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APPLICATION OF THE RAINFALL INFILTRATION BREAKTHROUGH (RIB) MODEL FOR GROUNDWATER RECHARGE ESTIMATION IN THE BIRIM NORTH DISTRICT OF EASTERN REGION, GHANA

BY

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DECLARATION

I, Nyarko Delaiah Antwi, hereby declare that this thesis is a result of an original research undertaken under supervision of Prof. Sandow Mark Yidana and Prof. Larry-Pax Chegbeleh towards the award of Master of Philosophy in Hydrogeology in the Earth Science Department, University of Ghana. And that, to the best of my knowledge, it has not been presented elsewhere for another degree except where due acknowledgement has been made in the text.

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ABSTRACT

Evaluation of Groundwater recharge from rainfall is essential for sustainable water resources management, particularly in arid and semi-arid environments. The Birim North District, where majority of the population relies on groundwater due to pollution of surface water sources, has already experienced a drop in groundwater levels attending to the cumulative impacts of human activities and climate change. This project applies the Rainfall Infiltration Breakthrough (RIB) methodology to estimate groundwater recharge in the shallow unconfined, saprolite aquifer system in the Birimian Province in Southwestern Ghana. The Water level fluctuation (WTF) approach was used to estimate groundwater recharge in order to check, augment, and confirm the Rainfall Infiltration Breakthrough (RIB) recharge estimates by comparing such groundwater recharge estimates. The specific yield values acquired from the previous studies were compared to those acquired using the linear regression model as a quality assurance measure. The validity of the analysis, i.e., the association between rainfall and groundwater level, was established as a result of this. The line drawn in the regression model for determining the specific yield corresponded to 0.06, which was close to the value (0.05) obtained from literature. The RIB model estimated local recharge at 2.9 % to 21.4% of mean annual precipitation (MAP). The WTF approach estimated recharge to be between 3.2 % to 22.6 %. The prediction showed that decreased rainfall had no effect on groundwater levels during the simulation period in the climate scenario analysis. However, the ratio of recharge rate to precipitation did not alter considerably; it was somewhat greater than the baseline. Correlation examination of rainfall and observed water level fluctuation (WTF) data at the monthly scale, along with recharge estimates derived from other approaches, indicated that the RIB results based on monthly data were plausible and could thus be utilized as recharge estimates.

These findings suggested that using these methods to estimate groundwater recharge provides opportunities for assessing temporal variations in groundwater recharge and thus facilitates groundwater resources management. The method can estimate groundwater recharge in similar regions with adequately long time series of rainfall and groundwater levels. The RIB model is particularly suitable for shallow unconfined aquifers with minimal transmissivity; nonetheless, the RIB model's utility for application in various climatic locations and hydrogeological circumstances needs to be further investigated. These strategies could be tested in the future in catchments that have similar conditions of physiographic and hydrogeologic systems to the current research region.

DEDICATION

This work is dedicated to my parents, Rev. Daniel Nyarko and Mrs. Beatrice Nyarko, for their unrelenting support and prayers at all times. To my siblings, Daniel and David Nyarko, for everything they ever did in my journey to this point. Not forgetting my family, friends, and everyone else I could not mention here.

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CHAPTER ONE

INTRODUCTION

1.1 Background

Groundwater recharge is a crucial step in replenishing groundwater resources. Groundwater has become increasingly constrained over time as a result of rising population, water shortages, change in climate, inadequate water service delivery, and degradation of surface water quality, particularly in semi-arid rural areas (Nemaxwi et al., 2019). Groundwater is a more potentially reliable source of water than surface water since surface water is more prone to pollution than groundwater (Banoeng-Yakubo et al., 2005; Duriez, 2005; Ghasemizadeh, 2015; Nemaxwi et al., 2019). The groundwater recharge rate is one of the most important factors in semi-arid areas for sustaining long-term groundwater use, making rational groundwater allocation decisions, and establishing successful water and environmental management strategies.

Estimating recharge is critical in any groundwater system analysis and evaluating the effects of abstractions for various uses (Sophocleous 2005). Estimating groundwater recharge is a major challenge for assessing sustainable groundwater development and management, particularly in arid and semi-arid regions where rainfall and recharge are low while evapotranspiration is considerable (Sun et al., 2013). Groundwater modelers have faced major challenges due to the complications of aquifer geology and the unpredictability associated with the meteorological data of a given area (Ahmadi et al., 2012; Ghafari et al., 2018; Sharda et al., 2006).

Ghana has abundant groundwater resources (Johnston and McCartney, 2010). Many towns and villages in Ghana have adopted groundwater as their primary source of water since it is both a practical and economically viable source of water (Nsiah et al., 2018). Aside from the benefits to society's economic standing and human health, the advantages of using groundwater are numerous. To name a few, its high quality in terms of chemical and bacteriological content, its ability to survive under climatic constraints, its readiness for use referring to the low treatment cost as compared to surface water, and the ease of installing hand pumps in remote places (Lutz et al., 2014). According to estimates, potable surface water can cost up to twice as much as water derived from aquifers in areas with less than 5,000 residents (Krautstrunk, 2012). While many communities in Ghana and across Africa already have access to potable water, population growth and the Millennium Development Goals necessitate ever-increasing access (Lutz et al., 2014). As a result, hand-pump wells are in limited supply.

Despite this, research demonstrates that change in climate will have a global impact on groundwater resources (Earman and Dettinger 2011; Epting et al., 2020; Wu et al., 2020). According to Kankam-Yeboah et al., (2009), Ghana will become water-stressed by 2025, even if climate change and its repercussions are not considered. Change in climate will exacerbate the problem by creating a 5-22 percent decline in groundwater recharge in 2020 and a 30-40 percent decrease in 2050 (Kwoyiga and Stefan 2019).

The Akyem area, which comprises unique hydrogeology, abounds in natural resources, mainly minerals and forest products, is located in the Eastern region of Ghana (Owusu, 2012). Granite and upper and lower Birimian rock formations comprising phyllite, schist, greywacke, metavolcanic, and quartzes underlie a large portion of the area. These rocks have a lot of potential for extracting groundwater (Birim North District, 2006; Nartey et al., 2011). It is part of the Birim North District and has a history of widespread artisanal gold mining activities in and along river systems. As a result, water from most rivers has become unsuitable for human consumption, requiring most communities to rely on wells or boreholes for water supplies (Attiogbe and

Nkansah, 2015; Attua et al., 2005; Ghana Business News, 13 March 2010). Thus, groundwater is the primary source of water in these communities (Frimpong et al., 2016). According to Asante and Asare (2008), the demand for adequate water to satisfy the ever-increasing domestic, industrial, and agricultural needs is increasing in the Birim North District. In addition to domestic water abstractions, mine dewatering activities and the impacts of unregulated, illegal surface mining activities have been threats to the sustainability of groundwater resources, especially in the shallow unconfined saprolite aquifer system in the area. Therefore, there is the need to assess the current levels of groundwater recharge to provide data to assist in resource governance and sustainable management.

Several hydrological and hydrogeological studies have been conducted in the research region. For instance, Attua et al., (2005) evaluated the water quality of rivers used as drinking sources in artisanal gold mining settlements in the Akyem-Abuakwa area using a multivariate statistical technique. Acceptable, reasonably good, bad, and seriously polluted are the four degrees of water quality discovered using cluster analysis. Only the Subri and Kadee rivers have been recognized as "safe" to drink. Discriminant analysis indicated that turbidity, arsenic, temperature, phosphate-phosphorus, and total dissolved solids are the most important variables for identifying the drinking water quality of river sources. Five components with eigenvalues larger than unity were found in the principal component analysis, accounting for more than 80% of the variation in water quality. More than 26% of the variation was due to artisanal gold mining, with arsenic and mercury being the primary pollutants in the area. Agricultural operations and the disposal of residential garbage are two further causes of contamination.

Nartey et al., (2011) measured contamination levels of mercury in rivers and streams near small-scale gold mining locations in Ghana's Birim North District. The overall concentrations of mercury recorded upstream were substantially lesser than the measured downstream amounts, according to the findings. In addition, the contents of mercury in the water samples measured in both seasons surpassed the WHO drinking water guideline level. During the dry season, however, one downstream total mercury content was beyond the recommended levels. In both seasons, the overall contents of mercury in sediments upstream and downstream surpassed the US-EPA guideline threshold of 0.2 mg/kg. In both seasons, the overall contents of mercury in wells in the research area exceeded the WHO recommended limit. In the wet season, the number of mercury contents in boreholes had lesser concentrations lower than in the dry season.

Frimpong et al., (2016) studied the assessment of groundwater quality in the Akyem Mine Area of Newmont Golden Ridge Ltd (NGRL). The study assessed the groundwater quality in the area and determined if there were significant changes in parameters, possibly due to activities of the mine and other anthropogenic activities. Results and analysis indicated that water from both shallow and deep wells is either neutral or weakly acidic and is as well within the WHO acceptable range for color and TDS. The study found that though there was external influence responsible for significant variations in TDS and pH values for shallow and deep boreholes, the influence has not been significant enough to cause parameters to exceed WHO guidelines. The following mining-related metal contaminants were below the detection limit and thus posed no health risk to the consumption of water from both shallow and deep wells. Hence, based on their research, there have not been significant impacts of the parameters under consideration on groundwater quality in the Akyem Mine area.

Although various studies have been conducted in the study area, there has been no previous research on groundwater recharge estimation, according to the literature in the area. There are no comprehensive recharge investigations for this specific location, either published or unpublished. There is, therefore, a need to provide a reliable estimation of the groundwater recharge to manage the groundwater resources. Furthermore, the geographical distribution of the rate of recharge is important for groundwater management and analysis. Infiltration of precipitation is the primary source of recharge to groundwater in the model area (AMEC Geomatrix, 2010). Hence understanding the groundwater recharge response to rainfall events is critical to understanding the catchment's aquifer system and its dynamics.

In African countries, several methods for groundwater recharge estimation have been employed with varying degrees of accuracy in recent decades (Beekman and Xu, 2003; Sun et al., 2013). The results of using these methods revealed that when alternative techniques and record data sets were employed, groundwater recharge estimates offered by several professionals differed substantially. Most groundwater recharge approaches have been designed for large-scale applications, while little information is available describing processes at the local scale (Sun et al., 2013). Various studies have been conducted to estimate rainfall-induced groundwater recharge (Baalousha 2005; Szilagyi et al., 2011; Tshelane et al., 2014). In spite of the numerous studies found in the literature, uncertainties in parameter estimation continue to be a concern due to the variety of factors that influence the rate of recharge (Liggett and Allen, 2009). Qablawi (2016) presented an analysis of four groundwater recharge estimation methods, as well as an assessment of their accuracy and applicability. The soil water balance, Chaturvedi formula, seasonal recession method (Meyboom method), and well-level data were the approaches used. Because the climate in the study area was sub-humid continental, the results of the Chaturvedi formula were less dependable than those of the soil water balance and the well-level data method. The Meyboom method's conclusions were also unreliable for two reasons: the approach predicts average groundwater recharge over a five-year period, and it cannot calculate a negative figure. There were also uncertainties with the soil water balance and the well-level data methods. Due to a paucity of evaporation data in some places, these methods may be difficult to apply. The computation of recharge may be influenced by the method employed to compute evaporation.

Groundwater is very vital in arid and semi-arid environments than in humid areas because they receive a small amount of rainfall. After the occurrence of rainfall in a dry season environment, the element of recharge is the utmost significant aspect. However, it cannot be measured directly, resulting in a scarcity of data (Majdabady et al., 2020). As a result, most approaches for estimating groundwater recharge do not appear to be totally appropriate due to the challenges associated with quantifying groundwater recharge. As a result, the rainfall infiltration breakthrough (RIB) technique was composed because it is a simple, cost-effective, and easy-to-use tool for estimating groundwater recharge (Ahmadi et al., 2014; Majdabady et al., 2020; Sun et al., 2013).

This study sought to estimate the rate of groundwater recharge that develops from rainfall by using rainfall infiltration breakthrough (RIB) to support water resources decision-making, which would be essential for sustainable water development and groundwater modeling in the study area.

1.2 Problem Statement

The Birim North district is one of the country's forested areas. Parts of these forests have been set aside to secure the long-term use of natural resources, culminating in the establishment of nine forest reserves to preserve a portion of the district's original vegetation (Attiogbe and Nkansah, 2015). However, a variety of variables have played a significant role in the district's forest cover decline. Poor farming methods, deforestation (particularly the actions of unlicensed chainsaw operators), bushfires, and reckless mining activities are among the most prominent of these. According to Attiogbe and Nkansah (2015), these human activities have caused the Pra River's headwaters to dry down during the dry season, especially along its banks. This usually results in long-term water shortages in areas that rely on it for their water supply. If mining activities damage these water resources, the problem may escalate (Birim North District, 2006).

Over the years, the growing use of groundwater and the resultant increased demand on limited groundwater supplies as a result of the increased usage of these aquifers as a source of water for domestic, commercial, and irrigation purposes have led to the introduction of mechanized pumps. Most of the surface watercourses in the district, including the Pra River and Birim River, have been contaminated by small-scale mining and anthropogenic activities (Frimpong et al., 2016). Due to the pollution of the surface water, Newmont Golden Ridge Ltd (NGRL) and the Government have provided the surrounding towns/villages with boreholes fitted with hand pumps. Thus, borehole and hand-dug wells represent the primary source of freshwater for residents in these communities (Frimpong et al., 2016). Despite the paucity and insufficiency of drinkable water in several places due to contamination of surface water bodies, the underground water reserve is rich (Nartey et al., 2011). Little is known or documented concerning the physical hydrological parameters of the Birim Basin. Thus, quantifiable parameters that are required to conduct a mass water balance analysis that could provide further sustainability investigations in this area are restricted. The long-term viability of water resources in this region is key to sustaining the lives of the communities in this area. The goal of this research was to create a modified groundwater recharge estimates for the Birim Basin using the rainfall infiltration breakthrough model. Monthly recharge estimates utilizing the modified RIB program and current data in the research area are provided, accompanied by sensitivity analysis. The effects on groundwater levels and the expected monthly recharge rate are investigated using scenarios analysis of data from the research region and varying precipitation inputs.

1.3 Objectives

The main objective of the study was to develop an improved groundwater recharge estimate from rainfall and groundwater level data in the Birimian Province in Southwestern Ghana.

The specific objectives include to:

- Adopt and modify a methodology for estimating groundwater recharge using rainfall and groundwater level data;
- Estimate groundwater recharge using RIB and WTF methods and compare findings;
- Determine the relationship between rainfall and groundwater recharge in order to determine the response of groundwater recharge to rainfall events;
- Evaluate various scenarios of climatic variability on the sustainability of the shallow unconfined saprolite aquifers in the study area.

1.4 Study Area

1.4.1 Location

The Birim North District was established in 1987 by the former Birim District Council as part of the government's decentralization effort to promote effective localized administration and speed up development in the area. It is located in the Eastern Region and is bounded to the north by Kwahu West, to the west by the Ashanti Region's Asante Akyem South, Amansie East, and Adansi South Districts to the south by Birim South District, and to the east by Atiwa and Kwaebibirem Districts. The region, particularly its headquarters, New Abirem, is strategically located near major commercial towns such as Nkawkaw, Oda, and Kade (Figure 1.1).

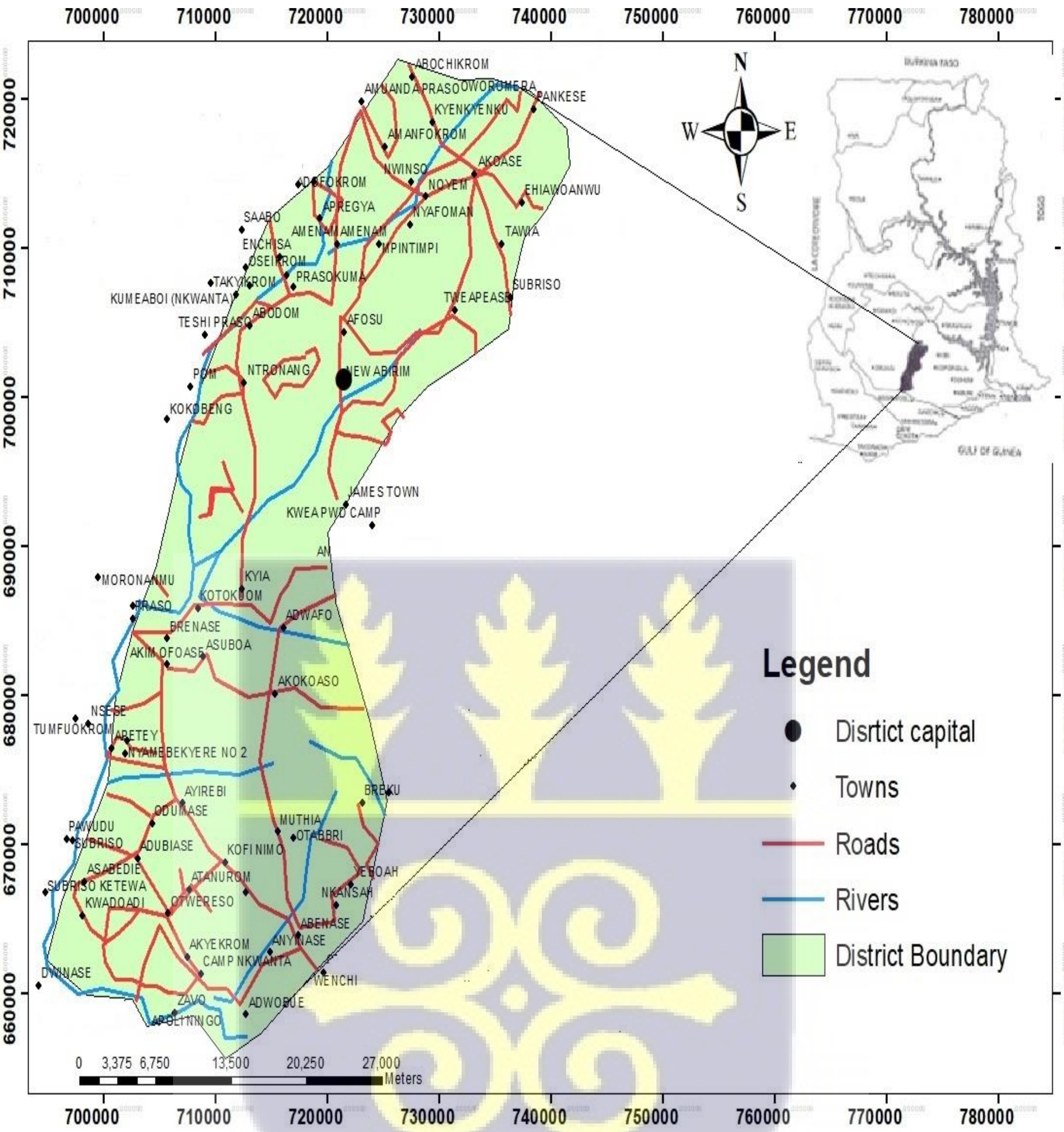

Figure 1.1: Map of Birim North District Assembly

The district's economy stands a better chance of improving with improved road conditions connecting it to key commercial areas. New Abirem is a nodal or confluence town since it is positioned at the intersection of the Nkawkaw-Oda-Kade highways (Birim North District, 2006). The area is located between latitudes 6.15° N and 6.35° N; and longitudes 0.20°W and 1.05°W (Owusu, 2012). The district has an estimated total land area of 1,250 km^2, accounting for roughly 6.47 percent of the Eastern Region's total land area.

1.4.2 Climate

The research area is located in a wet semi-equatorial climatic zone with a double maxima rainfall pattern. The climate is tropical, with a daily temperature range from approximately 14.0 ˚C to 36.4 ˚C, according to data collected by Clear Creek Consultants in 2009 (Clear Creek, 2010). The first rainy season begins in late March and lasts until the beginning of July. The second season begins in mid-August and runs until the end of October. The amount of rainfall received in the district is between 1500 mm and 2000 mm per annum. Monthly average temperatures in the model area range from approximately 24.3˚C in August to 26.6˚C in February and March (Clear Creek, 2010). Monthly average relative humidity at the Akyem Project site range from approximately 66.9 percent in January to 88.8 percent in July (Clear Creek, 2010). The district's agricultural operations, particularly the growth of food crops and tree plants like oil palm and cocoa, have benefited from a large amount of rainfall and temperate temperatures.

1.4.3 Relief and Drainage

The district is primarily undulating and hilly in nature. Nature is undulating and mountainous, rising to more than 61 meters above sea level in certain places. The highest peak in the district is Kwasiakwasi Mountain (N 06°28'09.6" & W 00°54'44.3"), which is found in the Kwasiakwasi Forest Reserve and climbs to a height around 800 m above sea level. Streams such as Nyanoma, Nkwasua, and Aprokuma originate in the Kwasiakwasi Forest Reserve.

The territory is primarily drained by two rivers, the Pra and its tributary, the Birim. These rivers' tributaries in the district include the Nwi, Suten, Mamang, Adechensu, Sukrang, Nkwas ua, Nyanoma, and Afosu. These rivers flow from the northeast to the southwest, eventually joining the Pra, which flows south and joins the sea near Shama in the Western Region (Figure 1.2). The Pra River forms the western boundary between the district and the three districts of the Ashanti Region. The Birim River also acts as the district's southern boundary (Attua et al., 2005; Birim North District, 2006).

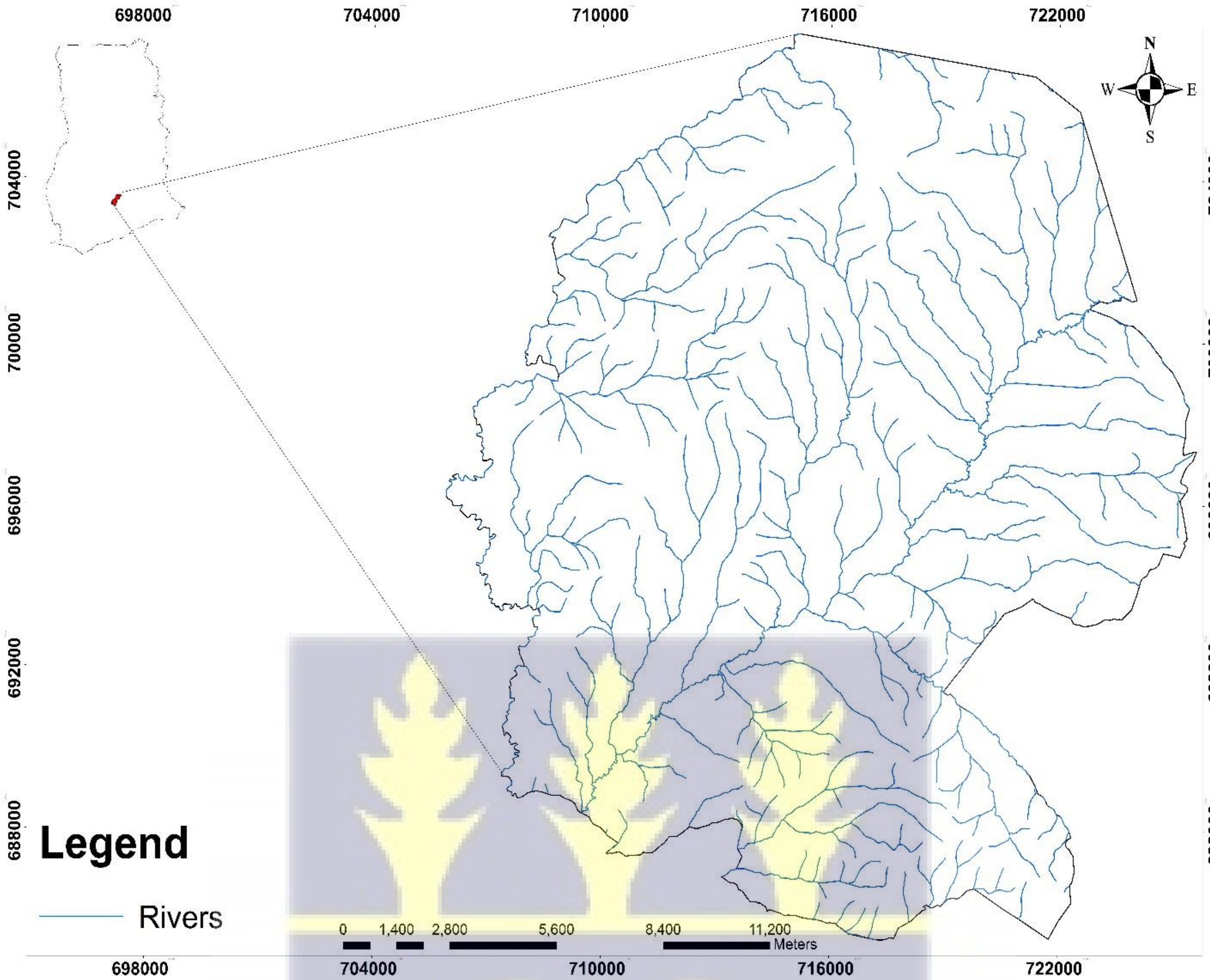

Figure 1.2: Drainage system of the study area

1.4.4 Vegetation and Soil

The district is located in Ghana's semi-deciduous forest area, which is made up of tall trees with evergreen undergrowth. It features a considerable number of economically valuable trees for the lumber sector. There are nine forest reserves in the area. It has much undergrowth, but due to the rapid expansion of the cocoa and oil palm businesses, illegal chain saw operators' activities, and frequent bushfires, the original forest is quickly becoming a secondary type (Birim North District, 2006).

