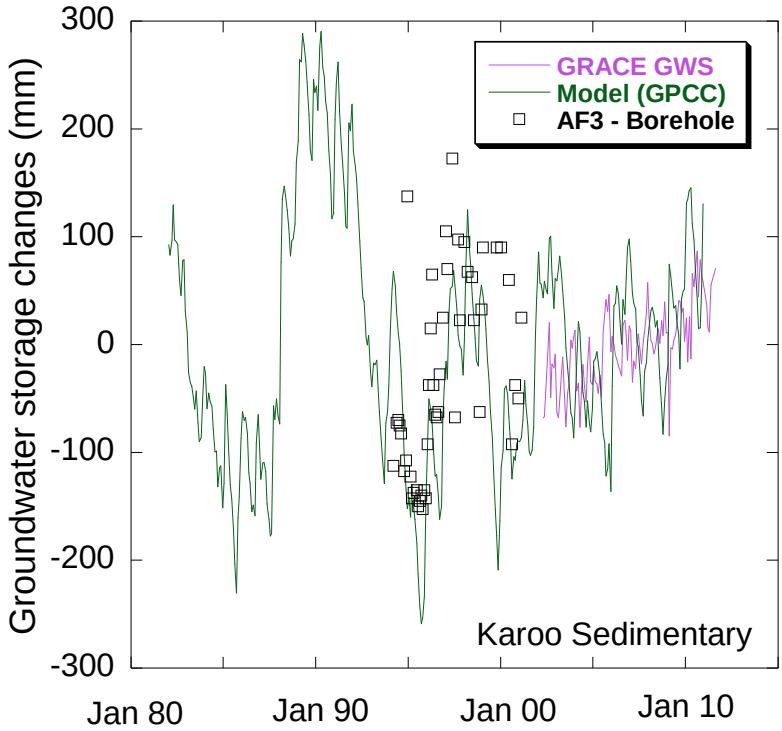

Figure 26 - Comparison of GRACE, modeled and experimental groundwater storage changes in the Karoo Sedimentary Aquifer

In the Ogallala Aquifer, obtained results concur with previous studies (Gurdak et al., 2006 and Gurdak et al., 2007) and indicate that groundwater level variations might be correlated to PDO Index (Figure 27). Gurdak et al., 2007 suggest that the controlling cycles of groundwater-level variability in the Ogallala Aquifer are driven by PDO-like climatic factors and not solely influenced by pumping variability. The majority of the variance in the Ogallala Aquifer precipitation time series was attributed to >PDO (>25 yr) and PDO periods (10–25 yr), with lesser control by ENSO. Additionally, the strength and direction of correlation between >PDO and PDO cycles and precipitation variability is consistent with the findings from McCabe et al. (2004), which indicate the majority of variance in drought frequency across the conterminous USA is attributed to positive AMO and negative PDO phases of variability. The climatic variability that dominantly controls the groundwater levels was also successfully identified by Gurdak et al., 2007. Even though groundwater of the Ogallala Aquifer is influenced by pumping, groundwater-level variability response to pumping was quantified as moderate in ENSO periods. The lack of PDO periodicities within the groundwater pumping records indicates that anthropogenic responses are PDO independent. Additional evidence demonstrated by comparing pumping responses with precipitation and climate indices, identified a rapid anthropogenic response to wet–dry fluctuations in shorter-term climate variability (e.g. ENSO).

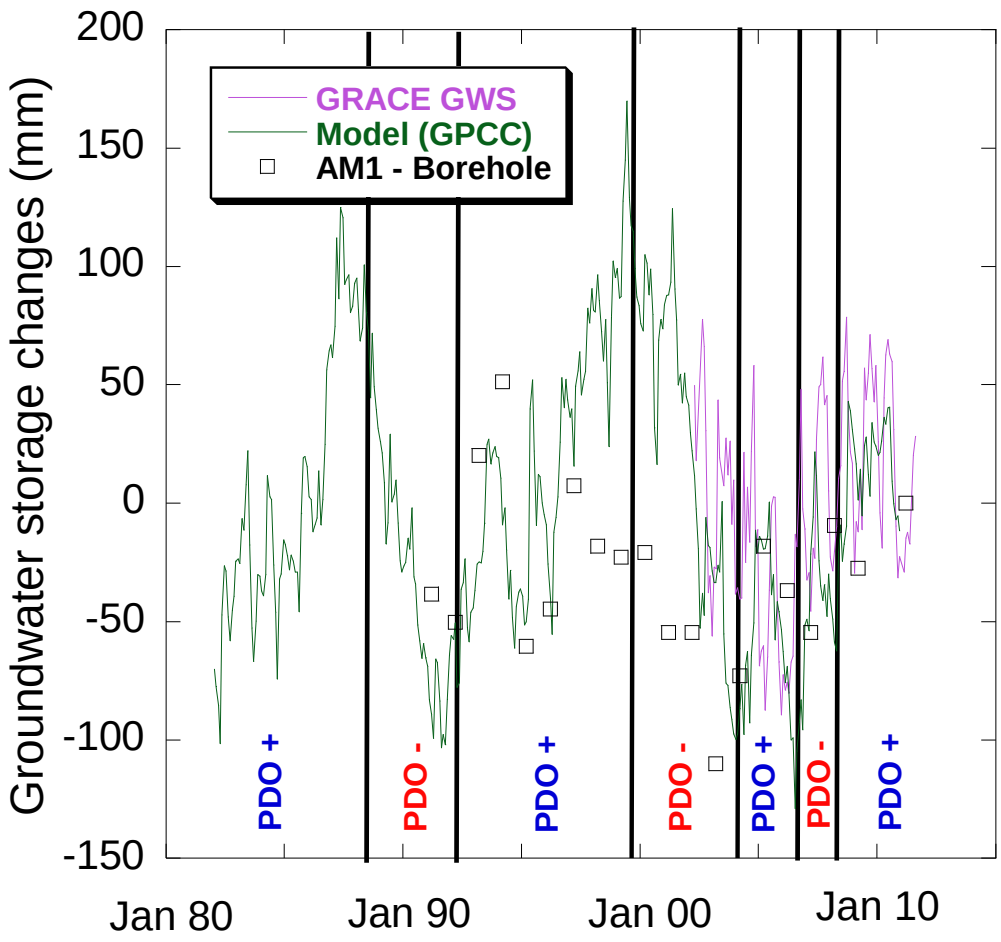

Figure 27 - Comparison of GRACE, modeled and experimental groundwater storage changes in the Ogallala Aquifer

Although, the model developed in this study does not take into account pumping, modelled groundwater-level variability in all studied aquifers is in good agreement with available ground-based observations. However, it is worth mentioning that in drought periods, obtained modeled results do depict decreasing groundwater storage, but are not as correlated to ground-based observations as in wetter periods (e.g. in the Ogallala Aquifer from early-2000s to mid-200s, and in the STAS from early-1990s to late-1990s). This result might be explained by the fact that during drought periods, reliance on groundwater is higher, and consequently there might be a tendency to over-pumping.

Shamsudduha et al. (2009) studied the trends in shallow groundwater levels within the East Ganges River Plain Aquifer, and have shown that rising mean groundwater levels could occur in some heavily abstracted areas (e.g. north of Dinajpur) (Figure 8 and Figure 29). Given that the predominant source of natural recharge of the East Ganges River Plain Aquifer is mainly through heavy monsoon precipitation over the aquifer area, projected scenarios of groundwater level would require thorough precipitation forecasting modelling. Forecasting monsoon precipitation is extremely complex, but studies have shown that the relationship between ENSO and Indian monsoon is not straightforward (Kumar et al., 1999 and Kumar et al., 2006). Recently, Krishnamurthy and Krishnamurthy (2013) have shown that the monsoon-ENSO relation is modified according to the phase of the PDO. Whether El Niño events bring more severe or less severe drought conditions over India may depend on whether the PDO is in warm or cold phase. It could then be expected that when warm (cold) phase of PDO and El Niño (La Niña) co-occur, more intense droughts (floods) and below (above) normal recharge are likely to occur over the East Ganges River Plain Aquifer. Obtained results in the East Ganges River Plain Aquifer concur with Krishnamurthy and Krishnamurthy as modeled groundwater storage changes substantially decrease when warm PDO phase and El Niño co-occur. Results are in good agreement with available ground-based observations from the late 1980s and early 2000s (Figure 28). This result might suggest that the controlling cycles of groundwater-level variability may be driven by climatic factors (i.e. precipitation) rather than pumping variability. Similar results are found for the Syr Darya Aquifer (Figure 30). However, it is worth mentioning that the relationship between PDO and hydroclimate in Central Asia is more robust than in monsoonal Asia, especially in warm periods (Fang et al., 2014).