The district's soils can be classified into five major groups. Swedru-Nsaba/Ofin Compound Association, Atiwa-Atukrom-Asikuma-Ansum Compound Association, Juaso-Manso-Debia Association, Bekwai-Oda Association, and Birim-Chichiwere Association are among them (Antwi 2010; Birim North District, 2006).

The major soil formation is the Swedru-Nsaba Ofin Compound Association. They are formed over granite. They can be found in the areas of Prankese, Nkwateng, Otwereso, and Abenase. This compound association is made up of two simple associations: Swedru-Nsaba and Nta-Ofin. Nta Offin evolved from the transported products of the former Swedru series. The Swedru Nsaba series, which is heavy in magnesia and potash, is excellent soil for tree and arable crops, but especially for cocoa. Because ofin soils are unsuited for tree crops, they are primarily used for cultivating dry season vegetables, sweet potatoes, sugarcane, and rice (Birim North District, 2006).

The soil in the Atiwa–Atukrom–Asikuma–Ansum compound series is restricted to a small portion of the district near Amuana Praso. Topsoils are dark brown, somewhat humus, silty, clay, and loam, with subsoils that are reddish-brown to red soft clay loam. Because of their high acidity and low base status, these two soil series are infertile. They are suggested for coffee, oil palm, other tree crops, and forestry (Birim North District, 2006).

The Juaso-Manso-Adubea Association is located near Noyem, Prasokuma, and Atobiaso. It is dark brown and shallow, and it helps with oil palm output. Bekwai- Oda Association is located in the vicinity of New Abirem, Ntronang. They are well-drained and suited for growing a wide range of tree and arable crops, including cocoa, coffee, citrus, oil palm, avocado pear, mangoes, yams, maize, cassava, and plantain. The Oda series also inhabits flat, rather large regions next to rivers and streams. They are ideal for mechanized irrigated rice farming (Birim North District, 2006).

The Birim-Chichiwere Association was discovered at Edubia. It was built on the River Birim deposits. It is moderately well-drained, deep, and easy to use with machines; it occurs on almost flat areas with little or no erosion, and it is appropriate for a wide range of tree and arable crops. Chichiwere is thought to be harmful to tree crops (Birim North District, 2006).

1.4.5 Geology and hydrogeology

Akyem is underlain by the Birimian metasedimentary and metavolcanic rock units (Figure 1.3). These include northeast-trending belts of folded, metamorphosed volcanic and sedimentary rocks of the early Proterozoic, which underlie the model area. The southeast side of the model area is underlain by rocks of the Birimian Supergroup. In this region, the Birimian terrain comprises northeast-trending belts of volcanic and volcanoclastic material separated by broad turbidite-dominated sedimentary basins. Tarkwaian sediments unconformably overlay the Birimian volcanic belts in the northwestern portion of the model area. The Birimian rocks consist of black phyllites, metasiltstones, metagreywackes, tuffaceous sediments, tuffs, and hornstones (Kesse, 1985). Tarkwaian sediments consist of conglomerates, sandstones, and phyllites. Structural features of the Birimian and Tarkwaian units display a strong northeast-southwest trend (GRRL 2005 and 2006; Ireland et al., 2001). The sandstone (quartzite) consists of variable amounts of feldspar, sericite, chlorite, ferriferous carbonate, magnetite, or hematite and epidote. Locally, the Akyem deposit is a shear zone hosted gold deposit almost at the contact between Birimian metasedimentary and metavolcanic rock. Mineralization is chiefly in the metasedimentary separated by graphitic shear. The deposit strike 070° and dips 60° to 65° SSE (Atule, 2007).

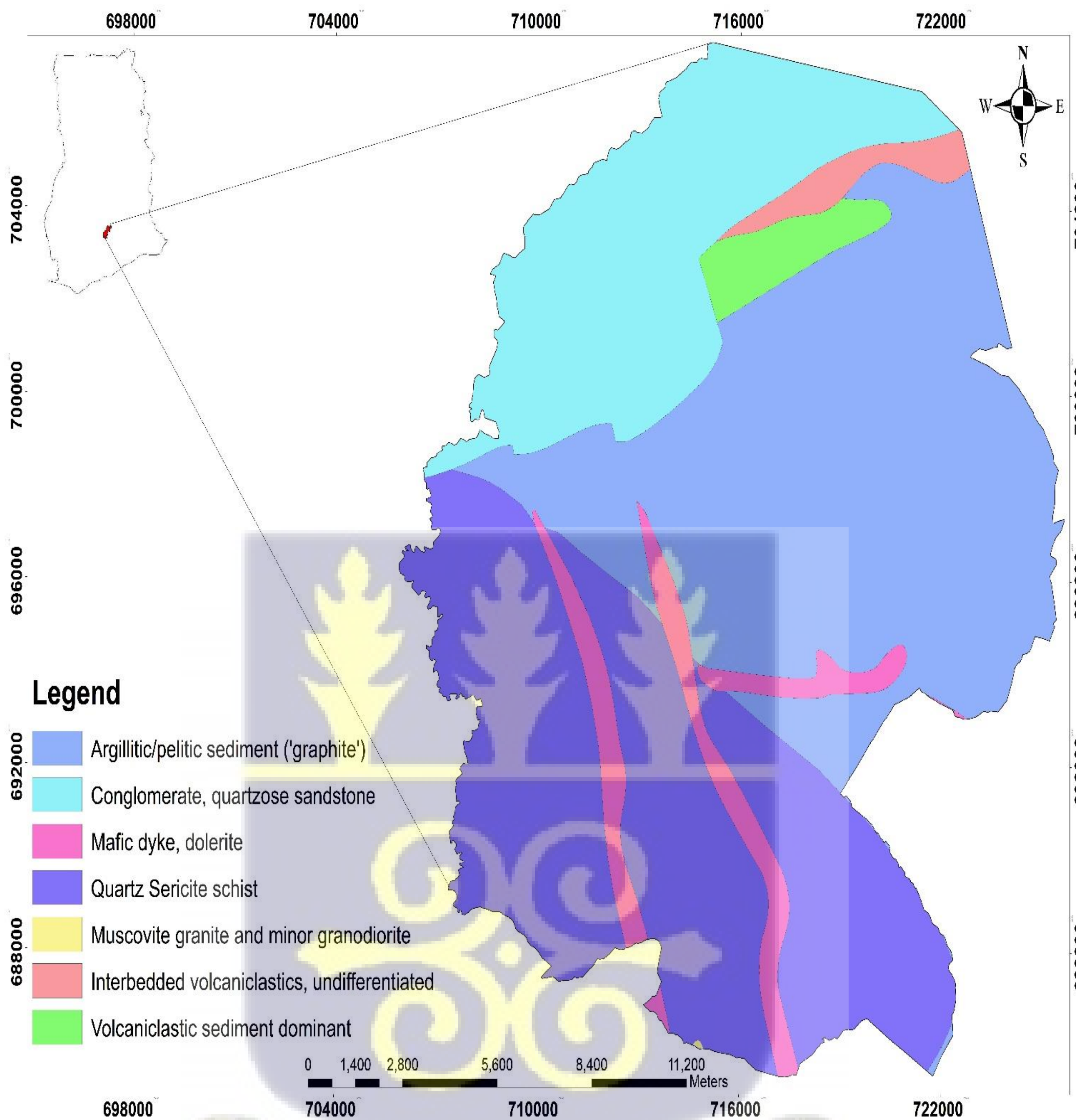

Figure 1.3: Geological map of the study area

In-situ weathering of bedrock defines the hydrostratigraphy in the model area. Weathering of bedrock to tens of meters depth is caused by the tropical climate. The pace at which rocks weather is affected by rainfall and temperature. Chemical weathering is accelerated by higher temperatures and more rainfall. Highly weathered bedrock (saprolite), moderately weathered bedrock (saprock), and fresh rock (bedrock) are the three basic hydrostratigraphic units. Localized zones of increased permeability have emerged from structural characteristics in bedrock and quartz veining. In the flood plains of streams and rivers, alluvial deposits lie on top of saprolite. The following are brief descriptions of four hydrostratigraphic units, as well as the hydraulic property values determined during studies by AMEC Geomatrix (2010), Golder (2004 and 2006), and Lycopodium Knight Piesold (2005).

(a) Alluvium

In the model area, alluvium is made up of fluvially deposited silt, sand, and gravel along streams and rivers. Alluvial deposits' lateral dimensions and thicknesses are poorly defined. Limited investigations conducted in support of the tailing storage facility along the Adenchensu river indicate that the alluvial deposits are small, measuring less than 200 meters wide and 2 meters thick (Lycopodium Knight Piesold 2005). Hydraulic conductivity of $1.6\text{x}10^{-3}$ centimeters per second (cm/s) was found in an aquifer test conducted in alluvium (AMEC Geomatrix 2010).

(b) Saprolite

The Saprolite is a near-surface unit that, because of its clay composition, may resist infiltration, but it lacks weathered quartz veins that might act as conduits for infiltration and groundwater flow (Anon., 2010). Saprolite is made up of decomposing bedrock that has been weathered in place into sand-silt-clay-sized particles with remaining quartz veins. The saprolite is usually covered by a thin layer of lateritic soil that is 1 to 5 meters deep. The thickness of the saprolite-laterite unit is usually less than 60 meters (Golder 2004). Topography affects the saturated thickness of the saprolite unit. The saturation thickness of the saprolite near the proposed Akyem mine pit ranges from 1 to 20 meters (Golder 2004). With topography, the saturated thickness of the saprolite unit varies. The saturated thickness of the saprolite near the proposed Akyem mine pit ranges from 1 to 20 meters (Golder 2004). Saprolite has a hydraulic conductivity range of $3x10^{-7}$ to $3x10^{-4}$cm/s with a geometric mean of 8 x10-6 cm/s (Golder 2004 and Lycopodium Knight Piesold 2005). The saprolite's upper, less permeable part can act as a semi-confining layer for this productive zone. In contrast, the lower, usually saturated part of the saprolite is characterized by lower secondary clay content, thus creating a zone of enhanced hydraulic conductivity.

(c) Saprock

The Saprock, on the other hand, is relatively impermeable, allowing groundwater to be stored and transmitted through fractures within the rock (Frimpong et al., 2016). Saprock is a type of bedrock that has been moderate to lightly weathered. The transition between saprolite and fresh bedrock is represented by this 1- to 20-meter-thick unit. The saturated thickness of the saprock around the proposed Akyem mine pit varies from zero (unsaturated) in upland places to 5 to 10 meters in lowland locations. Based on falling head tests and pumping tests, hydraulic conductivity estimations range from $2x10^{-5}$ to $2x10^{-2}$cm/s with a geometric mean of 4×10^{-4}cm/s (AMEC Geomatrix 2010; GRRL 2005; Golder 2006,).

(d) Bedrock

Groundwater in the bedrock occurs under confined conditions and is the dominant water-bearing associated with the fracture system. Greywacke and mafic volcanics are the most common bedrock types found in the area. Variations in fracture density, as well as the magnitude and degree to which fractures are interconnected, affect the hydraulic characteristics of bedrock aquifers (Anon. 2010). According to SGS's vertical electronic sounding surveys and dipole-dipole resistive surveys, most fractures are localized between 30 and 45 meters below ground surface (bgs) (GRRL 2005). The hydraulic conductivity estimates from tests performed on wells completed in bedrock range from 4.6×10^{-6} to 2×10^{-3} cm/s with a geometric mean of 1.2×10^{-4} cm/s due to the variety of fracture conditions (Golder 2006 and 2004, and Lycopodium Knight Piesold 2006, AMEC Geomatrix 2010).

Although the underlying matrix's general permeability is low, faults, shear zones, and fractures linked with structural trends are the most common groundwater transmission structures. A sequence of dykes that have been mapped perpendicular to the dominant structural trend may produce local barriers to groundwater flow in the area, whereas these primary permeability features may transmit relatively considerable amounts of water. Golder (2004) used packer testing in four exploration boreholes, and AMEC Geomatrix(2010) used aquifer testing in four deep wells to examine the hydraulic parameters of the Akyem fault and its associated structures. Hydraulic conductivity values for the structural zone ranged from 3×10^{-6} to 3×10^{-4} cm/s with a geometric mean of 1.5×10^{-5}cm/s. Generally, borehole depths drilled through rocks of the Birimian and Tarkwaian Systems range from 35 to 62 m, with an average of 42 m (Agyekum, 2004). Borehole depths in granitoid-underlain areas are similar, ranging from 35 to 55 meters, with an average of 50 meters (Carrier et al., 2008). In some areas, the regolith is tapped at relatively shallow depths using hand-dug wells. The Birimian and Tarkwaian Systems' productive zones have aquifer transmissivity ranging from 0.2 m^2 /d to 119 m^2 /d, with an average of 7.4 m^2 /d. Transmissivity within the regolith is slightly higher than that observed in the integrated aquifer system, ranging from 4 m^2 /d to 40 m^2 /d with an average of about 10 m^2 /d (Carrier et al., 2008). Borehole yields for the integrated aquifer systems in the Birimian and Tarkwaian Systems are generally low, ranging from 0.48 m^3 /h to 36.4 m^3 /h with a mean yield of 7.6 m^3 /h. The lower yields observed in these rocks are most likely due to differences in the degree of weathering within the granitoid (Yidana et al., 2008).

CHAPTER TWO

LITERATURE REVIEW

2.1 Introduction

This chapter gives a review of past research on groundwater recharge estimation, with a focus on the use of water table fluctuation and rainfall infiltration breakthrough methodologies. It also thoroughly discusses the recharge estimation models in groundwater studies and how concepts and methods have been used in other jurisdictions to achieve similar results. The chapter scrutinizes the methods that the literature suggests for groundwater recharge estimation and how they apply in the current study aiming to highlight the most beneficial methodology for the study area.

2.2 Groundwater Recharge

One of the essential variables in hydrological studies is groundwater recharge, but it is often one of the most difficult to estimate accurately. Groundwater can be recharged both by precipitation and/or surface water sources such as rivers and lakes infiltrating into the soil and rock layers of the ground (Bhattacharya et al., 2003). When assessing groundwater recharge, there is a need to distinguish between potential and actual recharge. Soil water percolating beneath the root system is a possible recharge; thus, soil water entering the aquifer is the actual recharge (Petersen, 2012). Most of the potential water for recharge would be retained at negative pressure (suction) in the vadose zone and is ineligible to be exploited. Instead, the actual recharge is the volume of water that actually makes it to the groundwater table and can be used (Sophocleous, 2004). Groundwater recharge may be categorized as (1) direct or indirect by reference to the origin of the recharging water, (2) piston or preferential flow by reference to the flow phase through the unsaturated region,

(3) point, line, or areal recharge by reference to the area on which it operates, and (4) current, short or long-term recharge by reference to the time scale over which it occurs (Boerner and Weaver, 2012; De Vries and Simmers 2002; Obuobie et al., 2012).

Groundwater recharge estimation is one of the vital issues in measuring the aquifer's sustainable yield in dry and semi-arid environments, as recharge rates are typically low relative to mean yearly rainfall and are therefore hard to estimate effectively (Xu and Beekman, 2003). There are several factors that govern groundwater recharge. These include the rainfall rate and length, the soil profile's previous moisture state, geology, soil properties, water table, and aquifer depth characteristics, vegetation and land cultivation, topography, and landscape (Ng et al., 2009). Albhaisi et al., (2013) noted that recharge has systematically risen over time as a result of degradation in South Africa's Berg River watershed. This was after being speculated that due to this shift in land use, evapotranspiration would decrease, and recharge would increase. In the soil, recharge is calculated by its composition, thickness, linkage, uniformity, and hydraulic elements, while the porosity of the aquifer determines the recharge magnitude.

2.3 Groundwater and climate change

Groundwater systems are influenced by climate change in a variety of ways. Climate change has the potential to influence the amount of soil infiltration, deeper percolation, and groundwater recharge in the hydrological cycle (Wu et al., 2020). Furthermore, increasing temperatures raise evaporative demand over land, reducing the amount of water available for groundwater recharge (Bates et al., 2008; Brielmann et al., 2009; Jesußek et al., 2013; Kipfer and Livingstone, 2008; Kurylyk et al., 2014; Menberg et al., 2014; Possemiers et al., 2014). Conversely, anthropogenic impacts on groundwater systems are primarily due to the pumping of groundwater and the indirect impacts of land-use changes and irrigation (Wu et al., 2020). Groundwater demand is fast rising in tandem with increasing population, while the change in climate puts extra strain on water resources and increases the likelihood of extreme drought (Wu et al., 2020).

Variations in the Earth's climate have the ability to have an impact on groundwater quality and quantity. There is a general consensus that the Earth's climate is changing and will continue to change in the future due to the impacts of global warming in the air (Earman and Dettinger, 2011). Future precipitation projections are not as consistent or reliable, but they often result in extremely arid areas and extremely moist areas (Intergovernmental Panel on Climate Change, 2007).

According to Earman and Dettinger (2011), overall warming combined with unknown precipitation changes is predicted to result in less freshwater availability in most situations than if the precipitation changes were the only factor. That is, even if precipitation stays unchanged (or slightly rises), warmer conditions will likely result in minimal recharge and runoff. Warming is expected to amplify runoff and recharge reductions where precipitation declines.

Many studies of the effects of change in climate on surface-water resources make no reference to groundwater and do not seem to account for groundwater contributions to streamflows in a meaningful way. This method is based partly on the observation that changes in climate have a greater impact on surface water than on groundwater, leading to the presumption that climate-change impacts on groundwater are not really severe. The impacts of variations in climate on groundwater systems are intrinsically unpredictable due to the uncertainty of future climate projections. Climate-groundwater interactions are typically incompletely defined and modeled, which contributes to ambiguity (Bates et al., 2008; Earman and Dettinger, 2011; Epting et al., 2020; Huggenberger and Epting, 2011). Higher total precipitation increases the quantity of water available for groundwater recharge at any particular place and may produce higher recharge devoid of other effects. On the other hand, less precipitation is likely to result in less recharge (Earman and Dettinger, 2011; Huggenberger and Epting, 2011).

The majority of research concentrates on entire catchment areas and solely looks at infiltrating precipitation water changes (i.e., "precipitation-fed aquifers"), with the exception of a few extensive assessments of extremely small aquifers (e.g., Malcolm and Soulsby (2000)). Similarly, in a review of Climate Change (CC's) influences on groundwater recharge, for determining the classification of groundwater resources, Smerdon (2017) emphasizes the need to know groundwater recharge mechanisms, which includes location and timing. The researcher reviews six scientific reports (Crosbie et al., 2013; Green et al., 2011; Kurylyk and MacQuarrie, 2013; Meixner et al., 2016; Moeck et al., 2016; Taylor et al., 2013) that show how the lack of certainty of future precipitation distribution and pattern from General Circulation Models (GCMs) results in different recharge estimates, to the point that modeling studies are frequently unable to estimate the degree and orientation (rise or fall) of prospective recharge situations.

2.3.1 Impacts of climate change on groundwater in Ghana

According to the Ghana National Climate Change Policy, climate change has caused a rise in temperature and a reduction in average yearly rainfall in the nation's natural borders (Ministry of Environment, Science and Technology, 2013).

In Ghana, attempts to comprehend climate variability and its influence on groundwater systems are inadequate, and there are ambiguities, as there are with rainfall, because most existing research is projection-based. Climate change and its effects on groundwater, on the other hand, have been a major cause of worry. The Volta Basin Authority conducted a transboundary diagnostic investigation, which found that climate change and its consequences are one of the Volta Basin's primary issues (covering parts of Ghana) (Mul et al., 2015). This supports Oyebande and Odunuga's (2010) conclusions that in the dry and semi-dry sectors of small areas, the Niger and Volta Basins will see a fall in streamflow and groundwater recharge as aquifer recharge has already dropped considerably. There is also a chance that the groundwater level will drop as a result of the reduced recharge.

According to the Council for Scientific and Industrial Research-Water Research Institute (CSIR-WRI) research on climate variability and water systems in Ghana published in 2000, climate change may lead to a broad reduction in groundwater recharge of 5-22% in 2020 and 30-40% in 2050 (Kankam-Yeboah et al., 2009).

In the Volta Basin, McCartney et al., (2012) used the Soil and Water Assessment Tool (SWAT) and Water Evaluation and Planning (WEAP) techniques to investigate the effects of climate variability on water systems. According to their research, the yearly mean rainfall, runoff, and average groundwater recharge would all decline by 2050. Under the A1B scenario, for example, groundwater recharge and mean yearly discharge will have been reduced by 53 percent and 45 percent by the end of the century, respectively.

There is, however, one aspect where these findings diverge. In a study, Obuobie (2010) estimated groundwater recharge in the White Volta River Basin, West Africa, in the perspective of future climate change and discovered that yearly recharge to the basin's groundwater, which is currently around 7% of annual rainfall, could increase to around 33% in the future for the period 2030-2039.

2.4 Groundwater Recharge Estimation Methods

Estimating groundwater recharge is critical for the effective and lengthy management of groundwater systems. Due to the difficulty of measuring groundwater recharge directly, a number of groundwater recharge estimation systems have been developed, ranging from physical methods to modeling approaches (Carrier et al., 2008; Scanlon et al., 2002). Since information is always inadequate and situations change over time and distance, most researchers believe that recharge estimation is effectively done in stages (Atta-Darkwa et al., 2013; Mare'chal et al., 2006). Methodologies based on surface water and unsaturated-zone data offer calculations of possible recharge, while those relying on groundwater data offer estimates of actual recharge, according to Scanlon et al., (2002). As a result, it is common practice to estimate recharge using numerous approaches in order to improve the accuracy of recharge estimates (Healy and Cook, 2002; Scanlon et al., 2002).

Some of the recharge rate estimation methods and techniques used are chloride mass balance, soil moisture balance technique, water table fluctuation, environmental tracer technique, and hydrological modeling.