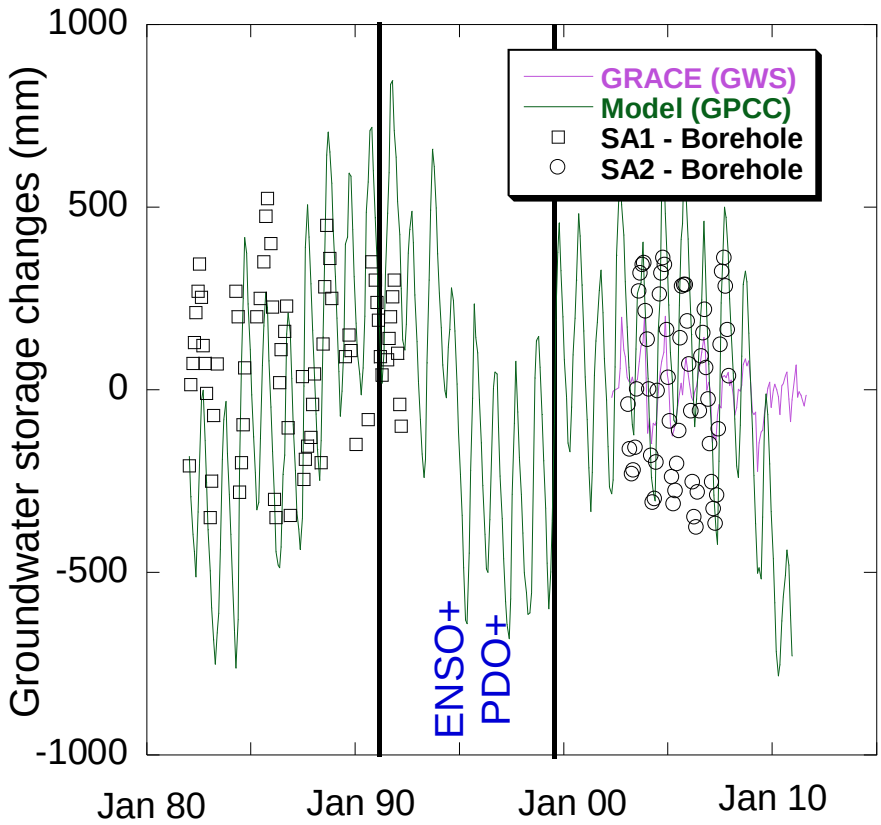

Figure 28 - Comparison of GRACE, modeled and experimental groundwater storage changes in the East Ganges River Plain Aquifer

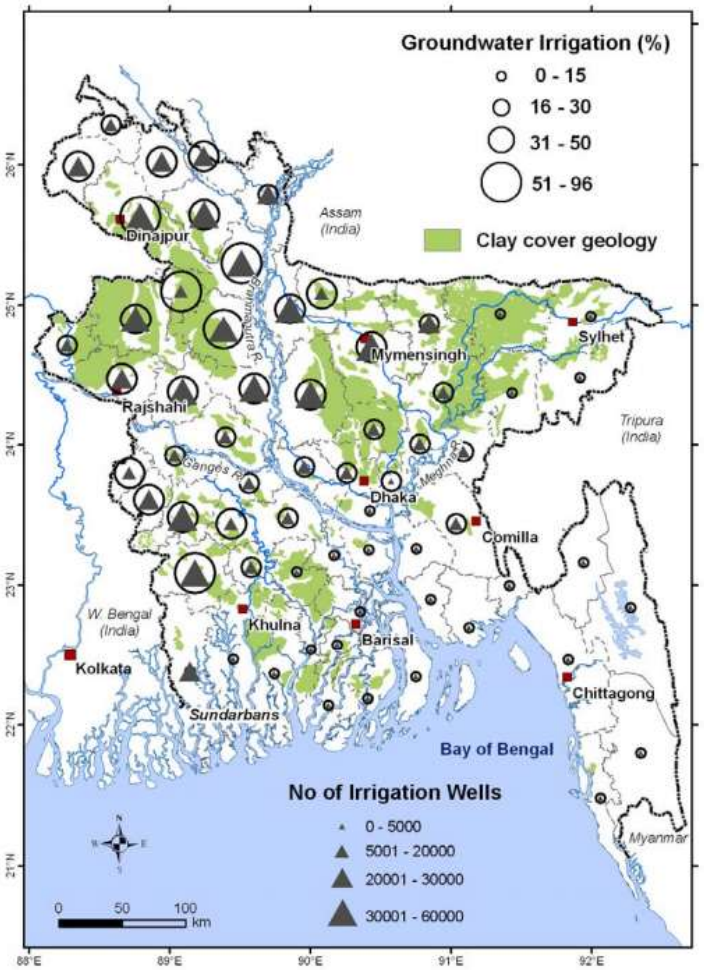

Figure 29 - Percentage of land in each of the 64 districts (broken gray lines) in Bangladesh irrigated with groundwater in 2003 (BADC, 2003). Total numbers of shallow and deep tubewells operated in each district in 2003 are also shown. Low-permeable regionally extensive surface geological units are shown in the background (Shamsudduha et al., 2012)