2.4.1 Chloride Mass Balance (CMB) Techniques

The Chloride Mass Balance (CMB) is a very useful tool for calculating recharge in dry and semi-arid regions due to the difficulty encountered in using conventional water balance methods (Allison et al., 1994). It is one of the broadly applied approaches in water resources development and management (Brunner et al., 2004; Martin 2006; Obuobie 2008; Sandwidi 2007). The premise behind this technique is the fact that the chloride ion has a conservative nature and does not easily react with other elements in water when traveling through the unsaturated zone to the saturated zone (Marei et al., 2010; Zagana et al., 2007). It offers a direct estimate of recharge and is the least expensive with low-time integrating properties. The CMB technique is easy to use and does not necessitate the use of any complicated equipment. However, if used alone, the approach may not provide a reliable representation of recharge rates. When evaluating the results, the method's uncertainties and assumptions must be taken into account. The Chloride Mass Balance method for estimating groundwater recharge depends on several assumptions; The following are the assumptions that must be met in order for the approach to be applied successfully:

- Apart from precipitation, there is no alternative origin of chloride in groundwater.
- In the system, chloride has a conservative nature, which means that it does not leach from or cannot be taken by aquifer materials and is not involved in any chemical processes.
- With regard to long-term precipitation and chloride content in the precipitation, steady-state situations are sustained.
- Surface run-on and runoff are negligible (Zhu et al., 2003; Marei et al., 2010).
- The depth of the groundwater table ought to be sufficient to avoid the vaporization of groundwater (Zhu et al., 2003; Marei et al., 2010).
- Within the basin, there is no chloride recycling.

By using this method, recharge in mm/year (R) can be estimated by equation (1) as described by Marie et al., (2010);

$$R = \frac{Cl(p)}{Cl(gw)} \times P \qquad \text{2.1}$$

Where R, Cl(p), Cl(gw), and P are respectively recharge rate, chloride content in precipitation, chloride content in groundwater, and average annual precipitation.

Consequently, in places where significant amounts of the groundwater chloride originate from mineral dissolution processes or are affected by reactions that reduce its amounts in groundwater, the method can be inappropriate and lead to underestimation of recharge. The approach cannot be used in regions where evaporates underlie the areas or where saline water is evaporating or mixing. Its application in fractured rock systems is complicated because extra chloride is created by the breakdown in a groundmass of the rock; time is required to establish a new balance between the rock matrix and cracks; and if additional chloride is made, CMB recharge rates are thought to be the bare minimum. The results could be influenced by wind recycling of dried salt, unrecorded discharge, and assimilation by harvested plants.

Several researchers around the world have adopted the CMB approach to predict groundwater recharge with varying degrees of success: Obuobie (2008), for example, employed chloride mass balance to estimate recharge in Ghana's Upper East Region; the latter study's estimates were significantly higher than the former study's. These results revealed the method's uncertainty and highlighted that there is no single extensive estimation method that can be applied at regional scales.

In the same area, Obuobie et al., (2010) again applied the chloride mass balance (CMB) method in the Upper East region of Ghana to calculate the recharge to groundwater aquifers. Furthermore, the recharge is linked with water demand in order to determine the groundwater's potential for enhancing water supply. Their findings indicated that groundwater recharge accounted for 3 to 19 percent of average annual rainfall, indicating a large groundwater potential, as usage in the research area was estimated to be a little more than 14 percent of recharge. However, their research revealed that long-term mean chloride contents in rainfall could not be calculated due to a lack of data for the study area. Mean monthly chloride readings recorded in 2006 were utilized as the basis for the study.

Earlier on, the Chloride Mass Balance method, together with the Soil Balance technique, was used by Carrier et al., (2008) to estimate groundwater recharge in parts of the Voltaian in Northern Ghana. The results of their estimates indicated that recharge ranges from 1.5% to 11.2% of mean annual rainfall. These variations in the estimated recharge rate indicated variability of geological characteristics within the unsaturated zone and the climatic heterogeneity in the study area. The research indicated that the assumption concerning the origin of Cl might contribute to errors in recharge estimate as the Cl data for rainfall was only available for a short period. Consequently, there is a large uncertainty for Cl in rainfall, which obviously translates into large uncertainty in recharge. As only major events were sampled, the values presented in the study are likely to be underestimated (by at least 10-15 %, if the Cl detection limit is used for minor non-sampled events). However, the comparison of Cl concentration values obtained here with those of another study available for northern Ghana (Martin, 2006) reveals Cl concentrations of similar magnitude, which would suggest that Cl values estimated here are representative. To reduce the inaccuracy related to Cl flow from run-on/runoff, field profiles were carefully placed to avoid steep slopes or depressions.

In portions of the Voltaian Basin in Northern Ghana, estimates of groundwater recharge using the CMB method by Yidana and Koffie (2014) suggested recharge in the range of 1.8 to 32% of the yearly precipitation and attributed the maximum estimates of recharge to localities with open wells which act in encouraging rapid groundwater recharge. According to their findings, the most major problem in using the CMB technique to estimate groundwater recharge is that other substantial origins of chloride in the zone saturation and unsaturation might not be effectively measured, resulting in inaccurate estimations of precipitation's actual contribution to groundwater chloride. Another issue with the CMB technique, they stated, was the likelihood of fluctuations in atmospheric chloride concentration that could cause the chloride amounts in the air to differ significantly from what existed during the recharge time being calculated. Wells and other groundwater outlets analyzed in the study area were settled in sandstones and the weathered zone, where chloride was not seen to be a dominant ion.

In a similar project, Mensah et al., (2014) used the Chloride Mass Balance approach to examine the representativeness of groundwater recharge in parts of Ghana's White Volta Basin. The chloride mass balance (CMB) technique, according to their findings, works well in a tropical climate in terms of obtaining reasonably accurate groundwater recharge estimates for groundwater resource appraisal. The results of this study are consistent with those acquired through mathematical model calibration and investigations of infiltrating rainfall evaporation rates. According to the CMB's predicted groundwater recharge, the shallow aquifer system in the area receives 0.9 percent to more than 21 percent of annual precipitation. This level of groundwater recharge indicates that commercially sustainable groundwater development in the region has a bright future. The analysis also found that recent derived atmospheric water, which might have been isotopically enhanced in the weightier isotopes of the components of the water molecule during the recharge process, is the source of groundwater recharge in the terrain. Other tracers should be utilized to ensure the reliability of the results anytime the CMB approach is used in groundwater recharge calculations, according to this study.

Adams et al., (2004) published a report on the groundwater recharge evaluation of the Central Namaqualand basement aquifers in South Africa. The CMB approach underrepresented alluvial aquifer recharge related to transitory rivers (i.e., Buffels River Town and Rooifontein). Extra chloride input from run-on during recharge times could explain this. Their analysis revealed that the CMB method's calculation did not have chloride data from run-on. Except for the Leliefontein specimen, which did not match the age expectation and could be due to a sampling or scientific mistake, the recharge rates correlated with the ^{14}C ages from separate boreholes. They came to the conclusion that the alluvial aquifer recharge rates were lesser than projected. The unaccounted chloride in the surface runoff flux was the cause of this.

In a similar climatic zone, Nyagwambo conducted a study in a crystalline aquifer in the tropics of Zimbabwe in 2006 using a similar methodology. The study concluded that the strategies produced a varied recharge, around 8% and 15% of yearly rainfall. The study found that all of the recharge estimating techniques utilized in the research had the flaw of relying too much on one crucial variable, such as chloride deposited for the CMB technique.

In nations other than Africa, Hay (1997) investigated groundwater recharge on the western edge of the southern Bridger Range in Montana, and the technique was applied in a more humid location of the intermountain west where streamflow from the mountain front is considerable. Rather than springs or mountain front wells, Hay looked at chloride concentrations in groundwater using groundwater boreholes filtered at varying depths over the valley floor. Rather than recharge occurring at the mountain front via diffuse surface flow or bedrock flow, the valley aquifer was projected to be recharged by streamflow from the mountain front and precipitation falling directly on the broad valley floor below the mountain front. As a result, the CMB equation's input parameter incorporates both precipitation and streamflow. The flow of several hill-front streams shown in the study was measured weekly at the mountain front and used to determine stream inflow into the region. Two streams were tested for chloride concentrations for use in the CMB calculation on two successive days in July, beginning in the three-year course of the study. When these measurements were collected, the streams were barely recovering to base flow, according to the author. The average mean annual values measured by the National Atmospheric Deposition Program at Clancy, Montana, over the ten years preceding the study were used to calculate the concentration of chloride in precipitation. Hay's CMB estimates are sensitive to chloride concentrations in groundwater and warn about the consequences of inter-annual variability in precipitation input and spring snowmelt timing, implying that alternating warm and cool temperatures reduce peak discharges and allow maximum water infiltration into the subsurface.

Contrary to the above-mentioned Nevada and Montana research, Russel and Minor (2002) used chloride concentrations of groundwater recorded from successive elevation springs inside a mountain block to calculate CMB for large alluvial or lake sediment-filled catchments. The altitude-reliant CMB application is based on the concept of changing recharge rates along the mountain front as a function of elevation. Russel and Minor's analyses revealed a nonlinear relationship between reducing spring elevation and increased ratios of spring water chloride concentrations to average precipitation chloride concentration, with lower elevations experiencing larger improvements in relative spring water chloride concentration. This association, according to the scientists, shows that recharge occurs within the mountain block at geographically varied, elevation-dependent rates. Furthermore, recharge occurs within a given watershed as a result of numerous characteristics such as slope, aspect vegetation, and various elevation-dependent parameters. From these analyses, the authors obtain CMB estimates of mountain block recharge to valley aquifers, not as a single calculation for the mountain block, as Dettinger did, but as elevation-dependent, area-weighted recharge calculations. The chloride concentration of groundwater obtained from springs along the mountain front, according to Russel and Minor (2002), is the most vulnerable to error, with estimates of average precipitation coming in second.

Aishlin (2006) conducted groundwater recharge estimates via CMB application at multiple watershed scales within the Dry Creek Experimental Catchment in a similar project to determine the fraction of yearly precipitation that is allocated for recharge, as well as the geographical diversity within the recharge. Incorporating streamflow discharge into the CMB calculation validates the recharge estimations as net groundwater recharge rates, indicating the amount of water accessible for groundwater flow pathways in deeper mountain blocks. Annual precipitation partitioned to net groundwater recharge was estimated to be zero to eleven percent for the entire watershed from July 2004 to June 2005. Nevertheless, usage at many drainage scales within the Dry Creek Experimental Watershed revealed that over the same period, 22 percent of yearly precipitation was divided into total groundwater recharge into higher altitude sub-catchments. The data from the second research period, from July 2005 to June 2006, were generally deemed inaccurate due to the mobilization of inter-yearly deposited unsaturated zone chloride. Using time-series data from spring and stream chloride concentrations applied to hydrograph separation, the time it takes for chloride to be mobilized in the unsaturated zone, as well as lateral and vertical migration toward bedrock infiltration and stream channels, were determined. Furthermore, gain/loss analysis utilizing time-series data from stream chloride concentrations shows evidence of streamflow loss to groundwater recharge. The disparities between the water years 2004-2005 and 2005-2006 show the significance of exercising caution when addressing the assumptions behind the implementation of chloride mass balance to estimate recharge, as well as the requirement for rigorous demarcation of a suitable multi-year integration interval for calculating the mean annual groundwater recharge.

In Nebraska, the CMB was employed to predict groundwater recharge in 2012. Sand that has been well-organized and is rough dominates the terrain. Boerner and Weaver carried out this study, which used the chloride mass balance concept, which is similar to the water budget approach, to predict yearly recharge across Nebraska. According to the chloride mass balance approach, most part of this area is recharged at a rate of less than 30 mm per year. They later found that CMB provided a good estimate of recharge across Nebraska and that comparing CMB estimates to recharge estimates from other methods is valuable.

Ping et al., (2014) conducted a study on a local level that used CMB for groundwater recharge calculation in a dry mountainous terrain in Southern Interior British Columbia. According to the research, recharge estimates of 1.1-1.9 percent of precipitation at the valley floor and 1.8-2.7 percent of precipitation on mountain areas appear to be helpful. These results, however, were not compared or verified utilizing alternative approaches. According to Boerner and Weaver (2012), comparing CMB estimates with recharge estimates from soil water balance or other approaches might be beneficial. They suggested that the application of the CMB method at the drainage scale appeared to be beneficial for predicting groundwater recharge quickly.

The assumptions required by the unsaturated zone CMB approach are generally difficult to verify and may therefore be causes of error. As a result, CMB-based recharge estimation should be utilized with caution and in conjunction with other recharge methods.

2.4.2 Water Table Fluctuation (WTF)

The WTF approach has been employed in many studies (Healy and Cook, 2002; Moon et al., 2004) to estimate recharge in unconfined aquifers. Because of its consistency, simplicity of usage, and reduced cost of usage in semi-arid regions, this technique is thought to be one of the most productive and appealing (Xu and Beekman, 2003). WTF is mainly caused by a number of different hydrologic phenomena caused by natural and/or anthropogenic means (Varni et al., 2013). Given the amount of accessible groundwater-level data and the ease of estimating recharge rates from temporal variations or geographical trends of groundwater levels, the WTF method is among the most extensively employed methods for predicting recharge rates. To be utilized, this approach needs information on specific yields and variations in water levels as time passes.

The method makes no assumptions about the transport of water through the zone of unsaturation; therefore, the availability of preferential flow pathways does not limit its application (Obuobie et al., 2012). The WTF approach assumes that increases in levels of water in unconfined aquifers are caused by recharge water entering at the water table and that all other elements of the groundwater budget, including lateral flow, are null in the time of the recharge period (Crosbie et al., 2005; Mensah et al., 2014). The recharge rate can be calculated as the product of the rise in water level and the specific yield of the groundwater aquifer property. The recharge can be expressed mathematically as

$$R = \frac{\Delta h}{\Delta t} \times S_y \text{..................} 2.2$$

Where R, Δh, Δt, and Sy are respectively the amount of recharge from precipitation (LT^{-1}), change in water table head during recharge period (L), period of recharge (T), and specific yield.

The following are some of the main hypotheses that this technique is based on:

- The rise and fall of water table levels in shallow unconfined aquifers are primarily attributable to groundwater recharge and discharge;
- The aquifer's specific yield is defined and consistent during the fluctuation of the water table; and

- Water level rise can be calculated using the pre-recharge water level recession (Healy and Cook, 2002).

These assertions are not always correct, and in some cases, they could be considered flaws in this method. Variations in groundwater levels, for example, might not necessarily be due to recharge or discharge. Other causes such as the combination of evaporation and transpiration, variations in atmospheric pressure, the existence of entrapped air, and earth tides could cause it, or it could be a response to fluctuations in stream stage for boreholes located near streams (Delin et al., 2006). An estimation of the specific yield (Sy) at a depth of water table fluctuation is needed to apply the WTF method. The Sy is defined as (Healy 2010):

$S_y = \varphi - S_r$……………………………………………………………………………..………2.3

where ϕ is porosity and S_r is specific retention

Acquiring a specific yield that is reflective of a vast region has proven to be problematic in previous research. Furthermore, in contrast to the premise of a stable and defined specific yield, specific yield values change with time (Delin et al., 2006; Loheide et al., 2005). Laboratory procedures, pumping tests, water-budget approaches, and water table response to recharge are the most regularly utilized ways to determine Sy. Typically, laboratory procedures determine Sy by measuring porosity and specific retention and applying the equation (2.3). Sy readings obtained in the lab and in the field have a high degree of variability. Depending on the space between measured and abstraction wells, aquifer tests yield values of Sy integrated over rather extensive areas. Walton (1970) presented a water-budget method that requires a basin-scale water balance, ideally in winter (when the soil is almost saturated and evapotranspiration is minimal). The balance can be solved for the change in groundwater storage, and Sy may be approximated using the water table fluctuation. The water budget equation is as follows:

$$P = ET + \Delta S + Q_{off} \ldots\ldots 2.4$$

where P denotes precipitation, ET denotes evapotranspiration, ΔS is the variation in water storage (in surface storing units, zone of unsaturation and saturation), and Q_{off} denotes surface plus subsurface water flows out of the basin. The variation in groundwater storage can be represented as a function of Sy (see equation 2.2, and then Sy can be found by rearranging equation 2.4). Because the change in groundwater storage must be inferred from the difference of other water budget elements, including significantly higher water quantities, the method may have huge inaccuracies. However, this method yields a mean value for a vast region rather than a local value.

The water table's response to the recharge method is determined by determining the ratio of water table rise to cumulative rainfall (Risser et al., 2005b) for all registered events in the examined area. The quantity of open pore space accessible in the zone of unsaturation is estimated by the height of the water table increases following a rainy event (i.e., Sy). For a shallow water table, the approach is applicable.

Crosbie et al., (2005) investigated various approaches for estimating Sy in two places, calculating the water retention curve in the laboratory through examination of the water table response and pumping tests. The rainfall–water table reaction appears to produce the best estimates after testing multiple methods. As a result, this paper provides a method based on field data (rainfall and water level fluctuations) that seems to be a version of the water table reaction to recharge events technique. The approach is a graphical strategy that produces outcomes with higher reliability as the number of occurrences observed grows (Varni et al., 2013). Because the Akyem catchment has comprehensive data records, a large increase in the water table episodes may be studied, making the Sy determination less questionable. Furthermore, because of the rapid response of the water table during recharge episodes, the records are particularly well suited for this research. Because of these considerations, the graphical technique suggested in this study was chosen for Sy estimate.

Shirahatti et al., (2012) cited some challenges with the application of the method. They connected to defining a typical figure for a specific yield and guaranteeing that water levels variations are because of recharge and are not because of variations in air pressure, presence of enclosed air, or other occurrences (such as abstraction). Healy and Cooks (2002) also indicated that this method for groundwater estimation has its own limitation and stated that:

- The water table fluctuation approach is excellent for shallow water table systems with abrupt rises and falls in water levels. Wetting fronts tend to disperse across vast distances; therefore, significant rises in deep aquifers are unlikely. The approach could potentially be used in systems that have substantial unsaturated zones and only have seasonal water level changes.
- Given that recharge results differ significantly within a catchment due to disparities in altitude, geology, land surface slope, vegetation, and other conditions, the boreholes must be situated so that the observed water levels are reflective of the area as a whole.
- Because the method does not account for a constant rate of recharge, given a constant rate of recharge equivalent to the rate at which it is drained away from the water table, the water levels would remain constant, and the water table fluctuation technique could forecast no replenishment.
- Further challenges include recognizing the source of water-level fluctuations and computing a specific yield value (Beekman and Xu, 2003).

There are no specific restrictions on the range of recharge that can be approximated in the WTF approach (Obuobie et al., 2012). The WTF approach can be used to estimate subsurface storage changes over longer time intervals (seasonal or annual). This technique has been employed by several researchers for groundwater recharge estimation. Obuobie et al., (2012) used the WTF technique in Ghana to study groundwater level fluctuations and estimate recharge to groundwater in the White Volta Basin for the 2006 and 2007 study years. The results of observed groundwater levels indicate large seasonal and regional water level changes, with a span of 1240–5000 mm in 2006 and 1600–6800 mm in 2007. Because water levels rose only during the wet season, the rainfall seasonality was revealed to be the primary recharge source in the research area. The recharge to groundwater in the White Volta basin is expected to range from 2.5 to 16.5 percent of annual rainfall, with an average of 7–8 percent. The findings are acceptable and within the recommended limits of estimates found in earlier hydrogeological investigations in other sections of the Volta Basin and in a number of dry areas in Africa using water table fluctuation and other approaches. The figures reached in this study, according to their research, were slightly greater than the findings discovered in several of the earlier investigations listed. Because the specific yield figures utilized in the research were derived from earlier research rather than examined for the specific aquifers in the research area, some degree of inconsistency was to be expected. They observed that adopting specific yield estimates established for the aquifers in the research area could increase the study's trustworthiness.

Later, Krautstrunk (2012) investigated the groundwater recharge in the Nabogo River Basin, Ghana, employing the WTF and CMB methods. Calculated estimates for specific yields were selected on the basis of the hypotheses that specific yield varies in the aquifer, and the aquifer is not confined. Specific yield estimates from the literature review were used to develop a small recharge estimation and a large recharge estimation to allow for a variety of specific yield projections in the aquifer of fractured rock. The values 0.001, 0.005, 0.002, and 0.008 for specific yield were chosen from certain areas in Zimbabwe and Australia with comparable climates and fractured aquifers to depict the small valued estimation representing the "semi-confined" parts of the catchment affected by Plinthite formation, and the values 0.05 and 0.08 from current research in the White Volta River Basin to reflect the large estimation representing the "semi-confined" parts of the catchment. The recharge estimates obtained in this research agreed with the findings acquired in the area and those expected for fractured rock aquifers. Many sources have given these approaches excellent marks for accuracy when compared to other ways, and they have been praised for being a quick and inexpensive way to estimate recharge (Sibanda et al., 2009). The CMB technique gave a mean rate of 37.06mm per year, or 3 percent of annual precipitation, in the Nabogo Basin, while the WTF technique produced a spectrum of 10-143 mm per year or 1-13 percent of annual precipitation. The research also revealed that both CMB and WTF methodologies have limits when it comes to calculating recharge, particularly when specific yield values are undetermined and time restrictions are imposed on the sample time. It was further noted that when the two procedures are used together to explain a specific yield, the margins of error are lowered, reducing the uncertainty associated with utilizing only one method.

More recently, Lutz et al., (2015) employed water table fluctuations (WTF) techniques in Ghana to analyze the variation of groundwater levels and trends of recharge in the North. The variety of recharge trends between the areas are demonstrated using the WTF approach and groundwater and Sy values acquired from the literature. The variety of trends in recharge is attributable to the varied spatiotemporal precipitation and the uniqueness of hydrogeological formations at the sites. Recharge seems to be fast at three of the four sites and is computed to be a higher fraction of overall yearly precipitation than certain values in existing works. The recharge at the other area is comparable to values found in existing works, which are around five percent of precipitation. According to a seasonal analysis of groundwater level fluctuations, groundwater levels rise in response to wet-season precipitation in the areas. In certain situations, such as at Kpataribogu, the increases are significant. All areas demonstrate a total net reduction in groundwater levels during the time of the research on a yearly basis. The drop could be linked to below-average precipitation deviations in 5 of the 7 research periods. Future research should look at what kinds of precipitation conditions contribute to aquifer recharge that is close to considerable, as well as what kind of temporal buffering (storage) the system can provide to eliminate the impact of drought and over-pumping.

In other countries, Varni et al., (2013) employed the WTF method to describe the recharge of groundwater in the Pampa plain, Argentina. An approach for determining the Sy is suggested in this study; it consists of a visual process that connects increases in groundwater level to sources of rainfall. As the occurrence being monitored grows, the approach produces increasingly reliable Sy values. The WTF approach was applied after an analysis of eighteen years of daily measurements yielded a Sy value of 0.09. The recharge numbers obtained are consistent with those calculated by other writers for the same area.

The WTF method was used as an additional supplement to the RIB model in this study because periodic and reliable water level readings were accessible in boreholes that were not impacted by human activities. It provides more reliable estimations regardless of whether a piston or preferred flow recharge mechanism is used in a given location. Its temporal scope is broad, ranging from a single day to a year. Furthermore, the procedure is simple to apply, requires little data, and is inexpensive (USGS, 2008).

2.4.3 Environmental Tracer (Isotope) Technique

Recently, the techniques based on the heat or chemical isotopic tracers are gaining much importance in the estimation of groundwater recharge. The estimation of recharge using tracers is predicated on the tracer's mass conservation and the premise that the tracer flows easily with water (Sharma, 1985). Environmental isotopes give insight into the evolution of geochemistry, recharge mechanisms, rock-water relationships, and the genesis of salinity (Abid et al., 2010; Cartwright et al., 2009; Chen et al., 2006; Demlie et al., 2007; Demlie et al., 2008). As such, they serve as a tool in the characterization of groundwater flow systems (Moore, 2002). Plummer (2003) indicates that the various environmental tracers offer different forms of information about the groundwater system and the aquifer. Thus, tracers are useful tools in estimating groundwater recharge, groundwater age; trace groundwater flows directions, identify water sources to the groundwater, etc. Commonly used environmental isotopes in hydrology include but are not limited to deuterium (2_H), 18_O, carbon (13_C and 14_C) and tritium (3_H) (Fontes and Edmunds, 1989). Oxygen-18 and deuterium are described below.

2.4.3.1 Oxygen-18 and deuterium

Precipitation has more weight isotopically than the vapor remaining in the air, including more 2 H and 18O, but water vaporizing from the ocean has less weight isotopically than the water left behind. The weightier water particles have a tendency to cluster in the liquid form as phase water transforms from liquid to gas, fractionating the hydrogen and oxygen isotopes. The standard mean ocean water (SMOW) isotopic ratio can be weighed to these environmental isotopic ratios (Petersen, 2012). The isotopic makeup of the groundwater is related to the Global Meteoric Water

Line (GMWL) of Craig (1961) or a Local Meteoric Water Line (LMWL) built for a given location. The GMWL is expressed by the equation:

$\delta^2 H = 8\delta^{18}O + 10$……………………………………………………………………...2.5

Allison et al., (1984) created the δD displacement procedure for the zone of unsaturation in Australia. The displacement from the GMWL or LMWL is applied to estimate recharge using stable isotope soil moisture data plotted on a $\delta^2\mathrm{H} - \delta^{18}\mathrm{O}$ diagram and the displacement from the GMWL or LMWL is applied to compute recharge employing:

$\Delta\delta = \frac{C}{\sqrt{Recharge}}$………………………………………………………………………….2.6

$\Delta\delta$ = displacement of either D or ^{18}O from the MWL

C = gradient through the inverse of the square root of moisture flows derived from displacements from the MWL for different places. Allison et al., (1984) discovered that in Australia, C equals 20.

The actual $\delta^{18}O$ values of precipitation reaching the ground hinge on numerous elements (Chen et al., 2006; Mook, 2006), and whether these elements will affect the precipitation in the various environments are discussed further below.

- Latitude: Temperatures that are cold diminish the possibility of weightier water condensing and precipitating as rain. As a result, the higher the latitude, the more ruined the precipitation is in $\delta^{18}O$.
- Proximity to the ocean: The heaviest precipitation tends to fall first in air masses. As a result, as air masses migrate inland, precipitation gets depleted in $\delta^{18}O$.

- Altitude: Colder temperatures at greater elevations often lead to precipitation with a lower $\delta^{18}O$ content. This may not affect certain places as some areas drop into a constrained range of elevations.
- Season: Colder winter temperatures result in precipitation that is considerably diminished in $\delta^{18}O$. This can have an impact in locations where the temperature, especially during the hot summer months, can cause rainwater to evaporate before it reaches the ground.
- Amount: When compared to small rains, heavy rainfall events tend to be depleted in $\delta^{18}O$. Varying isotopic ratios may emerge from the strength and duration of rainfall events, which might influence the isotopic signature.

Isotopes are rarely used in hydrological studies in Ghana (Yidana and Koffie 2014). However, impressive and extremely beneficial results have been documented where the methodology has been utilized. Acheampong (1996) and Acheampong and Hess (2000) studied groundwater recharge mechanisms in the southern Voltaian basin, applying stable and radiogenic isotope data from groundwater, surface water, and precipitation. Their research was crucial in attempts to link groundwater recharge from the Volta Lake to groundwater in the Voltaian, as it found no evidence of groundwater recharge from the Volta Lake. Later, Jorgensen and Banoeng-Yakubo (2001) linked measured elevated salinity levels of groundwater to the impacts of intrusion of saline water along Ghana's coastal belt utilizing comparable isotopic and hydrogeochemical data from groundwater in the shallow aquifers of the Keta Basin in the Volta Region. In some parts of the North, Pelig-Ba (2009) studied the stable isotope contents of groundwater and surface waters and predicted that the groundwater is directly recharged from local precipitation. However, as it travels from the atmosphere to the water table, it evaporates. In a similar area, Yidana et al., (2014) employed stable isotope data to link meteoric water to groundwater in areas of the Voltaian

Basin's Nabogo sub-catchment. Their study found that groundwater recharge within the weathered zone is of meteoric origin and shows indications of considerable evaporation of rainwater in the spectrum of 34 to 70%, which was attributed to high temperatures and low relative humidity.

The capacity of isotope tracers to maintain the current settings at the places and time of groundwater recharge determines their utility (Yidana and Koffie 2014). As a result, recently recharged groundwater can be distinguished from Paleo-recharged groundwater. For example, Datta (1999) examined the occurrence of groundwater and recharge in Delhi's National Capital Territory (NCT) employing 18O isotope profiles. These signals indicated that groundwater in Delhi region wells is a combination of changing quantities of various water sources, and the aquifer in the area is not a homogeneous system on a lateral scale.

In the same country, Négrel et al., (2007) employed records of stable isotope to identify the nature and manner of groundwater recharge in an Indian river basin. They linked groundwater recharge to rainfall in their study region, implying that infiltrating rainwater moves quickly across the unsaturated zone. Their study was able to classify the groundwater in the area into three separate end-member types. The researchers identified places in the basin that are highly impacted by human activities as a result of this investigation.

In Italy, Rapti-Caputo and Martinelli (2009) used a familiar approach to distinguish water underground in distinct stratigraphic layers according to variations in stable oxygen and hydrogen isotopes signatures. They presented a compelling argument for blending local surface waters with confined waters in the region. The use of stable isotope data in water studies can be seen in the assessment of the impacts of seawater on groundwater resources (Ma et al., 2007), the distinguishment of the lengths of regional flow systems (Salem et al., 2004), and the distinctions of groundwater recharged under varying climatic conditions (Bogaard et al., 2007; Dafny et al., 2006; Ophori, 1999; Peng et al., 2007, 2010a, b, 2012a, b; Rozanski, 1985;).

Hence these environmental isotope tracers can be used to adequately characterize groundwater recharge in a basin with a high level of certainty. However, it should be noted that recharge computed by tracers' technique is an average transport velocity grounded on Darcy's law and is not the most recommended technique when the recharge is very small (below 20 mm/year) (Wu et al., 2016).

2.4.4 Cumulative rainfall departure method (CRD)

This technique is based on the water balance equation and is predicated on the assumption that conditions of equilibrium evolve in an aquifer over time, i.e., the mean rate of losses equaling the mean rate of the system's recharge. In other words, rainfall events induce changes in groundwater levels. Monthly rainfall and groundwater level data are required, as well as details on aquifer components (storativity), abstraction, and the size of the recharge region. This approach enables the determination of aquifer properties and parameters such as i) recharge or specific yield, ii) the effect of pumping or iii) rainfall's natural reaction to being ascertained. The cumulative rainfall departure series is mathematically expressed by the relationship (Bredenkamp et al., 1995): The CRD model is consequently employed to simulate water level fluctuations. Bredenkamp (1995) defined CRD as follows:

$$CRD_i = \sum_{n=1}^{i} R_n - k \sum_{n=1}^{i} R_{av} (i = 1,2,3, \ldots, N) \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots .2.7$$

Where R is the rainfall in the storage, with subscript "I" showing the i-th month. "av" showing the mean of rainfall intensity. At the same time, k is the permeability coefficient, which is the result of the borehole abstractions.

It was believed that a Cumulative rainfall departure has a direct correlation with a monthly change in water level. Bredenkamp et al., 1995 derived:

$$\Delta h_i = \left(\frac{r}{S_y}\right) * CRD_i \dots\dots\dots\dots\dots\dots\dots\dots\dots\dots\dots\dots\dots\dots\dots\dots\dots\dots\dots\dots\dots\dots\dots\dots\dots\dots\dots\dots\dots\dots\dots\dots\dots 2.8$$

Where r is a fraction of the Cumulative rainfall departure that leads to recharge from rainfall, which was calculated as the departure (difference) from normal rainfall at a period.

Equation 2.8 may be used to determine the recharge-to-aquifer storativity ratio using linear regression between "CRD" and "Δh_i" (Bredenkamp et al., 1995).

A new CRD has been formulated to account for a trend in rainfall data as

$$CRD_i = \sum_{n=1}^{i} R_n - \left(2 - \frac{\sum_{n=1}^{i} R_n}{R_{av}}\right) \sum_{n=1}^{i} R_t \; (i = 1,2,3, \dots\dots N) \dots\dots\dots\dots\dots\dots\dots\dots\dots\dots\dots\dots\dots\dots\dots .2.9$$

Where "R_t A threshold value indicating aquifer boundary conditions" is generated during the modeling process. It may range from "0 to R_{av} with 0 representing an aquifer being closed and R_{av} meaning that the aquifer system is open, possibly being regulated by outflow" (pumping or natural outflows). Note that Equation 2.9 reduces to Equation 2.8 if rainfall events Ri do not show a trend (R_t =R_{av}). In this case, the cumulative rainfall average would conform to R_{av}.

If the other stress is essentially constant, it is considered that the CRD is the main factor behind the monthly water level variations. If the cumulative departure is positive, the groundwater level will rise; if it is negative, it will fall. Since:

$$rCRD_i = S_y \left[\Delta \mathrm{h}_i \frac{Q_{\mathrm{pi}}+Q_{\mathrm{out}}}{AS}\right] \qquad 2.10$$

where: r is that fraction of a CRD that leads to recharge, S_y is specific yield, $\Delta \mathrm{h}_i$ is water level change during the month i (L), Q_{pi} is groundwater abstraction (L^3 /T), Q_{out} is natural outflow, A is recharge area (L^2).

The method indicates that it is assumed

- When the rainfall exceeds the average for the specified time frame, recharge takes place.
- There will be no recharge if the rainfall is less than the mean (i.e., natural losses).
- Pumpage's impact is incorporated within the brackets, i.e., its impact is minimized based on the value of the fraction of rainfall that accounts for a recharge, which is typically less than 1.

The CRD approach struggles to explain the situation in which continuous departures are negative but measured water level rises, and the physical meaning of parameter r, which reflects the proportion of recharge, in the CRD method is problematic (Sun et al., 2013).

A large number of studies have been conducted to estimate groundwater recharge using the CRD method. Butterworth et al., (1999) studied the hydrological dynamics and water resources management in an arid-semi arid setting at an experimental watershed in south-east Zimbabwe. Sufficient rainfall datasets were employed to optimize groundwater levels in the research area from 1953 to 1996 in order to assess the effects of rainfall fluctuations on groundwater. Two distinct modeling approaches were used. Groundwater levels were forecasted as a factor of drainage, specific yield, and water table height using an ACRU soil water balance model that modeled daily rainfall and volatile demand. Furthermore, to simulate groundwater levels from monthly rainfall, the CRD technique was proposed. Both techniques successfully simulated observed groundwater levels from 1992 to 1996, and sufficient modeled patterns in previous levels were relatable. The findings revealed that substantial fluctuations in groundwater levels were a natural aspect of a shallow aquifer's response to rainfall variability. Long-term patterns in groundwater levels were evident, reflecting the influence of rainfall cycles. The average end-of-dry-season water level in the late 1970s was simulated to be over 3m higher than in the early 1990s. They noted that the CRD method is better suited in the routine use in the management of abstraction from water points and in studies where fewer data are available.

Setiawan et al., (2011) used the cumulative rainfall departure approach to estimate the possibility of landslides on the influence of climate variation in South Sumatera Province, Indonesia. Cumulative rainfall departure (CRD) was able to evaluate groundwater table recharge as the trigger factor of landslide by using the Geographical Information System (GIS) process. The highest landslide probability occurred in June, with the probability range from 0 – 0.64 and then divided into five levels of hazard: very minimal, minimal, medium, extreme, and very high. The results showed that landslides hazard probability map for 30 years projection, by using constructed spatial data-sets and model of slope stability that has done before.

Similarly, Petersen (2012) studied the groundwater recharge mechanisms and interconnections between groundwater and surface water from a conceptual standpoint in the Kruger National Park (KNP), South Africa. The CRD and stable isotopes (2H and 18O) were employed as the two techniques. An adapted version of the CRD, which includes the system's prolonged and brief-span memory, was utilized to find potential recharge mechanisms. Furthermore, the CRD approach was used to simulate a valid rebuilding of prolonged groundwater level patterns utilizing monthly rainfall data with respect to the mean rainfall throughout the whole time series 1936-2009. The stable isotopes (2H and 18O) were taken monthly for almost a year from accumulated rainfall tests, surface water (streams and rivers), and groundwater from wells (May 2010 to July 2011). The groundwater isotope make-up was utilized to determine if recharge was instantaneous or postponed. In order to identify surface-water interactions, surface water from streams and rivers was in comparison to groundwater in terms of isotopic make-up. The stable isotope data gathered from cumulative rainfall, groundwater, and surface water (streams and rivers) revealed that groundwater evaporates before infiltration. Due to the excessive vaporization levels in the KNP, the pattern of minor rain that falls has a small or no effect on recharge. The cumulative rainfall departure study of groundwater level changes revealed that aquifer recharge is affected by dry and wet seasons that last 6 to 14 years. Straightforward recharge by piston flow and indirect recharge via favored routes, mainly streams and rivers, are the primary recharge mechanisms in the southern half of the KNP, according to the CRD model and stable isotopic observations.

A revised and updated method called the rainfall infiltration breakthrough (RIB) was created by Xu and Van Tonder (2001) based on the CRD model. The RIB technique is similar to the CRD model in that it replicates groundwater levels, but it varies in that it considers the way in which recharge happens. As a result, it offers a tremendous opportunity to describe how recharge works under various settings. The model's limitations include the fact that it is suitable for shallow unconfined aquifers with poor transmissivity. As a result, it becomes problematic when there is a scarcity of data and when varied hydrogeological circumstances exist (Sun et al., 2013).

The rainfall infiltration breakthrough (RIB) model has been applied in many countries and has produced successful results. Sun et al., (2013) employed this model for recharge estimation in west coastal South Africa. In their study, they presented a monthly association between WTF and rainfall events in the RIB model for groundwater recharge calculation. Recharge estimates on a daily and monthly basis utilizing the improved RIB technique were provided in two research locations, one in a Riverlands sandy alluvial aquifer and another in an Oudebosch shallow unconfined aquifer of the Table Mountain Group (TMG), proceeded by sensitivity analysis. When employing long time series, the monthly RIB outcomes are more reasonable than those based on daily data, according to a correlation assessment between rainfall and measured WLF data at the monthly and daily scales, as well as recharge estimates obtained from other techniques. The sensitivity analysis revealed that the RIB model's recharge rate is particularly responsive to the variable of specific yield; thus, selecting an appropriate reflective specific yield of the aquifer must be done with care. Their research proved that the RIB model used in the two situations might be utilized to calculate the recharge of groundwater in similar places with suitably long time periods of rainfall and groundwater levels. According to the research, a consideration of the model's limitations shows that the model is appropriate for low transmissivity shallow unconfined aquifers. They also noted that due to a lack of data, they could only model daily and monthly recharge estimates. They concluded that the model's utility should be examined further using more data gathered in diverse hydrogeological settings and climatic areas.

Recently in the same geological and climatic zone, Mutoti (2015) assessed groundwater recharge in the upper Berg River basin, South Africa, applying chloride mass balance. The chloride mass balancing approach was utilized on a pilot scale in conjunction with the RIB and WTF methods to assess groundwater recharge and standardize the accessibility of groundwater. The average groundwater recharge estimates computed using the CMB, RIB, and WTF methods, respectively, were 27.6 percent, 23.67 percent, and 22.7 percent based on the findings. The findings suggest employing these methodologies has the ability to calculate groundwater recharge at the quaternary system. This is the main component for managing and regulating water in the country. These findings are consistent with past research performed in a similar basin, which found that average groundwater recharge ranged between 18.6 and 28 percent of precipitation. According to the study, no information relating to the precise specific yield of materials within the research region was accessible. Specific yield values for the proposed investigation were calculated using data from the aquifer test and the rainfall/groundwater level association. According to the research, these specific yield values were then compared to values taken from the literature.

Ahmadi et al., (2014) mapped and quantified the recharge rate in the semi-arid Neishaboor Plain, Iran. To analyze groundwater recharge, RIB and CRD techniques were used in the study. These two methodologies were utilized to simulate groundwater levels, and the method of efficiency was employed to reduce the root mean square (RMS) error between the observed and simulated groundwater levels. These findings showed that the observed readings perfectly correlated with the simulated groundwater levels. These two methods were then used to map the recharge areas. By using the CRD model, an average of 13 and 27 percent of CRD and return flow led to groundwater recharge for the entire research region, whereas RIB values were 11 and 19 percent, respectively. Using RIB and CRD, the mean recharge rate on a yearly basis for Neishaboor Plain was calculated to be 160 and 164 mm, respectively.

More recently, Majdabady et al., (2020) calculated the recharge of groundwater from agricultural irrigation and rainfall in Iran's Shyramyn Plain. The RIB model was used to estimate groundwater recharge from rainfall “r” and irrigation return flow “λ”. According to the outcomes of the work, the number of r ranged around a rate of 24.6 percent within the area, with a low and high of 1.1 and 27.0 percent, accordingly, while λ ranged with a low and high of 0.0 and 33.0 percent, accordingly. The most focus was traditionally placed on recharge from rainfall when estimating groundwater recharge. However, the study's findings revealed that recharge from irrigation return flow contributed a significant part in groundwater recharge, accounting for 66.3 percent greater than recharge from rainfall. The study concluded that the model allows the estimation of unidentified hydraulic characteristics and natural recharge with an acceptable spectrum of possible outcomes based on collected observation data.

There are numerous methods for estimating groundwater recharge. The majority of them necessitate a large number of field details; as a result, their practical application is restricted. The RIB model has been used in this study to get past this restriction. The RIB method, which is predicated on the water balance concept "(rainfall–groundwater level relationship)," was applied in this study because it requires few non-deterministic data such as groundwater level data, rainfall, aquifer properties, and inflow and outflow dataset. It is also a more reliable approach since actual mechanisms of rainfall water reaching the groundwater table and studied sources of recharge are considered. This method provides a spatial distribution of recharge that could be utilized as an excellent starting point for numerical groundwater modeling.

CHAPTER THREE

METHODOLOGY

3.1 Introduction

This chapter gives an outline of the research methods that were followed in the study. It provides information on where the data were collected, showing, therefore, the level of their accuracy. It further discusses how the data were processed to achieve the set of specific objectives.

3.2 Desk Study

The Desk study was conducted to acquire more information about the study area. The desk study entailed reviews of the geology, hydrogeology, geochemistry, as well as topographic maps covering the study area. The review also centered on various publications which employed techniques and procedures to estimate groundwater recharge. The sources of this information include Newmont Ghana, websites, and the University of Ghana.

3.3 Recharge Modelling

Lithological log data obtained from AMEC Geomatrix (2010) and literature were thoroughly analyzed and compared with the known geology and hydrogeology of the study region to maintain consistency. The data included details on the borehole IDS, their spatial positions (longitudes and latitudes) and locations, well depths (in meters), hydrogeological unit descriptions, and lithology types. The methodology is tested using data for the Birim North District and can be used and adapted for better water management in the future.

Xu and Van Tonder (2001) developed the RIB approach based on the cumulative rainfall departure (CRD) method utilized in South Africa (Bredenkamp et al., 1995). Both methods utilize the association between the departure of rainfall from the mean rainfall of a preceding time and water level fluctuations. The RIB formula is defined as:

$$RIB(i)_m^n = r\left[\sum_{i=m}^{n} P_i - \left(2 - \frac{1}{P_{av}(n-m)}\sum_{1=m}^{n} P_i\right)\sum_{1=m}^{n} P_t\right] \quad \text{3.1}$$

(i=1, 2, 3, …I)

(n=i, i−1, i−2, …N)

(m=i, i−1, i−2, … M)

$m < n < 1$

Where: RIB(i) is the cumulative recharge from m to n rainfall events.

r is the fraction of cumulative recharge by rainfall which contributes to the RIB (recharge percentage);

P_i is the rainfall amount at the i_{th} (daily, monthly, or annually)

P_{av} denotes the average precipitation across the whole time series

P_{t} is a threshold value indicating aquifer boundary conditions (P_{t} ranges from 0 to P_{av})

The value of 0 denotes a closed aquifer system, which means that the recharge at i_{th} time scale is solely dependent on preceding rainfall events from P_{m} to P_{n} ; while the value of Pav represents an open system, which indicates that the recharge at the i_{th} time scale is dependent on the difference between the rainfall's average of the whole time series and the mean rainfall of preceding events of rainfall from P_{m} to P_{n}. During the simulation process, both r and P_{t} values are determined.

Under natural conditions, it is considered that the recharge of groundwater using the RIB method is linearly proportional to water level changes. The link between water level fluctuations and natural rainfall can be illustrated by Equation (3.2)

$$\Delta h_i = \left(\frac{1}{S_y}\right) * RIB(i)_m^n \dots\dots\dots\dots 3.2$$

Where:

Δh_i denotes the water-level fluctuation, that is equivalent to the difference between the mean water level of the whole time series and the measured water level at i_{th} time scale; A positive value reflects a rise in water level, whereas a negative value represents a fall in water level.

S_y is the aquifer's specific yield.

Equations (3.1) and (3.2) revealed that the water-level fluctuation at i^{th} time scale (daily/monthly/annually) is influenced by preceding rainfall events from P_m to P_n (m<n), with a weighting factor that is a function of the moving average of a rainfall time series.

$$\left(2 - \frac{1}{P_{av}(n-m)} \sum_{1=m}^{n} P_i\right) \dots\dots\dots\dots 3.3$$

It is not always stable and can be either positive or negative, which is dependent on whether or not the amount of rainfall during the study period surpasses the moving average rainfall.

The relationship between the RIB model and water level fluctuations can be written as (Sun et al., 2013):

$$\Delta h_i = \left(\frac{1}{S_y}\right).\left(RIB(i)_m^n\right) - \frac{Q_p + Q_{out} + Q_{oth}}{A.S_y} \dots\dots\dots\dots 3.4$$

(i=1, 2, …I)

where:

A is the area of the catchment,

Q_p, Q_{out} *and* Q_{oth} denote groundwater abstraction, outflow, and volume changes caused by other activities, respectively.

Nevertheless, when the continuous departures are negative, but the observed water level rises, the CRD and earlier RIB approaches struggle to discuss these circumstances and the physical meaning of parameter r, which denotes the recharge percentage, in the CRD and preceding RIB methods is unclear. The formulation of the CRD approach is shown below.

$$CRD_i = \sum_{n=1}^{i} R_n - k \sum_{n=1}^{i} R_{av} (i = 1,2,3, \dots, N) \dots \dots \dots \dots \dots \dots \dots \dots \dots \dots \dots \dots \dots \dots \dots \dots \dots \dots \dots \dots 3.5$$

Where R denote the amount of rainfall with subscript "i" indicating the i-th month, "av" the average, and $k = \frac{Qouti+Qpi}{ASy}$, k=1 indicates that pumping does not occur and k>1 if pumping and/or natural outflow takes place. It was believed that a CRD has a linear relationship with a monthly water level change. Bredenkamp et al., 1995 derived

$$\Delta h_i = \left(\frac{r}{S_y}\right) * CRD_i \dots \dots \dots \dots \dots \dots \dots \dots \dots \dots \dots \dots \dots \dots \dots \dots \dots \dots \dots \dots \dots \dots \dots \dots \dots \dots \dots \dots \dots \dots \dots \dots 3.6$$

Where r is a fraction of the CRD that leads to recharge from rainfall, which was calculated as the departure (difference) from normal rainfall at a period.

Instead of considering the departure from average, the difference between consecutive departures should be considered as a recharge. If the difference is positive, the groundwater level will rise, and vice versa; recharge at the i^{th}time scale can be expressed as:

$$\mathrm{Re(i)} = RIB(i)_m^n - RIB(i-1)_{m'}^{n'} - \left(\frac{\Delta \mathrm{Q}}{A}\right)$$

$$= [\Delta \mathrm{h(i)} - \Delta \mathrm{h(i-1)}].S_y - \left(\frac{Q_{\mathrm{p}} + Q_{\mathrm{out}} + Q_{\mathrm{oth}} - (Q_{\mathrm{p}} + Q_{\mathrm{out}} + Q_{\mathrm{oth}})}{A}\right) \ldots\ldots 3.7$$

(*i*=2, 3, …*I*, *m*<*n*<*I*, *m'*<*n'*<*I*, *n*-*m*+1=*n'*-*m'*+1. *Re(i)*>0)

$$\mathrm{Re(i)} = \Delta \mathrm{h}(1).S_y - \left(\frac{Q_{\mathrm{p1}} + Q_{\mathrm{out1}} + Q_{\mathrm{oth1}}}{A}\right) \ldots\ldots 3.8$$

$$T_{\mathrm{Re}} = \mathrm{Re(i)} + \sum_{i=2}^{n} \mathrm{Re(i)} \qquad (\mathrm{i} = 2, 3, \ldots \mathrm{I}) \ldots\ldots 3.9$$

where:

Re (1) denotes the recharge for the first time step, and

Re(i) is the recharge estimate at the i^{th} time that could be daily, monthly, or annually.