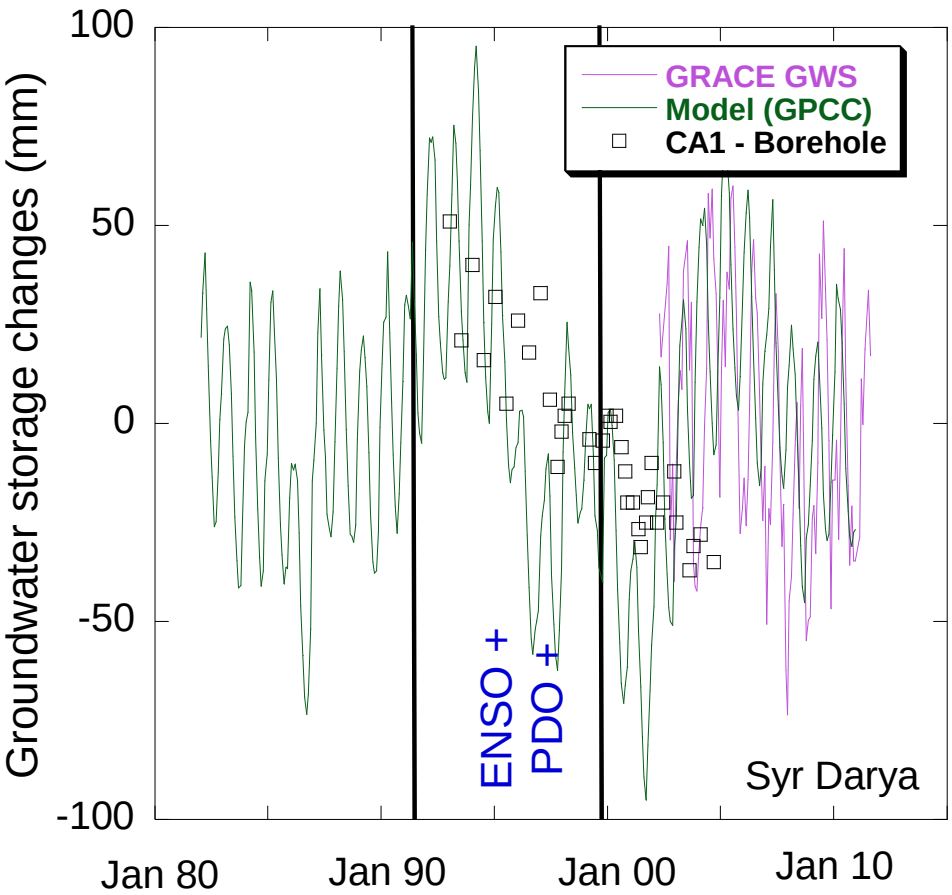

Figure 30 - Comparison of GRACE, modeled and experimental groundwater storage changes in the Syr Darya Aquifer

Correlation of groundwater storage changes with AMO in the Irhazer-Illuemeden Basin Aquifer might be more straightforward than for the other studied aquifers. A warm AMO phase is thought to be associated with a positive precipitation anomaly in the Sahel region (Folland et al. 2001, Knight et al. 2006, Martin and Thorncroft, 2014), which coincides with the generally monotonic increase of groundwater levels since mid-1990s (last AMO shift from cold to warm phase). Leduc et al. (1997) and Favreau et al., (2009) reported a water table rise from early-1930s to late 1960s (AMO warm phase), followed by a drop during the drought of mid-1970s and 1980s throughout the Sahel (AMO cold phase). It is worth mentioning that although the mid-1960s was wetter than today in the Irhazer-Illuemeden Basin Aquifer area, groundwater levels were the same or lower than those currently observed (Leduc et al., 1997). Several studies have attributed water level rise since mid-1990s in the Irhazer-Illuemeden Basin Aquifer as a consequence of land-use changes, such as native vegetation clearing (Leduc et al., 2001; Reij et al., 2005; Favreau et al., 2009). However, based on potential AMO-rainfall-groundwater storage changes correlation, this work tends to suggest that land-use change contribution is minimal with regards to the impact of variations in rainfall as groundwater storage changes decrease (increase) seem to be highly correlated with negative (positive) AMO (Figure 31).