T_{Re} is the total recharge in mm for the whole time series. If the value of Re(i) in Equations (3.7) and (3.8) becomes negative, no recharge on the i^{th} time scale is assumed.

3.3.1 Data requirement in RIB model

The CRD and RIB models required non-deterministic data such as groundwater level observations, rainfall, aquifer parameters such as specific yield, transmissivity, unsaturated thickness, lateral inflow and outflow, and groundwater extraction dataset. A field survey was also undertaken to verify the parameters necessary for recharge processes.

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Table 3.1: Sampled boreholes and their coordinates

ID	Easting	Northing	Elevation	Average Static Water Level
MW-8	719413.3	701927.6	177.00	1.24
NMW-8S	721018.5	701767.9	155.00	2.02
NMW-10S	716511.7	702502.8	198.00	18.72
NMW-12S	721335.1	700021.4	152.00	8.61
NMW-17S	718217.2	702177.5	185.00	20.45

3.4 Conceptual Model recharge processes

It has been said that a conceptual model must be constructed before a quantitative assessment of recharge in a given region can be done. A recharge process conceptual model seeks to provide answers to the questions of how, where, when, and why recharge occurs (Healy, 2010).

Static water levels were measured from 2011 to 2017. Depth to the bedrock in the area, according to AMEC Geomatrix (2010), is about 150m. However, the deepest Hydraulic heads recorded during this period were about 180m suggesting that the levels recess into the bedrock at this location. Levels were recorded monthly. The topography of the model area is defined by steep hills in the center surrounded by undulating lowlands. The landscape is mostly vegetated, with subtropical forests and agriculture dominating. The highest elevation of 200 m is found in the western part of the study area, while the lowest elevation of 150 m is found in the eastern part. Figure 3.1 shows the contour of the study area.

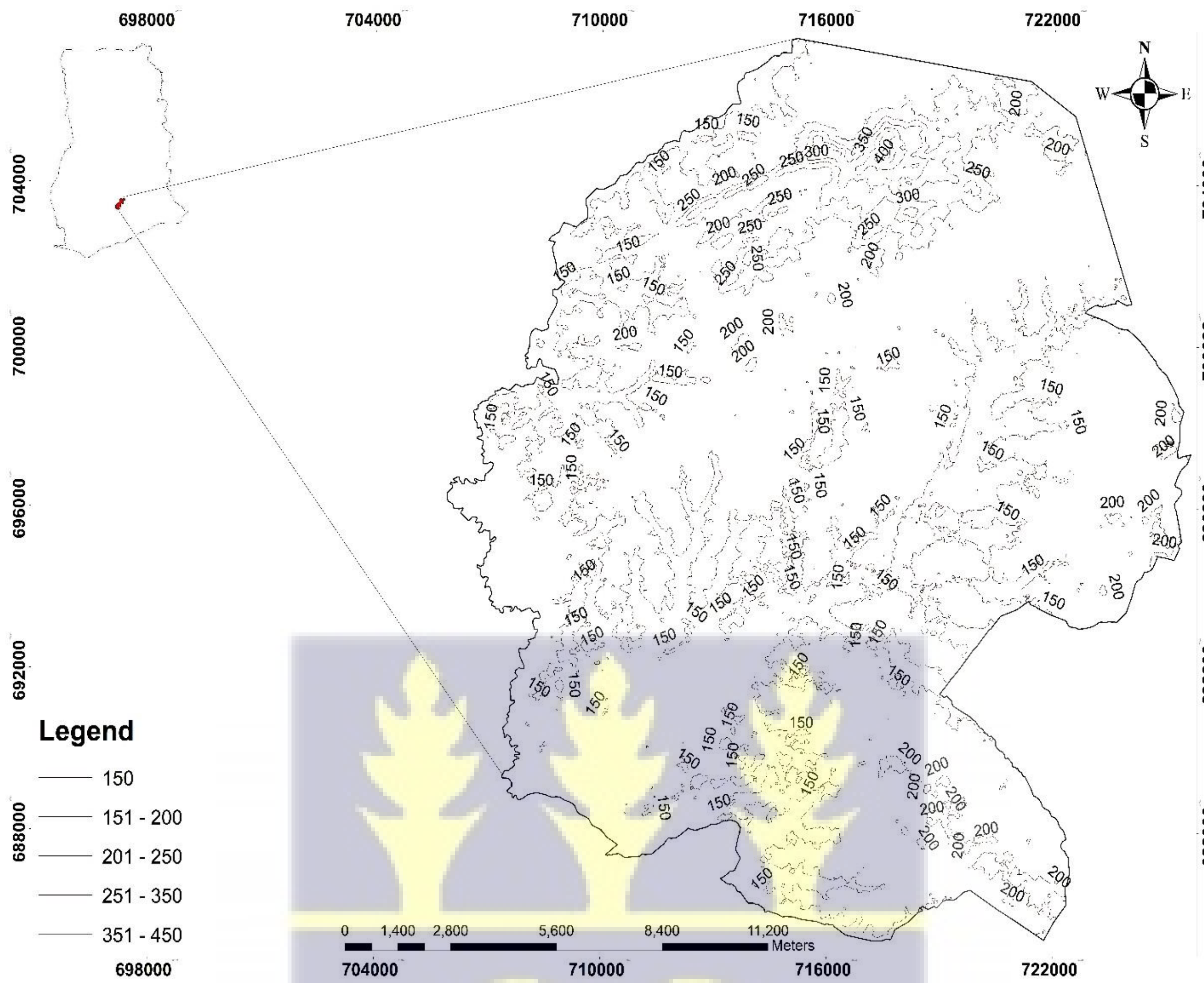


Figure 3.1: Contour map of the study area

In the study area, there are two rivers, the Pra river, and the Birim river. The major source of water for irrigation in the study area is groundwater. With one exception, all surface water features are ephemeral and flow in response to wet season precipitation. The lower Pra River reach is perennial, with groundwater discharging to the river sustaining it throughout the dry seasons. Peak stream discharge lags behind peak precipitation. Approximately 10% of the average annual precipitation enters the surface water system as runoff, and 1.2 % enters as baseflow. Infiltration of precipitation is the primary source of groundwater recharge in the model area. Recharge occurs during the wet season, with the greatest contribution most likely coming after heavy precipitation events. The study area receives 1,347 mm of rain each year on average. The majority of rainfall falls during the two wet seasons. The major rainy season begins in March with moderate rainfall, increases until June, and then decreases until July. The minor rainy season typically lasts from September through November. According to the work of Singh et al., (1984) and Taylor and Howard (1996), significant recharge only occur when precipitation intensities exceed 10 mm/day on two or more consecutive days. Groundwater discharges from the model region as base flow to rivers, evapotranspiration from riparian and wetland areas, and groundwater pumping from village wells for household use. Another source of groundwater discharge is evapotranspiration through wetlands and riparian regions. Precipitation, which averages 1,347 mm/year, has a significant impact on surface water flow and groundwater recharge, but approximately 82 percent is lost annually due to evapotranspiration. Approximately 10% of yearly precipitation is a runoff in streams and rivers, with the remaining 8% entering the groundwater system. Of the eight percent (80,350 m3/day that enters the groundwater system as infiltrating precipitation and runoff, around one percent (12,470 m^3/day) remains streams as baseflow, and about seven percent (67,650 m^3/day) is lost directly from groundwater by evapotranspiration (AMEC Geomatrix, 2010).

With the use of ArcGIS products and the Thiessen approach, the study area (Figure 3.2) was partitioned based on observation wells (5 OWs), and then transmissivity, specific yield, monthly rainfall records, abstraction wells, lateral groundwater inflow, and outflow data were provided for each sub-zone. Monthly rainfall from November 2011 to June 2017 (6 years) was used in the creation of the conceptual model in this study.

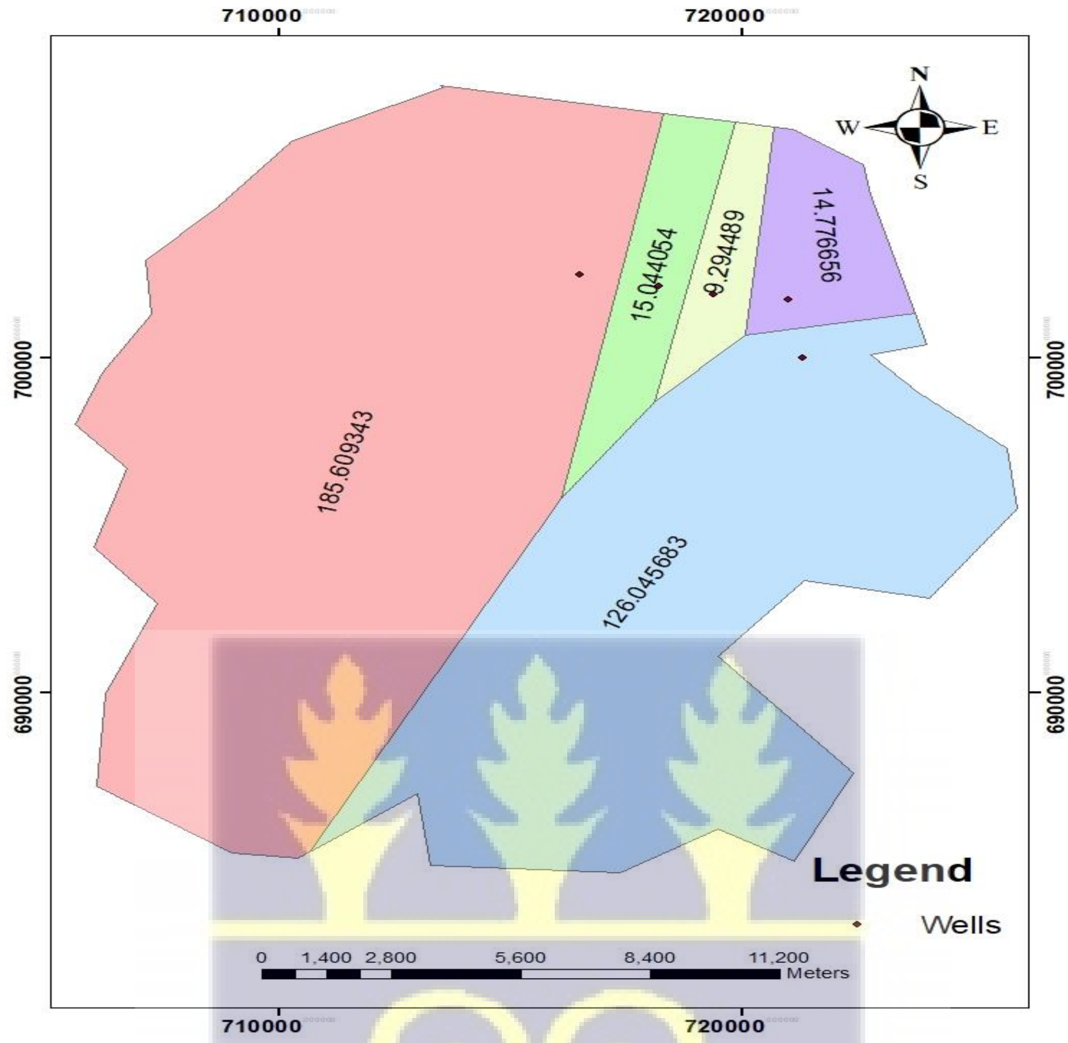


Figure 3.2: Location map and polygon of each borehole

A conceptual model was constructed for each of the five polygons (one for each observation well). Following the implementation of the Thiessen polygons, groundwater levels in the observation wells were measured monthly and documented in Excel spreadsheets for the length of the project. The pumping rate, monthly rainfall, inflow and outflow records, and groundwater level records for each polygon were then listed against the corresponding groundwater level data (observation well). These boreholes were chosen because they are fitted with automatic data loggers that record time-series data; hence, secondary data (monthly time series) from earlier periods were obtainable for these wells. The rainwater infiltration breakthrough model requires that these five monitoring boreholes be positioned away from bodies of water. Water level fluctuations from monitoring boreholes should not be close to surface water bodies unless the sources and sinks can be reliably defined.

The water table fluctuation method was used to estimate groundwater recharge in order to evaluate, enhance, and confirm the rainfall infiltration breakthrough model estimates by comparing such groundwater recharge estimates. Net recharge was estimated using seasonal changes in the water table in the research region. The water table fluctuations approach for calculating net recharge is appropriate for places with narrow vadose zones where water levels react immediately to precipitation (Saidi et al., 2009). The water table fluctuations were calculated using static water level observations from each of the five wells averaged over a six-year period. There are several methods for estimating the rise in water level (Atta-Darkwa et al., 2013; Obuobie et al., 2012). The difference between the highest and lowest mean water levels at each borehole location was used to calculate the water table fluctuation value in this study. The net recharge for each well location was calculated by multiplying the water table fluctuation data by the specific yield of the aquifer media type at each well location. In the Keta Strip in Ghana, Kippo (2012) employed this method to predict water level rise and net groundwater recharge in a quaternary aquifer.

The specific yield (Sy) at a depth of water table fluctuation must be determined in order to use the WTF and RIB techniques. Sy is typically determined using laboratory approaches, pumping tests, water-budget methodologies, the response of the water table to recharge, and regression analysis (Atta-Darkwa et al., 2013; Martin 2005; Mutoti 2015). The current study estimates Sy using literature and linear regression methods, as well as comparing the value to that of earlier studies. A graphical approach was presented to estimate Sy: a graph of rainfall values is displayed against water table fluctuations. The inverse of the slope of a line traced through the origin and instantly above all of the observed values gives the highest value of the specific yield. Because of the extensive data record, this Sy estimation method was chosen to provide a stable Sy value. This method is similar to determining the precipitation–groundwater rises ratios of all measured events and identifying the smallest value, but the graphical method has the advantage of clearly indicating whether there are several points near the lowest value of Sy or if there is one point that needs to be revised. The difficulty in identifying the specific yield value has led to a large range of values for the similar textural group being published in various literature. Natural heterogeneity in geologic material, variances in determination methods, and, most importantly, the amount of effort spent establishing the specific yield value have all been cited as reasons for these disparities (Gupta and Gupta, 1999 cited in Martin, 2006).

3.5 Calibration and Validation Methodology

The RIB model calibration was carried out in an easy-to-use Excel spreadsheet and coded with a visual basic application (VBA) developed by Xu and Van Tonder (2001). The software enables one to analyze, manipulate, estimate groundwater recharge and visualize data based on the measured rainfall and groundwater levels. Using the least-squares method, a solver function is used to minimize the difference between the measured and computed WLF values. The application may also fill voids in groundwater-level data and simulate water level fluctuations using abstraction and rainfall data by employing key parameters acquired from existing rainfall and water level data (such as length of related rainfall events, r value, and time lag). Appendix 2 gives an illustration of a screen printout of the RIB graphical interface. Calibration was performed for each individual observation well model to determine the fraction of cumulative recharge by rainfall (r), specific yield (Sy), lag time, and length of related rainfall events. The calibration target (objective function) was aimed at reducing the difference between the estimated and observed groundwater table elevation in each individual observation well (individual polygon).

The symbols used in the RIB program have been defined in Appendix 3, including the units and data types (e.g., one value/time series, input/output, etc.). Specific yield (Sy), rainfall time series, observed groundwater level fluctuations, catchment area (A), groundwater sinks/sources, and parameters for estimating the time lag between recharge and rainfall events are all input data. The sink/source terms ($Q_{\text{pumpage}}, Q_{\text{out}}, Q_{\text{oth}}$) were left empty due to the absence of data (in places where groundwater abstraction was modest, the impacts of these parameters on water-level changes were deemed minimal in comparison to rainfall input).

Before starting data inputs, the time scale was manually defined. This is dependent on the type of water level and rainfall data available (monthly, annual, or daily basis). By clicking the 'Start button, the time scale was activated. By pressing the 'Graph' button, groundwater level fluctuations and recharge were also estimated. The chart was automatically updated and showed measured/computed groundwater level fluctuations (using the CRD and RIB models), rainfall, and recharge estimates using the RIB approach. The time length and lag were adjusted to fit the estimated and observed WLF value curves.

3.6 Sensitivity analysis

Following the calibration of each model, it is recommended that a sensitivity analysis be performed. Sensitivity analysis is necessary for calibration, optimization, risk assessment, and data collection. The goal is to verify the model's stability in the face of modest alterations in some of the main aquifer parameters. A model that is particularly sensitive to any of the parameters is considered unstable and so unreliable for forecasting scenarios. Monthly rainfall and water level fluctuations were used as inputs in this study to analyze individual implications on the projected recharge rate.

3.7 Scenarios Analysis

A model can be used to forecast future groundwater balance, flow, or contaminant transport conditions, among other things, according to Kumar (2015). The model may also be used to evaluate different remediation alternatives. However, any model prediction in a groundwater recharge analysis is no better than an estimate due to mistakes and ambiguities. As a result, all simulated results should be expressed as a variety of possible results that take into consideration the assumptions made as well as the ambiguity in the model input data and parameter values. In this study, the RIB model was used to run simulations in one of the boreholes to forecast trends of groundwater level under conditions of reduced rainfall. As a result, the following scenarios were run at one borehole:

- Baseline conditions: Applying measured rainfall records from January 2012 to July 2017; whilst maintaining other parameters.
- Reduction in measured rainfall by 10%; while maintaining other parameters.
- Reduction in measured rainfall by 20%; while maintaining other parameters.

- Reduction in measured rainfall by 30%; while maintaining other parameters.

Flowchart of Modelling Protocol

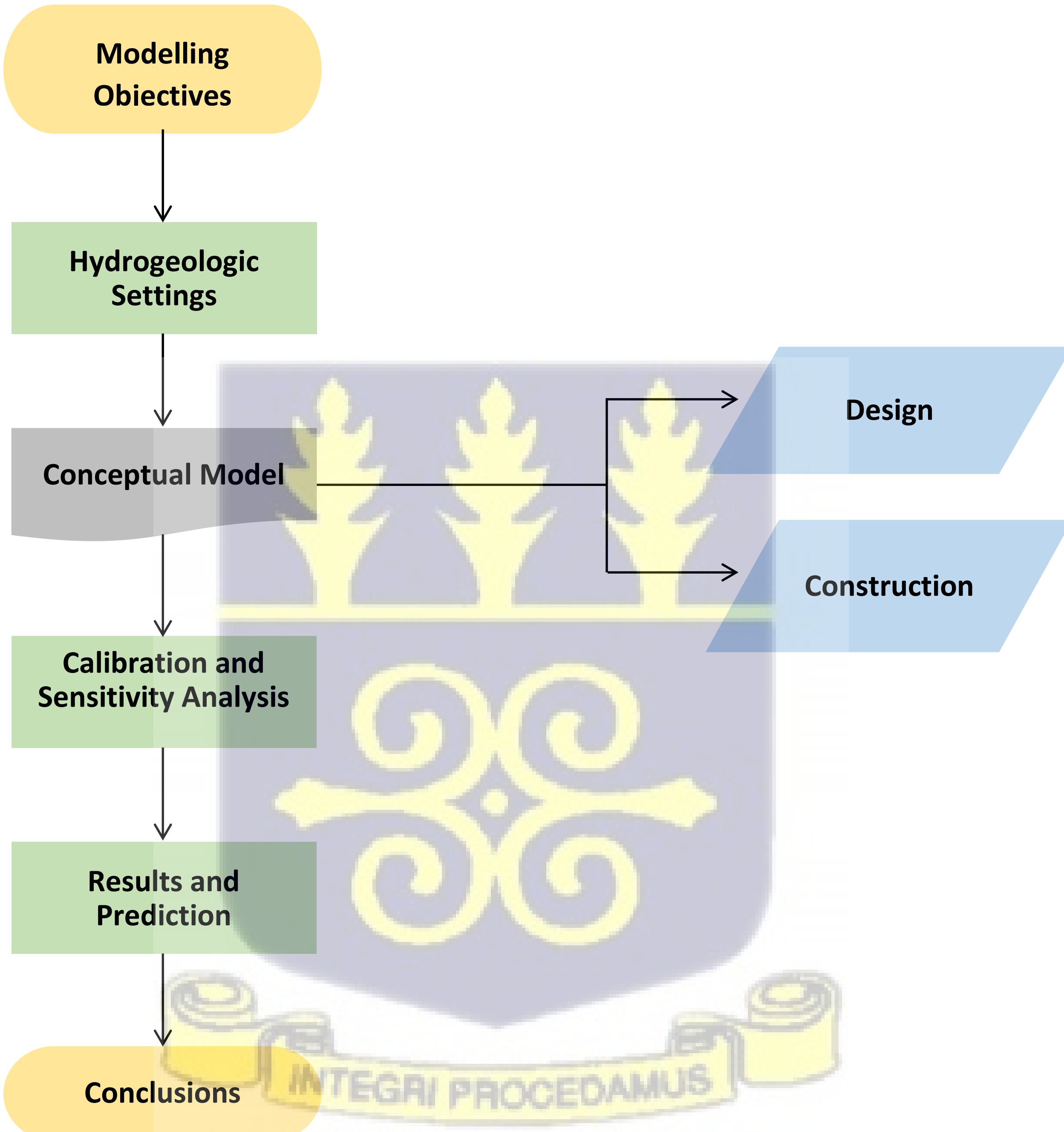


Figure 3.3: Groundwater Recharge Modelling Process

3.8 Study Limitations

The study time was limited due to the time required for data collection and processing. The study's main limitation was the difficulty in measuring the specific yield of an aquifer when using the rainfall infiltration breakthrough (RIB) and the water table fluctuation (WTF) techniques of assessing groundwater recharge. The pumping test technique is the most often used method for determining specific yields. However, due to restricted resources (lack of access to other wells in the area and inability to undertake pumping tests), the specific yield was taken from literature, limiting data availability.

CHAPTER FOUR

RESULTS AND DISCUSSIONS

4.1 Introduction

A thorough understanding of how to compute recharge value is essential for efficiently assessing groundwater availability. Natural groundwater recharge estimations produced utilizing the rainfall infiltration breakthrough (RIB) and the water table fluctuation (WTF) techniques are presented and discussed in this chapter. The goal is to validate the groundwater recharge estimations from the RIB model and the WTF approach. The claim is that groundwater recharge results acquired from the RIB model are comparable to estimates derived from the WTF technique.

4.2 Estimation of specific yield

To use the RIB and the WTF methods, an assessment of the specific yield at a depth of water table fluctuation is necessary. The association between rainfall and groundwater level rise for individual recharge events is demonstrated in Figure 4.1. Water level fluctuation data was acquired from one of the wells chosen for application with the rainfall infiltration breakthrough (RIB) and the water table fluctuation (WTF) techniques:

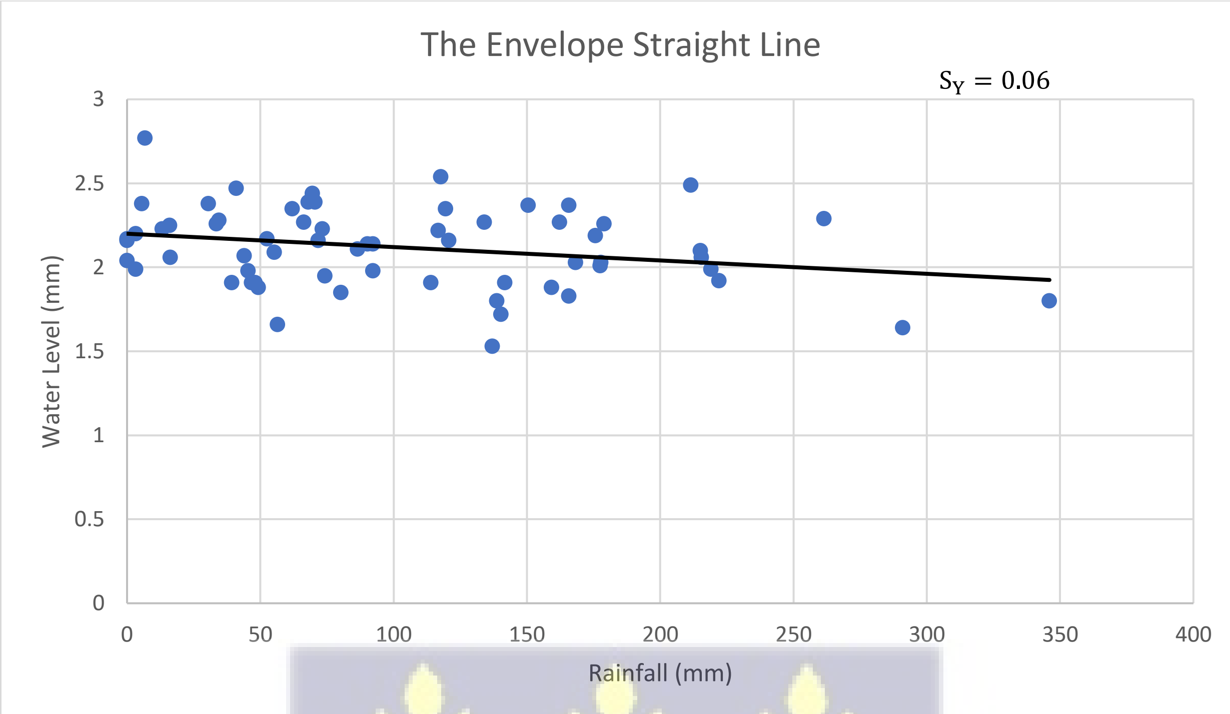


Figure 4.1: Graphical determination of specific yield using the envelope straight line of the precipitation–groundwater rises points of each recorded rainfall–recharge occurrence.

The relationship between rainfall and groundwater level increase for individual recharge events is indicated in Figure 4.1. The line drawn corresponds to Sy = 0.06. This value obtained from the linear regression method was in close range with the value (0.05) obtained from the literature. This number is the greatest value because if all the rainfall recharge events occurred when the soil and vadose zone profile was not at field capacity, this value could be less. Acquiring a specific yield that is typical of a massive region has proven challenging in previous research. Furthermore, unlike the assumption of a constant and known specific yield, specific yield values change with time (Delin et al., 2006; Loheide et al., 2005; Sophocleous, 1985). As there is no laboratory determined specific yield values for the study basin, the specific yield value was also selected from literature, based on the geologic material of the aquifers in the study area and guided by the range of specific yield value (0.01– 0.05) in Shahin (2002), for the weathered zone material in neighboring Burkina Faso. Ghana and Burkina Faso, to a large extent, have the same geology. The values 0.05 and 0.08 were selected from a recent study in the White Volta River Basin (Oboubie et al., 2012 after Sinha and Sharma 1988) to represent the high end of the estimation. This technique has been tested by Varni et al., (2013) using the 2-year-long series of data and Mutoti (2015) in his research to estimate groundwater recharge in the upper Berg River catchment, South Africa.

4.3 Recharge estimation using RIB model

The rainfall infiltration breakthrough model was chosen to estimate groundwater recharge based on hydrogeological data. This method is based on the water balance principle (the relationship between rainfall and groundwater level) and requires non-deterministic variables such as groundwater level observations, rainfall, and aquifer features (Sy).

The RIB model was applied in 5 Thiessen polygons created based on existing observation wells. The monthly recharge as a percentage of monthly rainfall and as well as lag time was calculated for each polygon. Figures 4.2, 4.3, 4.4, 4.5, and 4.6 illustrate the plotted graphs of rainfall inputs and computed groundwater recharge in mm as bars. On a monthly basis, the simulated groundwater level fluctuation derived using the RIB technique closely matches the measured values after calibration of the time lag (lag-month) and the length of associated rainfall events (length-month). The term lag-month was 0 months for boreholes NMW10S, NMW12S, and NMW17S, indicating that percolating rainwater reached the water table in less than a day. For boreholes MW8 and NMW8S, the term lag- month was 9. This indicates that it took less than 270/279 days for percolating rainwater to reach the water table. The term length months were 2, 2, 1, 2, 1 for boreholes MW8, NMW8S, NMW10S, NMW12S, and NMW17S monitored, respectively. The duration of the recharge event is prolonged with increasing thickness of the unsaturated zone.

Using the rainfall infiltration breakthrough method (Eq. 3.4), the predicted annual groundwater recharge was 2.92% in borehole MW8 with a specific yield of 0.05. In borehole NMW8S, the annual groundwater recharge was estimated to be 3.36%, with a specific yield of 0.05. Annual groundwater recharge was estimated to be 21.01% in borehole NMW10S, with a specific yield of 0.05. In borehole NMW12S, the predicted annual groundwater recharge was 7.32% with a specific yield of 0.05, while the predicted annual groundwater recharge in borehole NMW17S was 21.36% with a specific yield of 0.05.

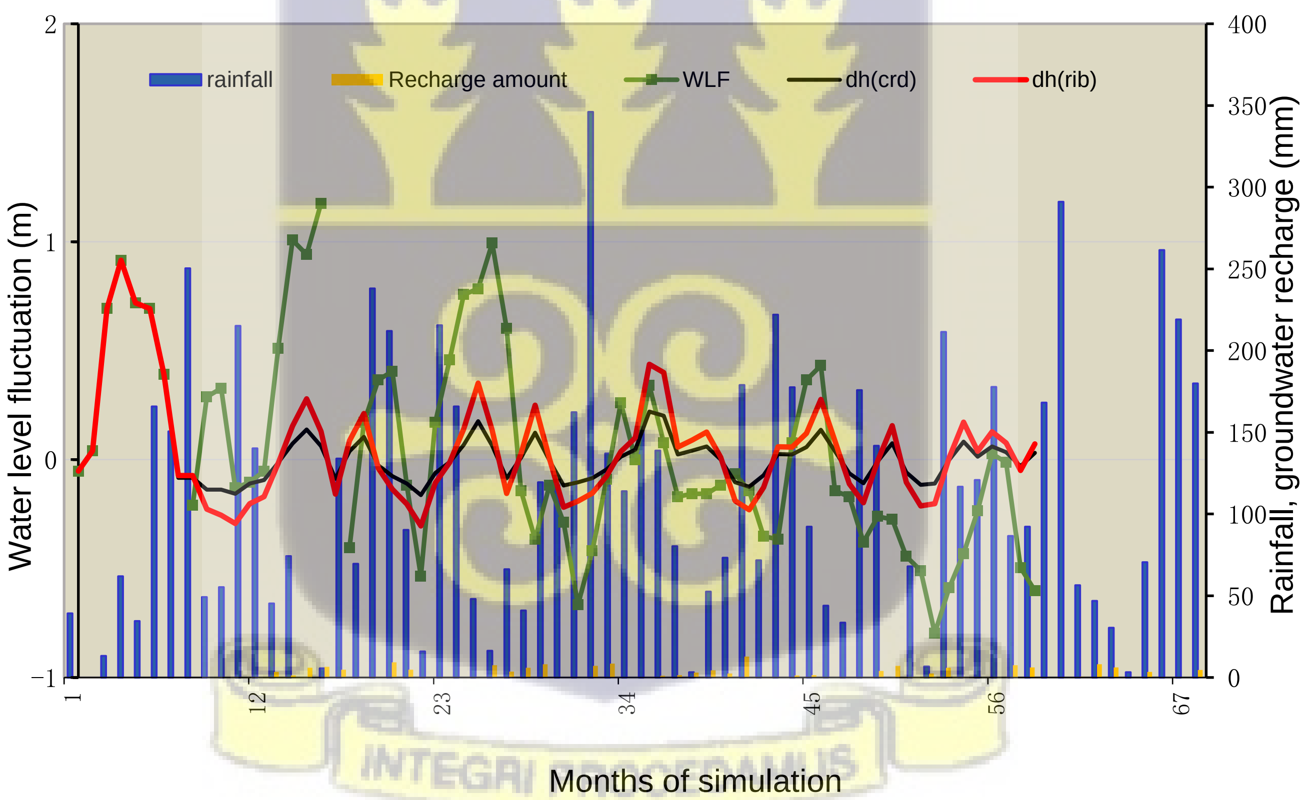


Figure 4.2: Monthly rainfall, observed WLF as well as calculated WLF and groundwater recharge in borehole MW8

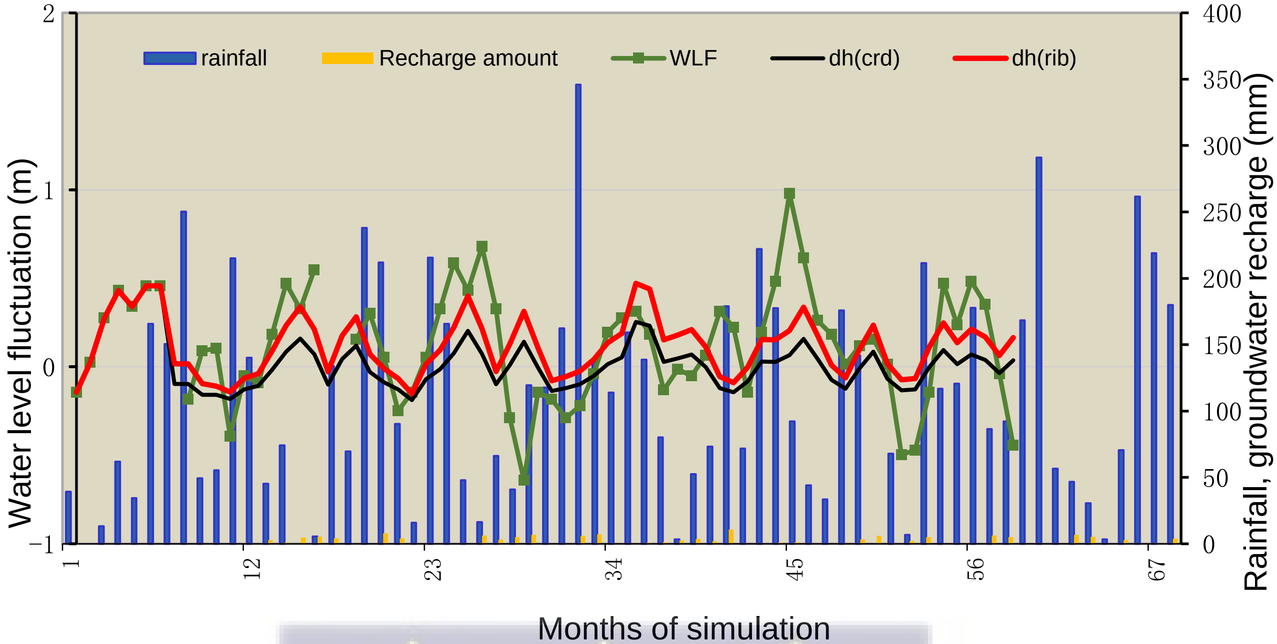


Figure 4.3: Monthly rainfall, observed WLF as well as calculated WLF and groundwater recharge in borehole MW8S

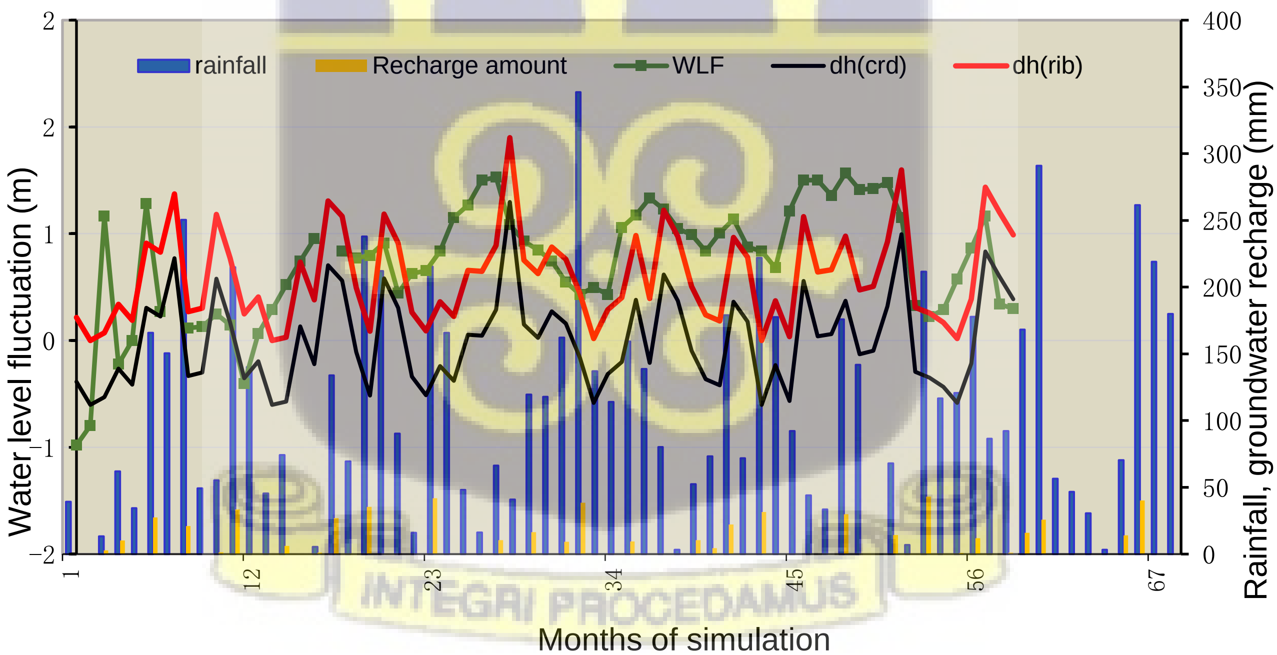


Figure 4.4: Monthly rainfall, observed WLF as well as calculated WLF and groundwater recharge in borehole NMW10S

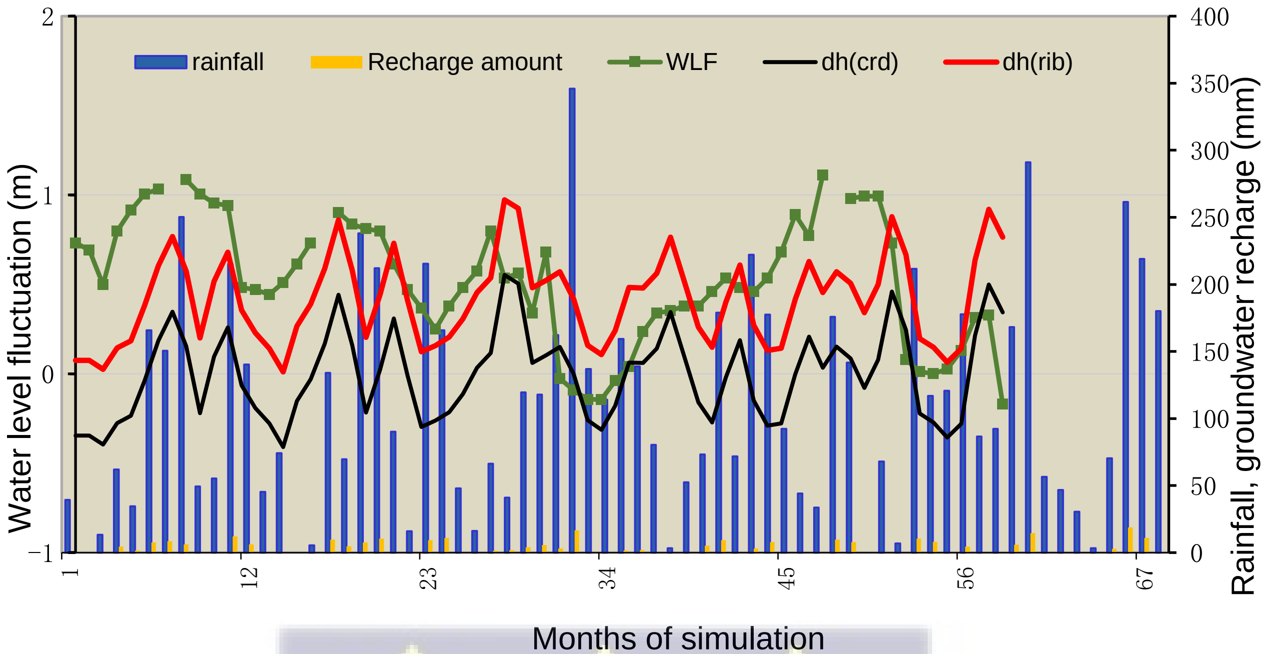


Figure 4.5: Monthly rainfall, observed WLF as well as calculated WLF and groundwater recharge in borehole NMW12S

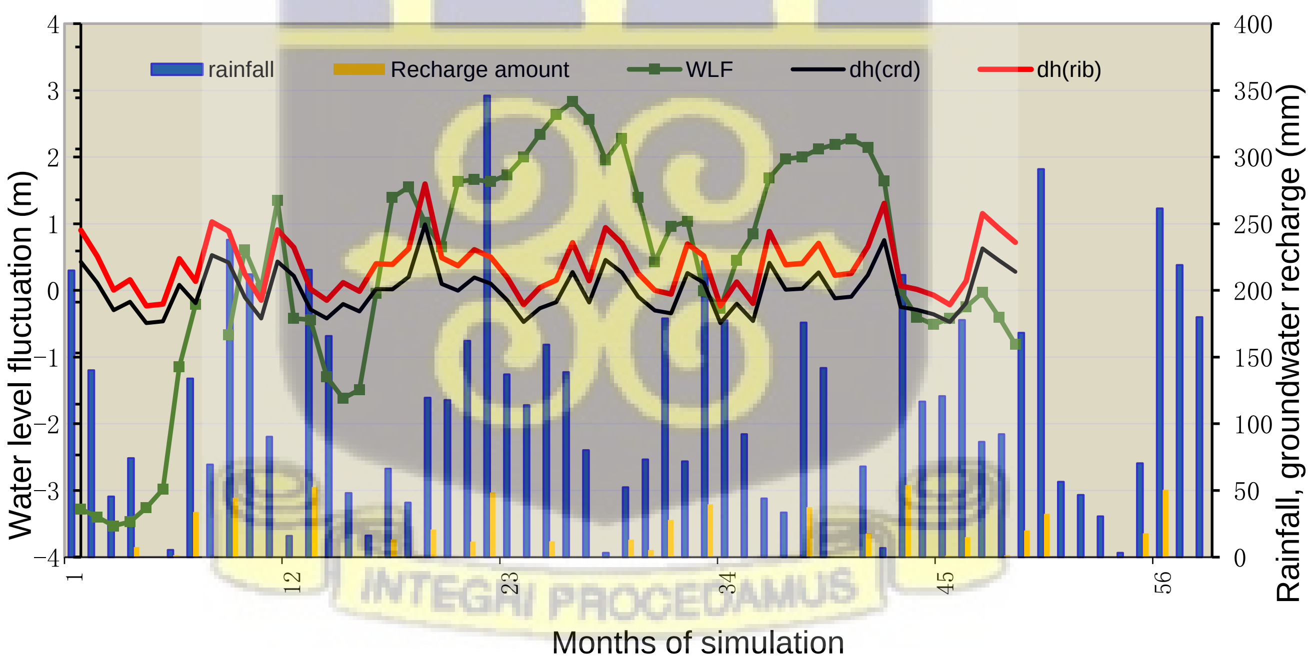


Figure 4.6: Monthly rainfall, observed WLF as well as calculated WLF and groundwater recharge in borehole NMW17S

The simulations were done using a monthly time series of rainfall and groundwater levels. The lengths of the simulation of monitored boreholes MW8, NMW8S, NMW10S, and NMW12S were 68. The length of the simulation was 58 months for borehole NMW17S. Simulation in borehole NMW17S began on 9/1/2012 and ended on 6/1/2017. The simulation started for boreholes MW8, NMW8S, NMW10S, and NMW12S from 11/1/2011 to 6/1/2017. There are some gaps that occurred in water levels in all boreholes. No water level data were available for the 6th month of 2012 and the 5th month of 2013 for all boreholes. There were 2 months water level data gap that occurred from 5/1/2013 to 6/1/2013 in borehole NMW8S. There was also no water level data for the 6th month of 2016 for borehole NMW12S.

4.3.1 Results from rainfall infiltration breakthrough model

Table 4.1 shows the results for the lag time and length of related rainfall events on a monthly basis with a specific yield of 0.05.

The RIB model was applied in five existing observation wells. For each of the five observation wells, the monthly groundwater recharge was calculated as a percentage of the monthly rainfall. Using the RIB model and a specific yield of 0.05, monthly groundwater recharge was estimated, ranging between 2.92 % and 21.36% of the mean monthly rainfall as shown below (Table 4.1), which implies that less than a third of the cumulative rainfall acts to recharge groundwater.

The results of the RIB model indicated that the length between rainfall infiltration and groundwater recharge for most of the boreholes have different lag and length times, which may reflect the different characteristics of hydrogeological systems in the study area. The MW8 and NMW8S polygons in the aquifer outlet had the largest lag time (9 months). This indicates that rainfall that occurred over the course of two months reaches the water table after nine months. As expected, increasing the depth to groundwater increases the lag times. The delayed time increases as the depth to the groundwater table increases, and percolation in the unsaturated zone reduces due to evaporation from the unsaturated zone, and so recharge to the saturated zone decreases (Ahmadi et al., 2014; Majdabady et al., 2020). Also, the relatively low hydraulic conductivity of some geologic materials limits permeability due to the availability of clay materials, increasing the lag times as the duration of the recharge event is prolonged. These were found to be in good agreement with the physiological conditions of each sub-zone, depending on the depth of the water table. Other than rainfall, atmospheric pressure, the impacts of evapotranspiration, and entrapped air on short-term water level fluctuations can be considerable (Healy, 2010).

The recharge estimates acquired using the RIB model in this study are in the same range as estimates from groundwater studies conducted in similar geology. A study conducted by Atta-Darkwa et al., (2013) quantified groundwater recharge in the river Oda catchment. The recharge estimates for the research site varied from 133-467 mm for the 14 wells placed for the work, which represented 9- 31 percent of the 2009 annual rainfall and 47.6-427.9 mm, in 2010 indicating 4-34 percent of the annual rainfall. Similarly, Sanwidi (2007) applied comparable technique for the Kompienga Dam Basin in Burkina Faso close to Ghana and predicted the recharge to be from 5.3-29.4 percent of the annual rainfall.

Based on these similar results, it can be stated that the RIB approach can predict recharge if the specific yield is available and requirements specified in the preceding chapters are achieved.

Table 4.1: Groundwater recharge estimates from RIB model

	Time lag	Length of rainfall events		
Borehole	Monthly scale (month)	Monthly scale (month)	Specific yield	Groundwater recharge (monthly)
MW8	-9	2	0.05	2.92%
NMW8S	-9	2	0.05	3.36%
NMW10S	0	1	0.05	21.01%
NMW12S	0	2	0.05	7.32%
NMW17S	0	1	0.05	21.36%

The Spearman correlation coefficients in Table 4.2 show that there is a significant association between monthly rainfall and observed water level data. As a result, recharge rates computed on a monthly basis are quite close to reality. This is consistent with groundwater flow mechanisms, which state that precipitation accumulated in the unsaturated zone is moved downwards by the next infiltration/percolation event without disrupting moisture distribution and preferential flow in the unsaturated zone with relatively high infiltration and/or percolation capacity by the next infiltration/percolation event (Sun et al., 2005). Water-level data can be corrected using a variety of methods (Crosbie et al., 2005; Tamura et al., 1991; Toll and Rasumssen, 2007; Von Asmuth et al., 2008).

Water level fluctuations occur primarily on a monthly basis as a result of rainfall, pumping, evapotranspiration, or other phenomena. The simulated water level computed using the RIB approach closely matches the observed values after calibration, indicating the reliability of the RIB model.