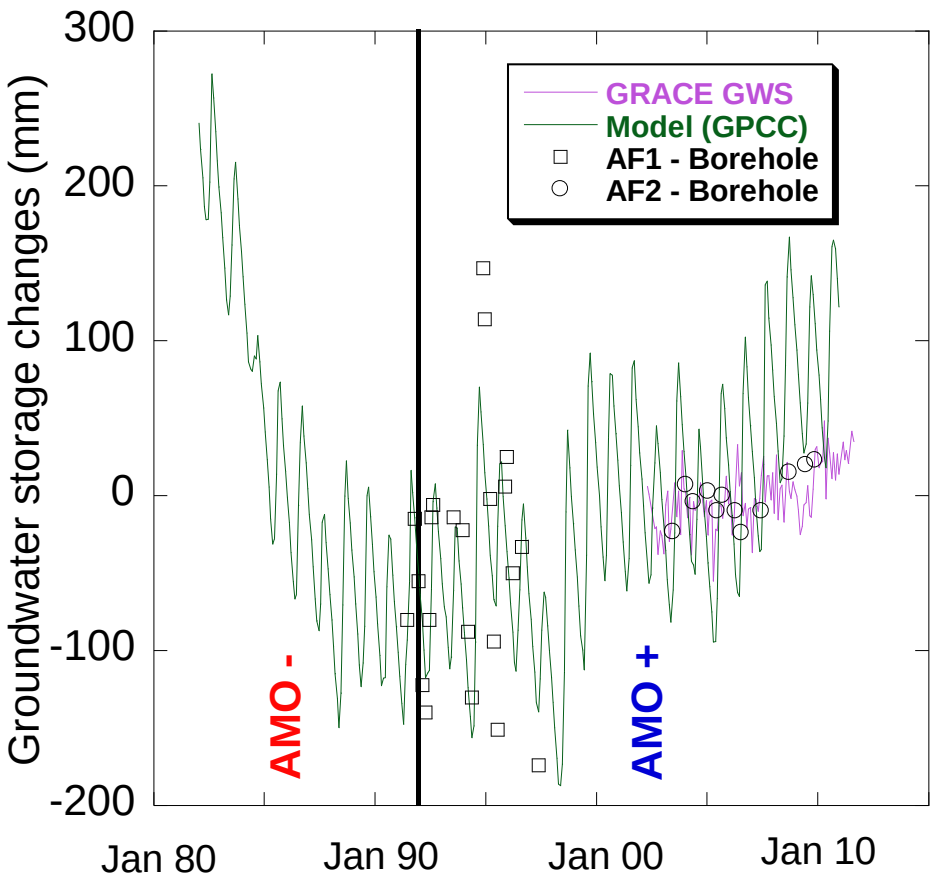

Figure 31 - Comparison of GRACE, modeled and experimental groundwater storage changes in the Irhazer-Illuemeden Basin Aquifer

5. Conclusion

Policy making process is lengthy, especially in the field of natural resources, because of existing uncertainties. This is particularly true for groundwater resources management. Indeed, despite its importance, groundwater is still poorly understood and often undervalued. This may in part derive from the nature of groundwater: a complex, hidden resource that is difficult to conceptualize. Satellite observations have revolutionized our understanding of water availability at continental scale. GRACE data has proven to be an extremely powerful tool to monitor large aquifers changes in *GWS*, which complement much higher spatial-resolution *GWS* changes from ground-based monitoring and modeling analyses. It is now known that global change impacts groundwater availability and quality (Green et al., 2011, Treidel et al., 2012, Taylor et al., 2013). However, for the time being, GRACE data does not allow drawing any exhaustive conclusion about the impact of climatic oscillations cycles on groundwater resources because of its limited timescale of about 10 years. Additional decades of GRACE observations combined with ground-based data (or more) would be required to thoroughly evaluate the impact of climatic oscillations cycles on groundwater resources in large aquifers. However, given that most of large aquifers are located in highly populated areas and have great regional social and economic importance, there is an urgent need to expand the dynamic and usable "toolbox" of innovative approaches (McNeeley et al., 2012) that can be used by water managers and scientists to improve water resource planning and operations in the context of the current scenario of limited data and future climate uncertainty (Milly et al., 2008, Voss et al., 2013). Innovative approaches also have to consider the existing capacity and knowledge gap that exists between developed and developing countries. For instance, processing of satellite data is costly and would require enormous capacity training in developing countries. This work has developed a simple model based on the water mass balance equation that has allowed to satisfactorily reconstructing groundwater fluctuations since 1982 in several large aquifers in North America, Southern Africa, Central and South Asia. The simplicity of the model developed in this work provides a powerful tool that could be easily replicated and applied in regions with knowledge and capacity gaps. Obtained results might indicate that substantial variability in groundwater storage changes (and consequently groundwater levels) on interannual to multidecadal timescales might be more correlated to climatic factors (e.g. ENSO, PDO, and AMO) than temporal trends in groundwater pumping, even in over-exploited aquifers such as the Ogallala and East Ganges River Plain Aquifer. Thus, a better understanding on the impacts of climatic oscillations on rainfall will allow better predictions of groundwater levels fluctuations. Water decision makers can begin to use information presented here to:

1) strengthen links between Department of Water Affairs and Meteorological Agencies (at national, and regional level) as groundwater levels variability might be more correlated to precipitation variability than abstraction variability.

2) help identify locations and optimal time periods for water storage and managed artificial recharge (MAR) operations in accordance to ENSO, PDO, and AMO phases, and

3) consider updating groundwater abstraction permits in accordance to ENSO, PDO, and AMO phases.

A key limitation of the model developed in this work is that anthropogenic contribution (e.g. land-use change) is indirectly taken into consideration through evapotranspiration. Therefore, future work should focus on extracting the contribution of climatic factors and anthropogenic activities, especially in overexploited aquifers such as the Ogallala and East Ganges River Plain.