Table 4.2: The Spearman correlation coefficients between rainfall and observed WLF Borehole Monthly scale

Borehole	Monthly scale
MW8	-0.4
NMW8S	-0.5
NMW10S	0.5
NMW12S	0.2
NMW17S	0.5

The RIB model is most suitable for low-transmissivity aquifers, according to Sun et al., (2013). The Rainfall infiltration breakthrough is a water balance approach that depends on examining groundwater level fluctuations in shallow unconfined aquifers due to rainfall (Ahmadi et al., 2014). As a result, in order to assess groundwater recharge in the research area, the rainfall infiltration breakthrough model was used. When used in places where groundwater levels respond clearly to rainfall, this model can provide accurate predictions of groundwater recharge.

4.4 Sensitive Analysis

A sensitivity analysis for water level and rainfall was also undertaken at the five borehole sites – MW8, NMW8S, NMW10S, NMW12S, and NMW17S (Sy = 0.05). A constant distribution was applied with perturbations of ±10% around the observed water level and rainfall values; therefore, the two extra-time series show the uniform distribution's extremes/boundaries (maximum and minimum). The model was run on the smallest and largest time series. The simulation results offered a range of recharge variability, which defined the sensitivity to the essential model inputs. Table 4.3 displays the monthly sensitivity analysis results of groundwater recharge assessed with the RIB model.

The results in Table 4.3 reveal that a perturbation of the water level time series data by a factor ±10% leads in a range of recharge of 2.61-3.22% of MAP in MW8, 3.04-3.71% of MAP in borehole NMW8S, 18.92-21.93% of MAP in borehole NMW10S, 6.59-8.07% of MAP in borehole NMW12S and 17.49-21.36% of MAP in borehole NMW17S. The results also show that when the rainfall time series data is perturbed by a factor of ±10%, the recharge percentage ranges from 2.65-3.42% of MAP in MW8, 3.06-3.73% of MAP in borehole NMW8S, 19.09-23.34% of MAP in borehole NMW10S, 6.66-8.14% of MAP in borehole NMW12S and 19.42-23.74% of MAP in borehole NMW17S. The sensitivity of recharge calculated using the RIB model is higher for boreholes NMW17S and NMW10S than for boreholes MW8, NMW8S, and NMW12S, based on these analyses.

Table 4.3: Summary of inputs of the sensitivity analysis of groundwater recharge calculated with the RIB model on a monthly basis

Borehole	Mean monthly recharge range (mm)	% of MAP
	Outputs of recharge based on water level fluctuations (±10% perturbation)	
MW8	2.36-2.89	2.61-3.22
NMW8S	1.99-2.43	3.04-3.71
NMW10S	7.39-8.57	18.92-21.93
NMW12S	3.37-4.14	6.59-8.07
NMW17S	9.01-11.01	17.49-21.36
	Outputs of recharge based on rainfall input (±10% perturbation)	
MW8	2.62	2.65-3.42
NMW8S	2.20	3.06-3.73
NMW10S	8.21	19.09-23.34
NMW12S	3.75	6.66-8.14
NMW17S	10.01	19.42-23.74

4.5 Climate scenarios

The RIB model was used to run four simulations to forecast groundwater level and recharge changes under the condition of reduced rainfall. The following scenarios applied at borehole NMW8S were considered:

- Baseline conditions: Using recorded rainfall data from November 2011 to June 2017; Sy =0.05; aquifer area = 14.80 km^2
- Reduction in recorded rainfall by 10%; Sy =0.05; aquifer area = 14.80 km^2
- Reduction in recorded rainfall by 20%; Sy =0.05; aquifer area = 14.80 km^2
- Reduction in recorded rainfall by 30%; Sy =0.05; aquifer area = 14.80 km^2

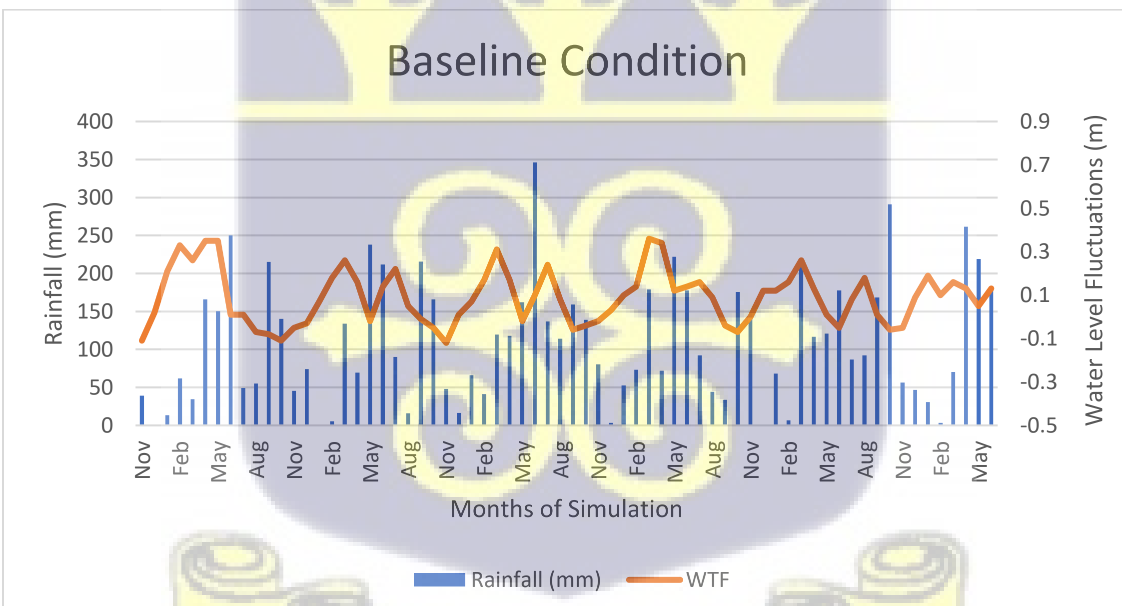


Figure 4.7: Baseline conditions of measured rainfall and groundwater level fluctuations for borehole NMW8S

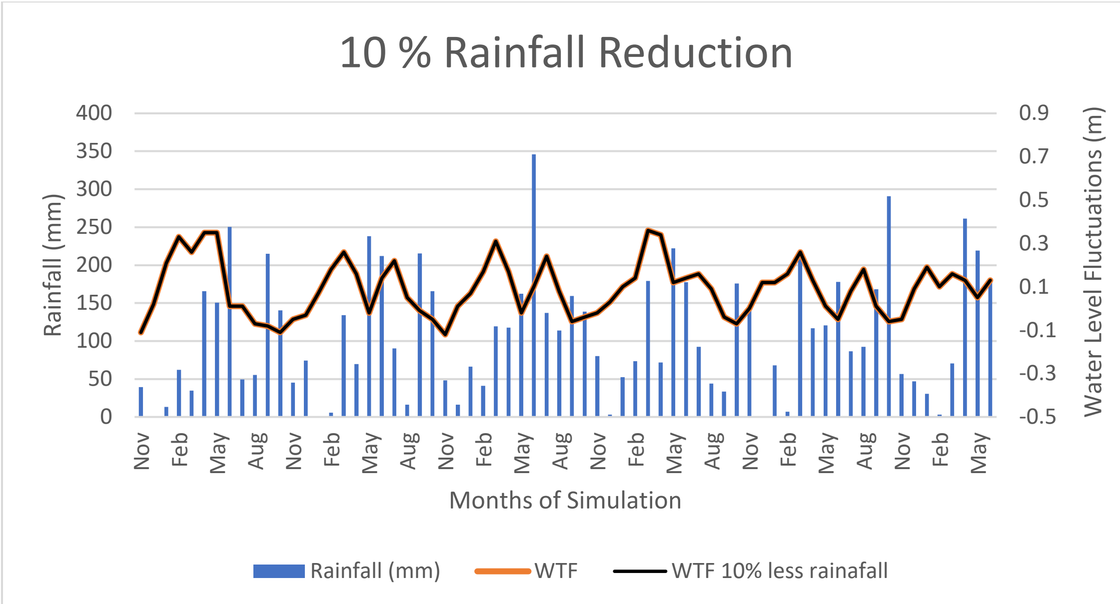


Figure 4.8: Measured rainfall and groundwater level fluctuations with 10 % less rainfall for borehole NMW8S

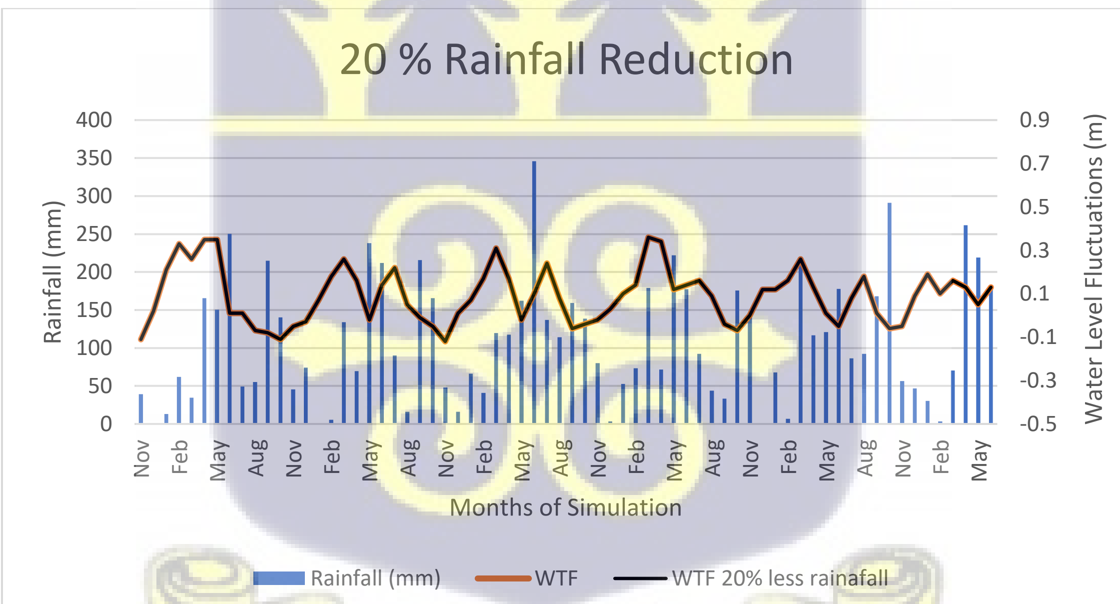


Figure 4.9: Measured rainfall and groundwater level fluctuations with 20 % less rainfall for borehole NMW8S

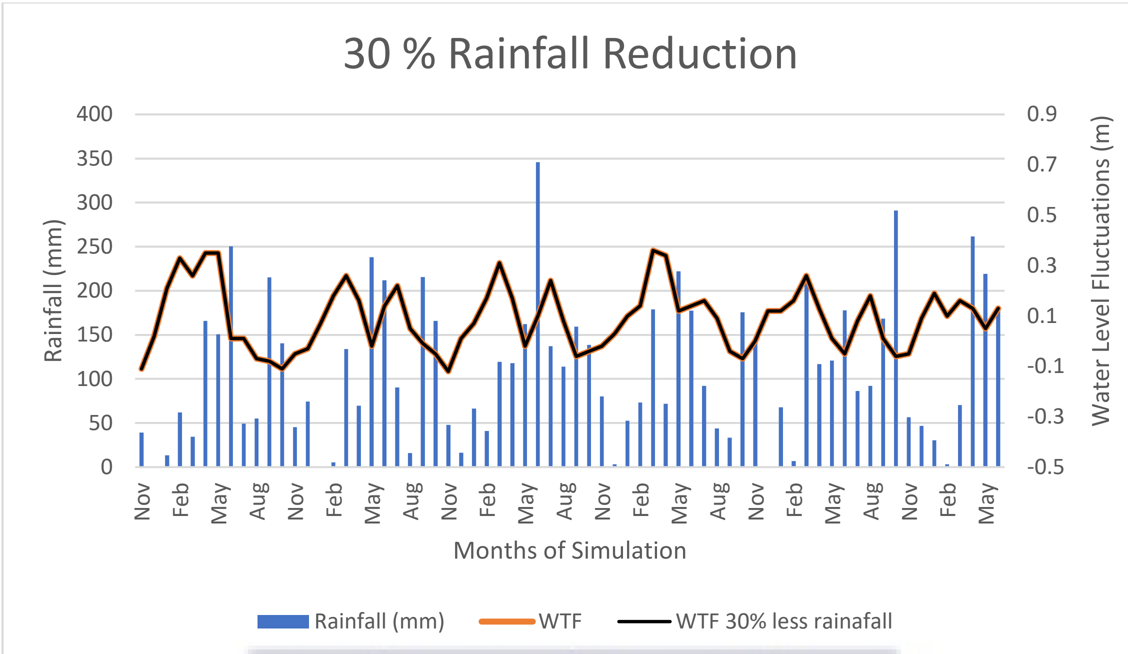


Figure 4.10: Measured rainfall and groundwater level fluctuations with 30 % less rainfall for borehole NMW8S

The projections of fluctuating groundwater levels are presented in Figure 4.7-4.10; the mean recharge percentage for the baseline condition is about 3.36 percent from 2011 to 2017. Reduced rainfall had little effect on groundwater levels over the simulated period, according to the prediction. The ratio of recharge rate to precipitation, on the other hand, did not alter considerably; it was somewhat greater than the baseline (mean recharge percentage is 3.73%, 4.20% and 4.80% of corresponding rainfall with 10%, 20%, and 30% rainfall reduction, respectively). It can therefore be concluded that a decrease in rainfall has no effect on groundwater levels in this model but increases the recharge amount. One limitation of the RIB model is that it ignores additional components like evapotranspiration and runoff. Changes in groundwater levels may not always be caused by recharge or discharge. It may also result from other factors, such as variations in atmospheric pressure, the presence of trapped air, earth tides, the combination of evaporation and transpiration, or it may be a reaction to stream stage changes for boreholes near streams (Delin et al., 2006). That explains why the groundwater levels remained unaffected even with reduced rainfall in this scenario analysis.

The RIB model is theoretically unsuitable for recharge estimation in highly arid places when rainfall (P) is less than 100 mm/a because rainfall recharge is too small to induce a rise of water level, i.e., $\Gamma \infty 0$. The aquifer's hydraulic conductivity (K) cannot be excessively high. When the hydraulic conductivity of an aquifer is high, for example, the outflow from the aquifer exceeds the inflow. In other words, even with a very high K value, recharge can still occur without causing a rise in the water table. As a result, applying the Rainfall infiltration breakthrough (RIB) method to aquifers of fractured rock should be approached cautiously, as some fractures may serve as a principal conduit for water transport; and recharge from rainfall may move into the aquifer without causing water level fluctuations. The water table depth (D) should be shallow in terms of time lag. The time lag increases as the thickness of the unsaturated zone increases. Individual rainfall events, for example, could barely play a substantial effect in water level changes for water table depths of 200 m in northern African countries. In such instances, the parameter Γ approaches 0.

4.6 Recharge estimation using water table fluctuation (WTF) method

The graphs below (Figure 4.11 to 4.15) show the variation in groundwater levels in response to rainfall from all the wells that were analyzed between certain time periods.

Figure 4.11 shows the water level in response to rainfall from the MW8 borehole between the period of 2012-2017.

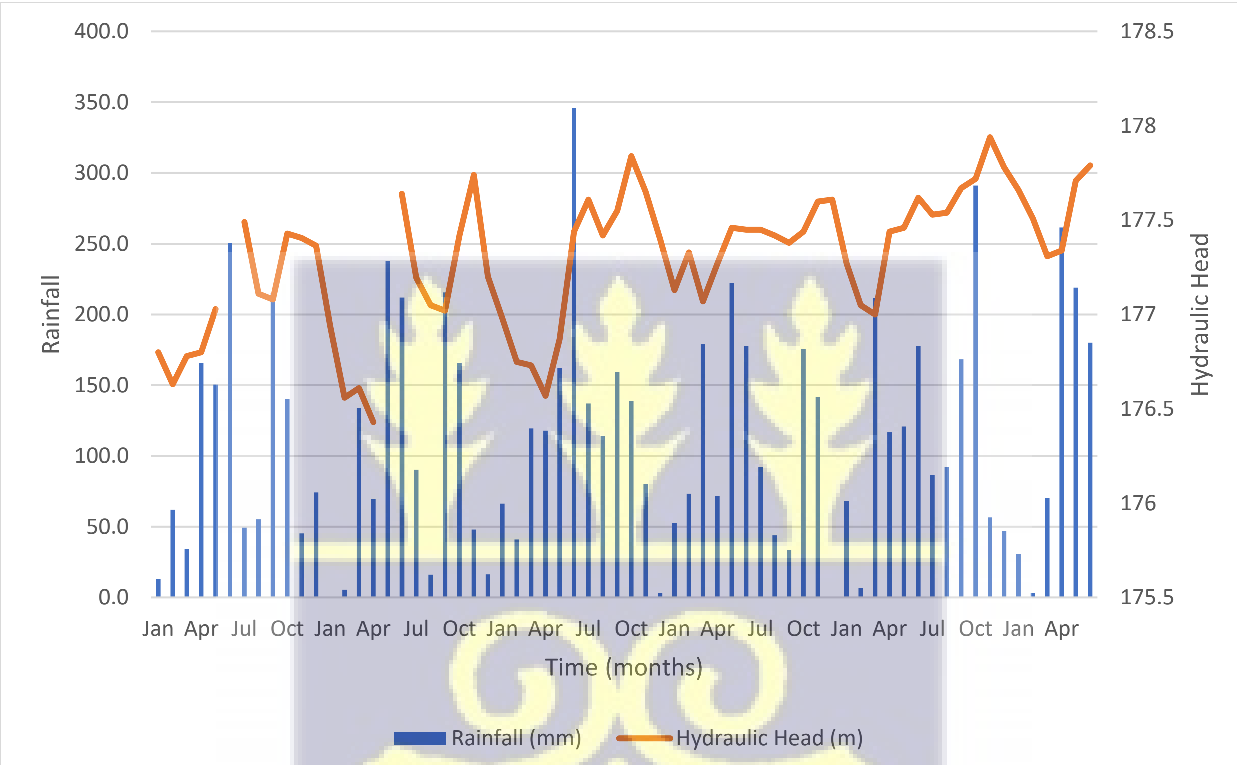


Figure 4.11: Water table fluctuation in response to rainfall in MW8 from 2012 to 2017

The graph in Figure 4.11 illustrates that the water levels only rise when there has been a rainfall event. The graph shows that the area receives a lot of rain during the major rainy season (late March to July) and less rain during the minor rainy season (September to November). The maximum rainfall, 346 mm, was recorded in June of 2014, and during that time, the water level in borehole MW8 climbed by 0.58 m. This rise implies that a significant amount of rainfall has been recharged into the groundwater system. The graph also shows that there was less rainfall in 2013, with the peak rainfall of 238 mm occurring in June 2013, resulting in low water levels during that time.

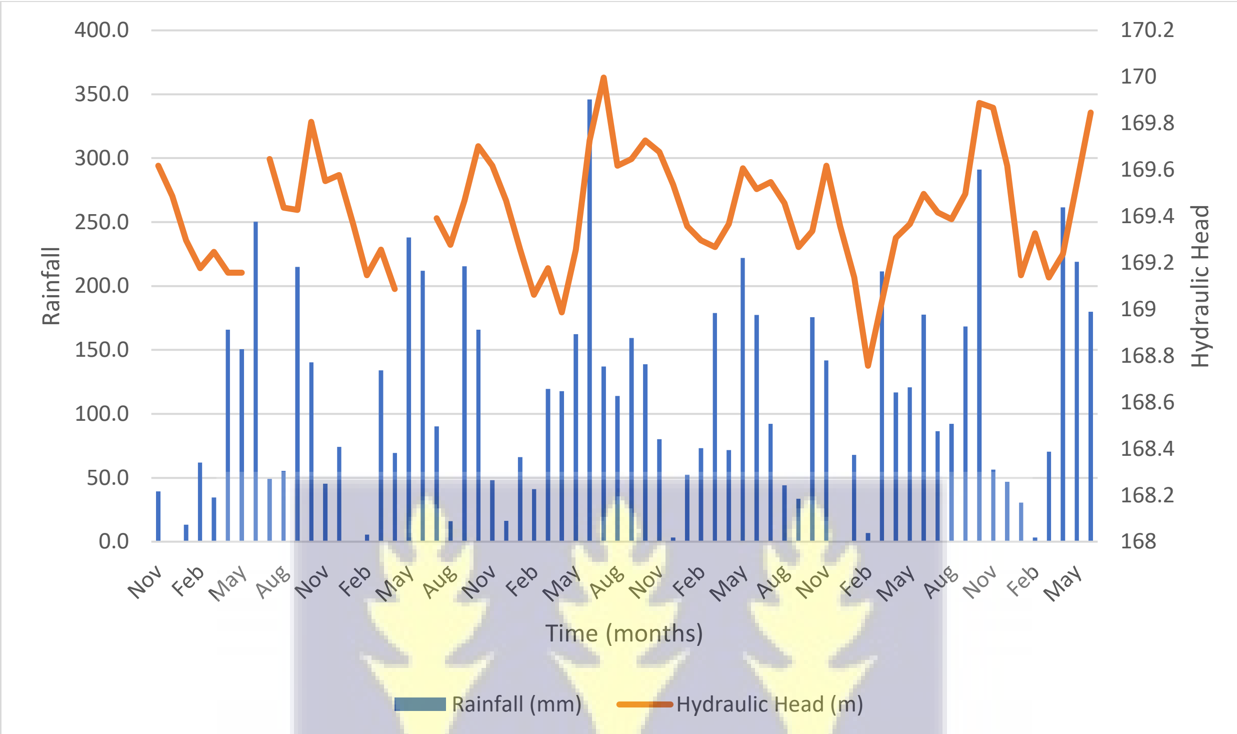


Figure 4.12: Water table fluctuation in response to rainfall in NMW8S from 2012 to 2017

The graph in Figure 4.12 illustrates that the only time that water levels rise is when there has been a rainfall event. The graph shows that the area receives much rain during the major rainy season (late March to July) and less rain during the minor rainy season (September to November). The most rainfall was recorded in June of 2014, with a volume of 346 mm, and the water level in borehole NMW8S climbed by 0.47 m during that occurrence. This rise implies that a significant portion of the rainfall has been recharged to the groundwater. The graph also shows that less rainfall fell in 2013, with the maximum total of 238 mm falling in June 2013, resulting in low water levels during that time period.