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Annex 1 – Groundwater level fluctuations

Borehole ID	Aquifer	Latitude	Longitude	Location	Time frame	Source
Americas						
AM1	High Plains	NA	NA	Averaged from approximately 9000 wells across the High Plains	1990-2013	McGuire, 2014
Africa						
AF1	Irhazer-Illuemeden Basin			Maradi	1991-1998	
AF2	Irhazer-Illuemeden Basin			Koutakroukou	2003-2012	
AF3	Karoo Sedimentary	28.3	-28.8	Caledonspoort	1994-2001	GGMN, 2015
AF4	Stampriet Transboundary Aquifer System	-23.7	18.4	Stampriet	1986-2001	JICA, 2002
AF5	Stampriet Transboundary Aquifer System	-24.5	18.2	Ncojane	2005-2007	Matsheng, 2007
Central Asia						
CA1	Syr Darya					
South Asia						
SA1	East Ganges River Plain	23.7	89.4	Sarsa	1982-1992	BGS and DPHE, 2001
SA2	East Ganges River Plain	NA	NA	Averaged from a total of 236 monitoring locations and river levels averaged from a total of 298 gauging stations across Bangladesh	2003-2008	Shamsudduha, 2011

Annex 2 – GRACE Linear long-term trends, errors, and R^2 and P-value of TBAs

TBA name	Estimated trend [mm/year]	1 sigma error [mm/year]	Estimated trend [km3/year]	1 sigma error [km3/year]	R^2	P value
Amazonas	4.2	0.3	15.6	1.2	0.63	1.91E-24
Amu-Darya	-12.8	0.7	-2.4	0.1	0.75	2.88E-33
Aquifer Vallee de la Benoue	17.3	0.6	3.3	0.1	0.9	2.57E-54
Aquifer extension Sud-Est de Taoudeni	1.1	0.3	0.4	0.1	0.1	0.001137
Baggara Basin	-1.4	0.6	-0.3	0.1	0.11	0.00039
Cambodia Mekong River Delta Aquifer	7.6	1.6	1.4	0.3	0.23	1.61E-07
Coango	0.2	0.5	0.1	0.1	0	0.93871
Cuvette	-8.3	1.3	-6.6	1	0.11	0.000513
East Ganges River Plain Aquifer	-4	0.3	-0.7	0.1	0.59	5.02E-22
Edwards-Trinity-El Burro	-3	0.5	-0.3	0.1	0.35	1.97E-11
Indus River Plain Aquifer	-8.5	0.4	-3	0.2	0.82	5.88E-41
Irhazer-Illuemeden Basin	3.2	0.2	1.6	0.1	0.63	1.98E-24
Irtysh-Obsky	2.3	0.3	2.1	0.3	0.3	8.94E-10
Jezira Tertiary Limestone Aquifer System	-17.5	1	-2.3	0.1	0.75	1.15E-33
Judith River	14	0.8	2.3	0.1	0.78	5.34E-37
Karoo Sedimentary Aquifer	8.6	0.2	1.2	0	0.93	1.66E-64
Karoo-Carbonate	-5.5	0.7	-3	0.4	0.33	7.62E-11
Khorat Plateau Aquifer	3.2	0.5	0.3	0.1	0.28	3.25E-09
Lake Chad Basin	-4.8	0.2	-9.4	0.5	0.81	3.26E-40
Lower Mekong River 2 Aquifer	1.7	1	0.2	0.1	0.03	0.05611
Merauke - Ketu Basin Aquifer	-19	0.8	-4.1	0.2	0.82	1.38E-41
Middle Heilongjiang - Amur River Basin	2.7	0.4	0.3	0	0.34	4.79E-11
Northern Great Plains	6.8	1.1	5.3	0.8	0.25	2.80E-08

Northwest Sahara Aquifer System (NWSAS)	-0.5	0.3	-0.6	0.3	0.14	5.41E-05
Nubian Sandstone Aquifer System (NSAS)	0.2	0.2	0.4	0.4	0.02	0.12275
Pantanal	-2.8	0.5	-0.6	0.1	0.15	4.62E-05
Península de Yucatán-Candelaria-Hondo	-6.4	0.4	-0.9	0.1	0.67	1.58E-27
Pre-Caspien	-5.8	0.8	-1	0.1	0.32	2.18E-10
Saq Ram	-8.6	0.2	-1.7	0	0.94	2.21E-68
Senegalo-Mauretanian Basin	4	0.4	1.2	0.1	0.41	1.12E-13
Sistema Aquífero Guaraní	10	0.4	12.2	0.5	0.87	5.98E-48
South of outer Himalayas Aquifer	-28.5	1.2	-9.1	0.4	0.86	3.95E-47
South_pred_ural aquifer	1.9	0.5	0.2	0	0.24	7.09E-08
Stampriet Transboundary Aquifer System	5.4	0.3	0.5	0	0.69	4.71E-29
Sudd Basin	1.4	0.5	0.5	0.2	0.04	0.044453
Syr Darya	-2.7	0.4	-0.8	0.1	0.33	9.83E-11
Syrt	-9.8	0.9	-1.5	0.1	0.51	5.76E-18
Système Aquifère de Tindouf	-2.2	0.4	-0.4	0.1	0.3	6.60E-10
Tanganyika	-29.5	1.6	-5	0.3	0.76	4.76E-35
Taoudeni Basin	-2	0.2	-2.2	0.2	0.59	1.57E-22
Taurus-Zagros	-21.8	0.8	-4.4	0.2	0.86	1.10E-47
Umm_north	-6.4	0.5	-1.9	0.1	0.65	5.52E-26
Umm_south	-2.7	0.1	-3.3	0.2	0.65	6.58E-26
Volta Basin	19.5	0.8	2.5	0.1	0.83	4.58E-42
Voronka groundwater	21.4	1	2.1	0.1	0.81	6.76E-40
Weathered basement	-3.5	0.9	-0.4	0.1	0.19	2.03E-06
Yenisei Upstream	-2.5	0.6	-0.3	0.1	0.15	3.23E-05
Yrendá-Toba-Tarijeño	6.6	0.3	3.1	0.2	0.78	9.11E-37
Zeya River Basin	9.7	0.5	0.7	0	0.83	6.00E-43

Annex 3 – GPCC / CRU datasets comparison at local scale

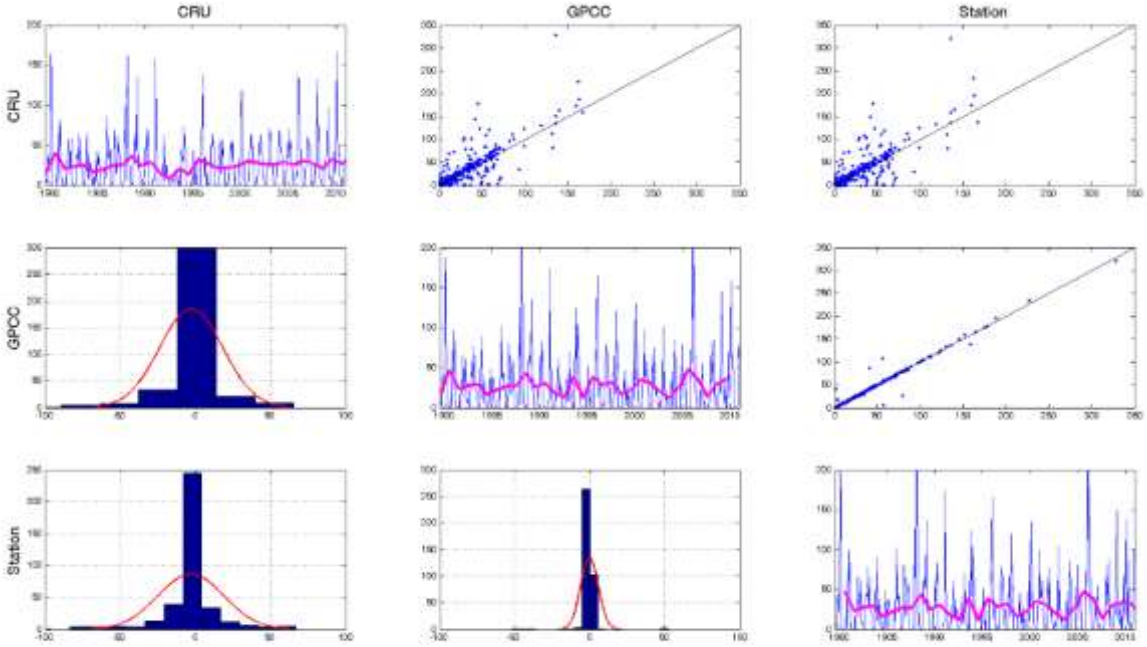

Figure 32 - Analysis of precipitation datasets (GPCC and CRU) and ground measurements for Tshane station (Botswana)

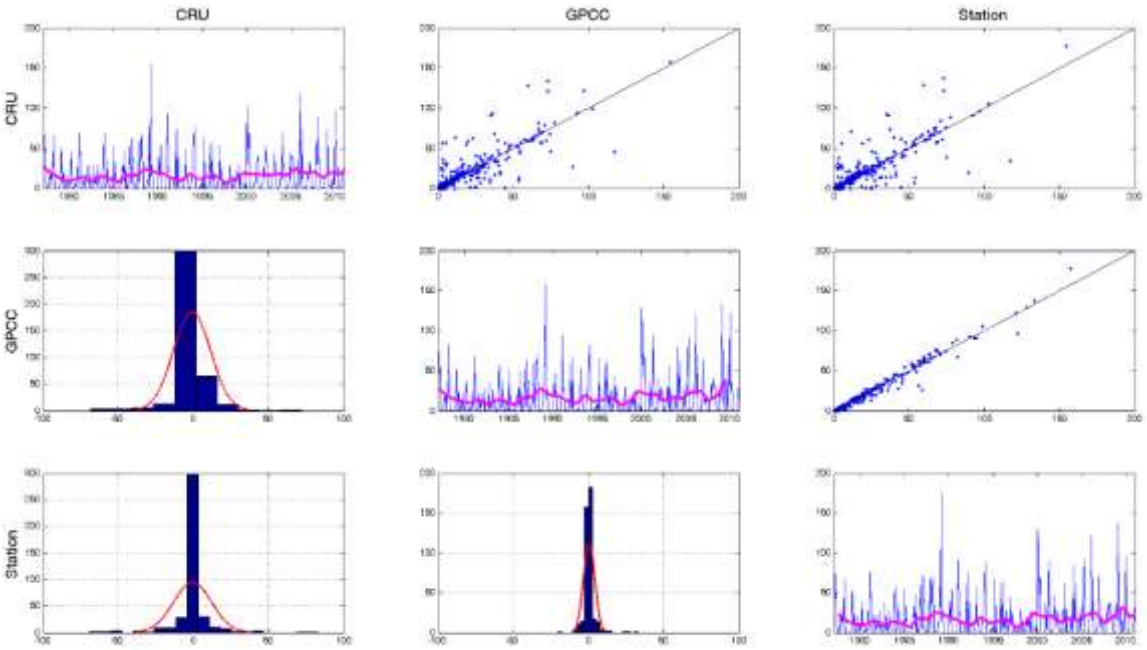

Figure 33 - Analysis of precipitation datasets (GPCC and CRU) and ground measurements for Gemsbok Park – Twee Rivieren station (South Africa)

Résumé

Les eaux souterraines constituent la principale source d'eau potable dans le monde entier et jouent un rôle essentiel dans le soutien alimentaire et production énergétique. Elles fournissent de l'eau potable à au moins 50 % de la population mondiale (UNESCO - WWAP , 2009) et représentent 43 % de toute l'eau utilisée pour l'irrigation (Siebert et al., 2010). Les eaux souterraines jouent un rôle important dans l'adaptation de la société à la variabilité et au changement climatique, en particulier parce qu'elles sont moins exposées aux effets du changement climatique que l'eau de surface (Van der Gun, 2012, Taylor et al., 2013). Sa capacité tampon unique offre un atout majeur pour réduire le risque de pénurie temporaire d'eau, et de créer les conditions pour la survie dans les zones plus exposées aux effets des changements climatiques (par exemple, semi-arides et les régions arides). Alors que l'extraction des eaux souterraines a augmenté de plus de 300 % au cours des 50 dernières années avec de grands avantages socio-économiques (Van der Gun, 2012), son développement et son utilisation est souvent tombée en dehors des cadres de gouvernance (FAO, 2015).

Un des enjeux majeurs pour améliorer la gouvernance des eaux souterraines vient du fait qu'une part importante de celles-ci est situé dans un nombre limité de très grands systèmes aquifères de plus de 100 000 km² (Margat et Van der Gun, 2013), en ajoutant donc une dimension transfrontière qui doit être prise en considération dans l'élaboration des politiques de gestion. Les preneurs de décisions exigent de plus en plus des outils de gestion innovants pour faire face aux vastes conséquences des changements climatiques. Gravity Recovery and Climate Experiment de la NASA (GRACE) - première mission en mesure de surveiller à distance les variations totales de stock d'eau (y compris les eaux souterraines) par satellite - a fourni de nouvelles informations de la dynamique des grands aquifères (Scanlon et al, 2012, Richey et al, 2015..) depuis 2002. Toutefois, étant donné que la dynamique des eaux souterraines n'est pas uniquement liée aux caractéristiques temporelles de pompage, mais est également affectée par la variabilité climatique interannuelle et multi-décennales (Shamsudduha et al. 2012, Kuss et Gurdak, 2014), plus de temps d'observation que celui que GRACE actuellement permet est nécessaire pour séparer les impacts respectifs aux activités anthropiques (utilisation des terres, abstraction) et climatiques sur les ressources en eau. Ainsi, il est nécessaire d'étendre les informations fournies par GRACE dans le «passé» pour mieux évaluer l'évolution actuelle et future des ressources en eaux souterraines.

Cette étude vise à ouvrir la voie à de meilleures politiques de gestion dans les grands aquifères en « reconstruisant » l'évolution des changements de stock d'eaux souterraines en fournissant une première évaluation quantitative des effets potentiels des activités anthropiques et des oscillations climatiques des cycles tels que El Niño/La Niña, Oscillation décennale du Pacifique, et l'oscillation multidécennale de l'Atlantique sur les grands aquifères (zone > 100 000 km²) situés dans les régions arides / semi-arides et tempérées . La validation est effectuée en comparant la variation du stock d'eaux souterraines obtenus par GRACE, un modèle développé dans le cadre de ce travail et des mesures au sol.