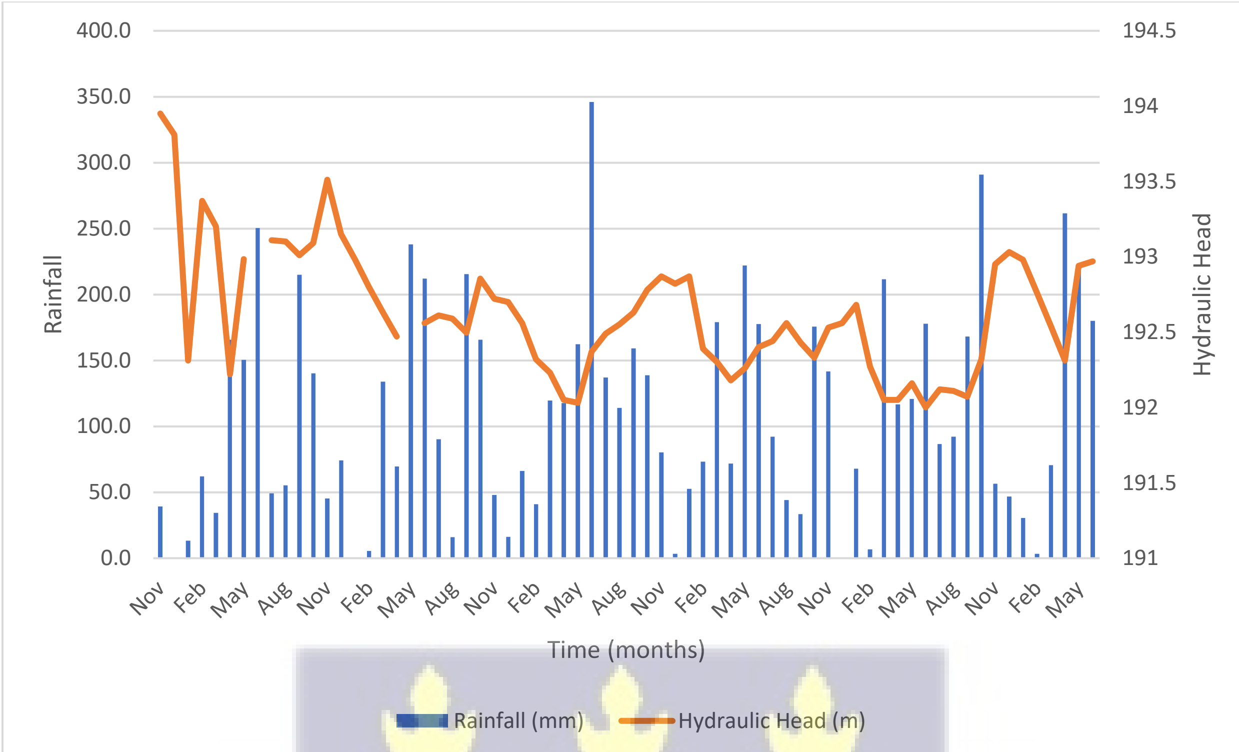


Figure 4.13: Water table fluctuation in response to rainfall in NMW10S from 2012 to 2017

Figure 4.13 demonstrates that the only time the water levels rise is when there has been a rainfall event. The graph shows that the area receives a lot of rain during the main rainy season (late March to July) and less rain during the minor rainy season (September to November). The most rainfall, 346 mm, was recorded in June of 2014, and the water level in borehole NMW10S grew by 0.34 m during that event. This rise implies that a large portion of the rainwater has been recharged into the groundwater system. The graph also shows that rainfall was lower in 2013, with the maximum total of 238 mm falling in June of that year, resulting in low water levels.

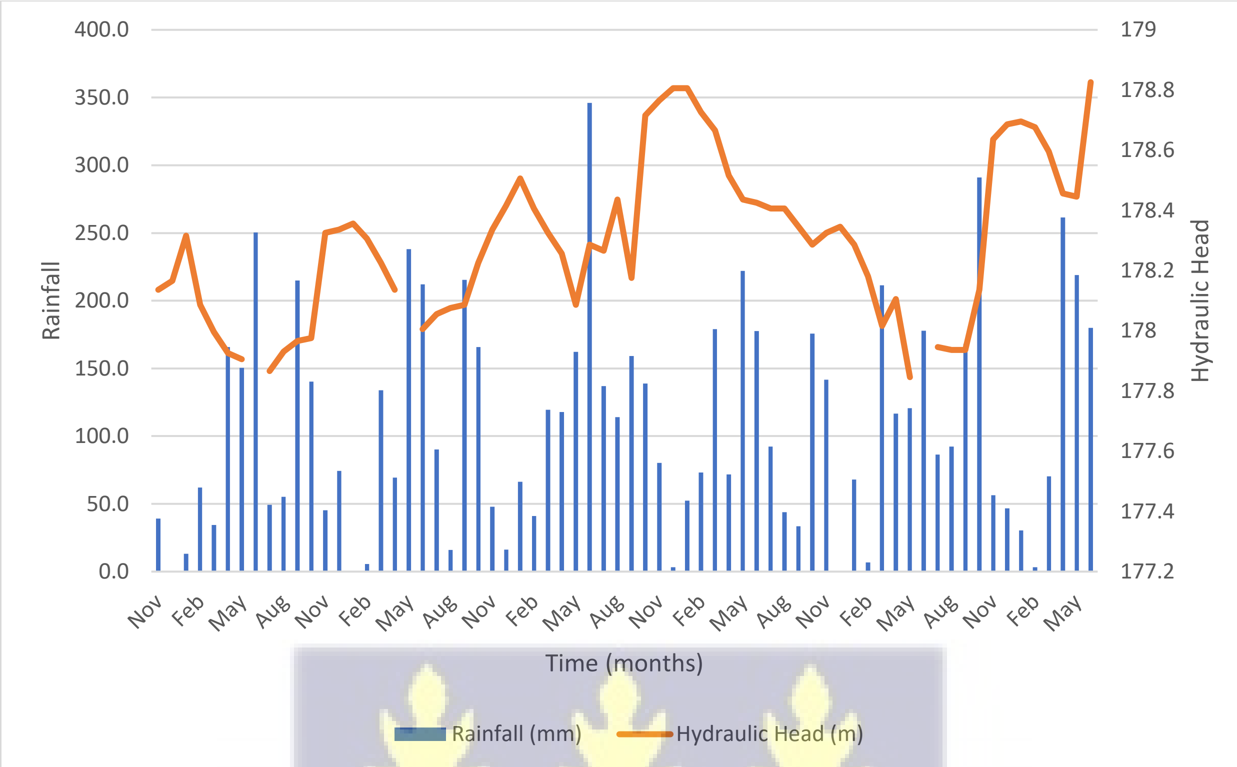


Figure 4.14: Water table fluctuation in response to rainfall in NMW12S from 2012 to 2017

The only time the water level elevates is when there has been a rainfall event, as shown in Figure 4.14. The graph indicates that the area receives a lot of rain during the major rainy season (late March to July) and less rain during the minor rainy season (September to November). The most rain fell in June of 2014, totaling 346 mm, and the water level in borehole NMW12S grew by 0.20 m during that time. This rise suggests that a significant portion of the rainfall has been recharged into the groundwater. The graph also shows that there was less rainfall in 2013, with the peak rainfall of 238 mm occurring in June 2013, resulting in low water levels during that time.

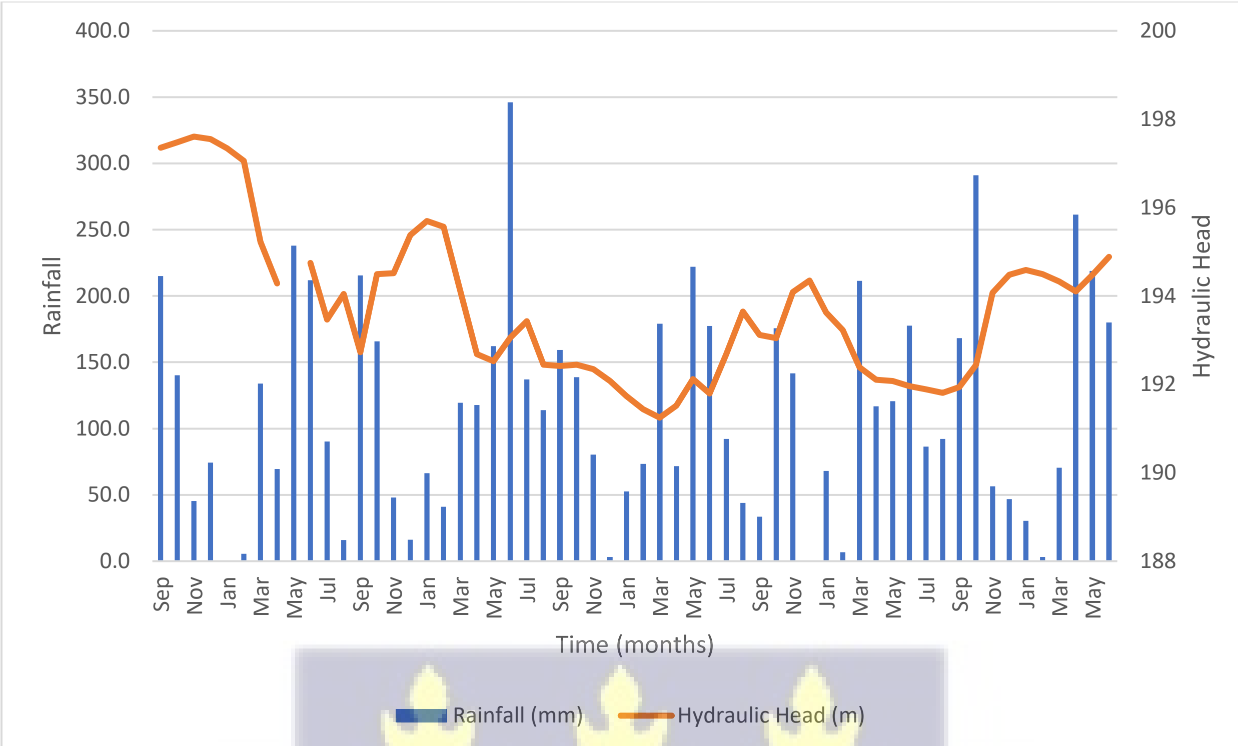


Figure 4.15: Water table fluctuation in response to rainfall in NMW17S from 2012 to 2017

The graph in Figure 4.15 illustrates that the water levels only rise when there has been a rainfall event. The graph illustrates that the area receives a lot of rain during the major rainy season (late March to July) and less rain during the minor rainy season (September to November). The maximum rainfall, 346 mm, was recorded in June of 2014, and during that time, the water levels in borehole NMW17S climbed by 0.53 m. This rise implies that a significant amount of rainfall has been recharged into the groundwater system. The graph also shows that rainfall was lower in 2013, with a peak of 238 mm in June 2013, resulting in low water levels during that period.

Within the research area, the extent to which levels of water fluctuated in the boreholes varied greatly. The position of the boreholes was largely responsible for the diversity in water level rise observed by these boreholes. Because the rainy season in the research area began in April, groundwater levels in all boreholes did not begin to climb until June and July, when approximately forty percent of the annual rainfall had fallen. The two to four months between the beginning of the rainy season and the start of the rise in groundwater level could be thought of as a period of soil replenishment caused by moisture shortages from the previous dry season. In the study area, the lag indicates that there are impacts of threshold and a non-linear correlation between recharge and rainfall. Furthermore, the lag indicates that most of the boreholes in the studied area do not recharge quickly. In the Atankwidi basins, Martin (2005) observed comparable results. Sandwidi (2007) also discovered similar findings at the Kompienga dam. Because the rise in water level occurred primarily during the rainy season, groundwater recharge is virtually completely obtained from seasonal rainfall, according to a careful analysis of well hydrographs and groundwater level data for the six research periods. During the dry season, there was a little amount of recharge, probably due to regional groundwater flow. As a result, it is acceptable to conclude that aquifers outside the study area have little impact on the recharge of groundwater in the research region.

4.6.1 Annual Recharge

Using the Sy values of 0.05 (or 5%) indicated in the previous section, the following is an estimate of groundwater recharge. The annual recharge can be calculated by adding the annual cumulative individual rises in groundwater level. The mean annual rainfall of borehole MW8, NMW8S, NMW10S, NMW12S, and NMW17S, for the period, were 1652.9 mm, 1066 mm, 1066 mm, 1066 mm, and 1106.9 mm, respectively (Table 4.4, 4.5, 4.6, 4.7, and 4.8).

In borehole MW8, the most annual recharge happened with 66 mm in 2013, which was 5.4% of the annual rainfall, despite having one of the lowest rainfalls that year (1210.8). The least recharge happened in 2015 with 27 mm, or 1.3 percent of the 2128 mm of precipitation, despite having a higher amount of rainfall that year than in past years (Table 4.4). The mean annual recharge for the 6-year period was 45 mm, or 2.9% of precipitation (Sy = 0.05). Recharge in borehole MW8 is generally low despite having higher precipitation values. In reality, water level changes are caused by a variety of factors other than rainfall, such as baseflow, inflow, and outflow of the aquifer, and abstraction of groundwater, among others (Sun et al., 2013).

Table 4.4: Annual precipitation (mm), annual level variation (m), net annual level rise (used to calculate recharge) (m), net annual level decline (m), and recharges (%P) for Sy = 0.05 for borehole NW8

Year	**P (m)**	***Δh (m)***	**Net rise**	**Net decline (m)**	**Sy**	**R**	**%**
2012	1256	176.63	177.49	0.86	0.05	0.043	3.4
2013	1211	176.43	177.74	1.31	0.05	0.0655	5.4
2014	2127	176.57	177.84	1.27	0.05	0.0635	3.0
2015	2129	177.07	177.61	0.54	0.05	0.027	1.3
2016	2130	177.00	177.94	0.94	0.05	0.047	2.2
2017	1065	177.31	177.79	0.48	0.05	0.024	2.3
Mean	1653	176.83	177.73	0.90	0.05	0.04	2.9

The percentage recharge and precipitation values in borehole NMW8S are shown in Table 4.5. The greatest annual recharge at borehole NMW8S was 56.5 mm, or 3.9% of annual rainfall, in 2016, though the maximum percentage recharge period was in 2017, with 4.6% of rainfall (35.5 mm of 765 mm of P). In 2015, the least recharge was 18 mm, or 1.4% of 1263 mm of precipitation, which was also the lowest in the era. For the 6-year period, the average annual recharge was 37 mm, or 3% of precipitation (Sy = 0.05).

Table 4.5: Annual precipitation (mm), annual level variation (m), net annual level rise (used to calculate recharge) (m), net annual level decline (m), and recharges (%P) for Sy = 0.05 for borehole NMW8S

Year	**P (m)**	***Δh (m)***	**Net rise**	**Net decline (m)**	**Sy**	**R**	**%**
2012	1256	169.16	169.81	0.65	0.05	0.0325	2.6
2013	1211	169.09	169.70	0.61	0.05	0.03075	2.5
2014	1485	168.99	170.00	1.01	0.05	0.0505	3.4
2015	1263	169.27	169.62	0.35	0.05	0.0175	1.4
2016	1443	168.76	169.89	1.13	0.05	0.0565	3.9
2017	765	169.14	169.85	0.71	0.05	0.0355	4.6
Mean	1237	169.07	169.81	0.74	0.05	0.04	3.1

The maximum percentage recharge period in borehole NMW10S (Table 4.6) occurred in 2012 when 5.1% of rainfall was received (65 mm of 1256 mm of P). The lowest recharge occurred in 2013, with 26 mm, or 2.1 percent of the total precipitation of 1211 mm. This year has been one of the driest in the region. For the 6-year period, the mean annual recharge was 42 mm, or 3.5 percent of precipitation (Sy = 0.05).

Table 4.6: Annual precipitation (mm), annual level variation (m), net annual level rise (used to calculate recharge) (m), net annual level decline (m), and recharges (%P) for Sy = 0.05 for borehole NMW10S

Year	**P (m)**	***Δh (m)***	**Net rise**	**Net decline (m)**	**Sy**	**R**	**%**
2012	1256	192.22	193.51	1.29	0.05	0.0645	5.1
2013	1211	192.47	192.98	0.51	0.05	0.0255	2.1
2014	1485	192.03	192.87	0.84	0.05	0.042	2.8
2015	1263	192.18	192.87	0.69	0.05	0.0345	2.7
2016	1443	192.00	193.03	1.03	0.05	0.0515	3.6
2017	765	192.31	192.98	0.67	0.05	0.0335	4.4
Mean	1237	192.20	193.04	0.74	0.05	0.04	3.5

The greatest annual recharge at borehole NMW12S (Table 4.7) was 36 mm, or 2.4% of annual rainfall, in 2014, though the maximum percentage recharge period was in 2016, with 2.9% of rainfall (42 mm of 1443 mm of Precipitation). In 2013, the least recharge was 19 mm, or 1.7 percent of the total precipitation of 1211 mm. This has been one of the region's driest years. For the 6-year period, the mean annual recharge was 28 mm, or 2.2% of precipitation (Sy = 0.05).

Table 4.7: Annual precipitation (mm), annual level variation (m), net annual level rise (used to calculate recharge) (m), net annual level decline (m), and recharges (%P) for Sy = 0.05 for borehole NMW12S

Year	**P (m)**	***Δh (m)***	**Net rise**	**Net decline (m)**	**Sy**	**R**	**%**
2012	1256	177.87	178.34	0.47	0.05	0.0235	1.9
2013	1211	178.01	178.42	0.41	0.05	0.0205	1.7
2014	1485	178.09	178.81	0.72	0.05	0.036	2.4
2015	1263	178.29	178.81	0.52	0.05	0.026	2.1
2016	1443	177.85	178.69	0.84	0.05	0.042	2.9
2017	765	178.45	178.83	0.38	0.05	0.019	2.5
Mean	1237	178.09	178.65	0.48	0.05	0.02	2.2

The maximum percentage recharge period in borehole NMW17S happened in 2013 when 19.0% of rainfall fell (231 mm of 1211 mm of Precipitation). The least recharge happened in 2012, with 13 mm of precipitation, or 2.6 percent of 475 mm, which was also the smallest in the era (Table 4.8). The mean annual recharge throughout the 6-year period was 1107 mm, or 10.1% of precipitation (Sy = 0.05), indicating substantial annual fluctuation in comparison to the precipitation that produced it. The main component of the recharge of groundwater in this borehole is rainfall.

Table 4.8: Annual precipitation (mm), annual level variation (m), net annual level rise (used to calculate recharge) (m), net annual level decline (m), and recharges (%P) for Sy = 0.05 fore borehole NMW17S

Year	**P (m)**	***Δh (m)***	**Net rise**	**Net decline (m)**	**Sy**	**R**	**%**
2012	475	197.36	197.61	0.25	0.05	0.0125	2.6
2013	1211	192.73	197.34	4.61	0.05	0.2305	19.0
2014	1485	192.08	195.70	3.62	0.05	0.181	12.2
2015	1263	191.25	194.35	3.10	0.05	0.155	12.3
2016	1443	191.81	194.48	2.67	0.05	0.1335	9.3
2017	765	194.11	194.89	0.78	0.05	0.039	5.1
Mean	1107	193.22	195.73	2.51	0.05	0.13	10.1

The variation in recharge outcomes across the six research periods can be credited to annual rainfall distribution and intensity fluctuation. Due to variations in soils, land use/land cover, climate, and physiography, recharge varies spatially within the study area. The recharge estimate acquired in this work is consistent with outcomes obtained from water table fluctuation investigations conducted around the world. Martin (2006) used this technique in Ghana's Atankwidi watershed, estimating recharge to be between 1.8 and 12.5 percent of annual rainfall in 2003 and 1.4 to 10.3 percent in 2004. Later, Obuobie (2008) used a similar approach to estimate recharge in Ghana's White Volta Basin, which ranged from 28.0-150.0 mm in 2006, providing 3.5-16.5 percent of mean annual rainfall, to 32.0-204.0 mm in 2007, indicating 2.5-16.0 percent of mean annual rainfall, with a specific yield range of 0.01-0.05. It can be deduced that considerable relative variances in estimated recharge are caused by changes in the estimate of specific yield.

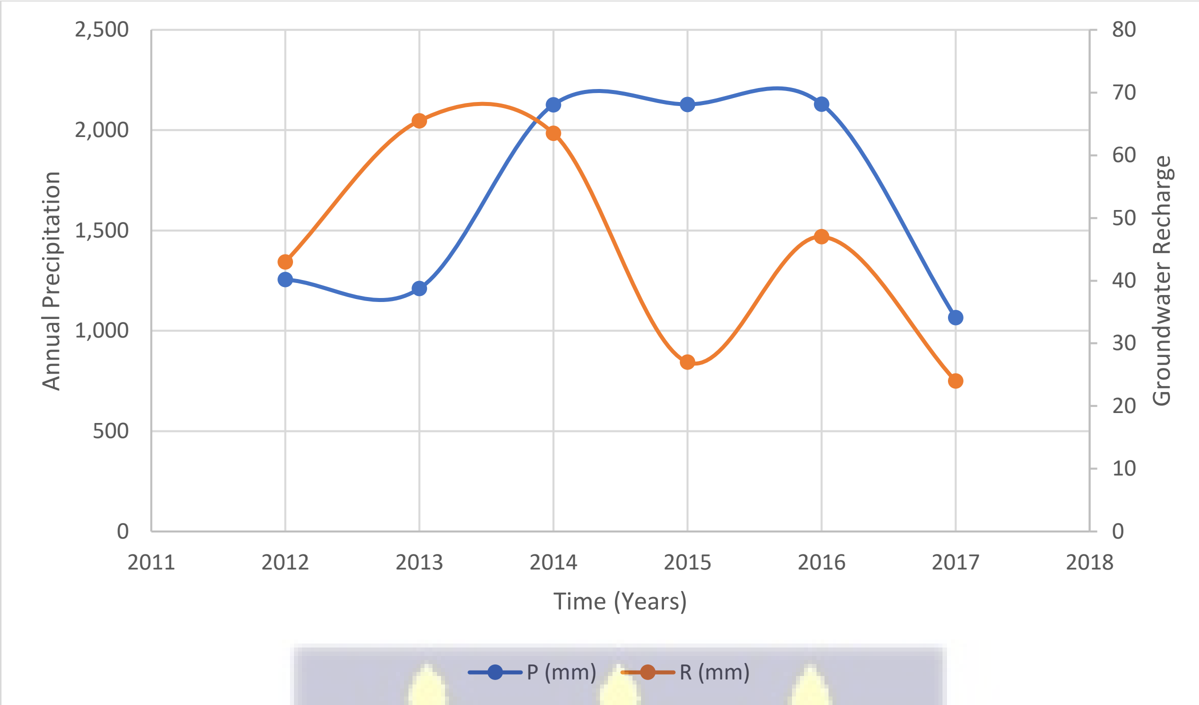


Figure 4.16: Annual variations in precipitation and groundwater recharge in borehole MW8 (2012–2017)

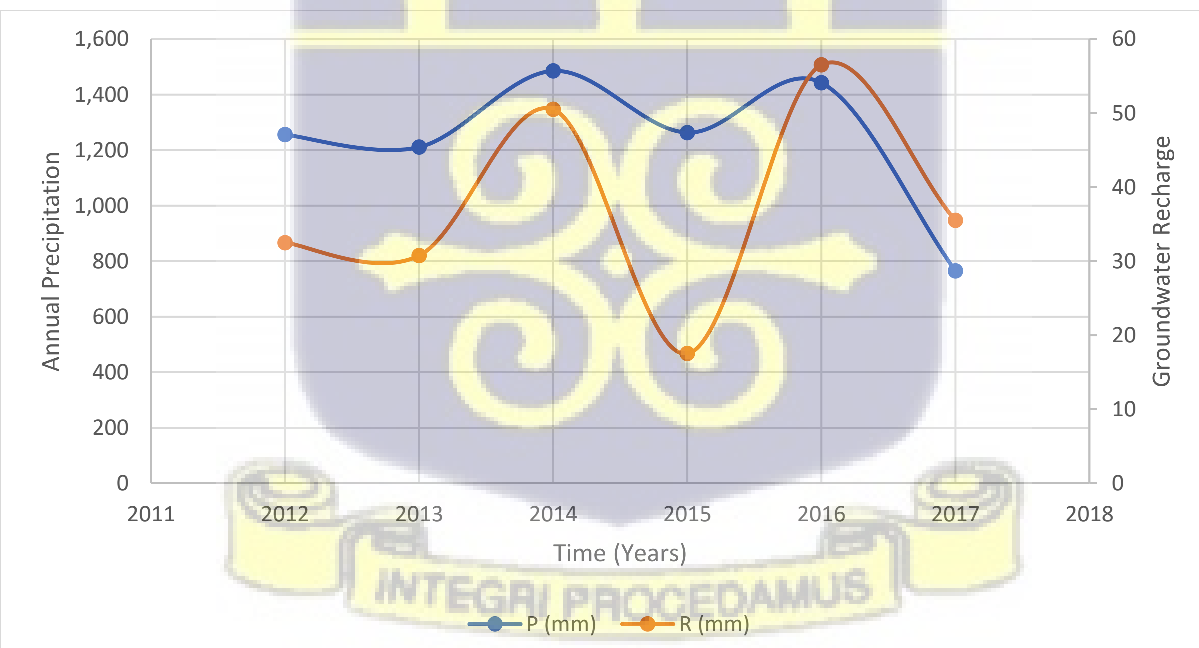


Figure 4.17: Annual variations in precipitation and groundwater recharge in borehole NMW8S (2012–2017)

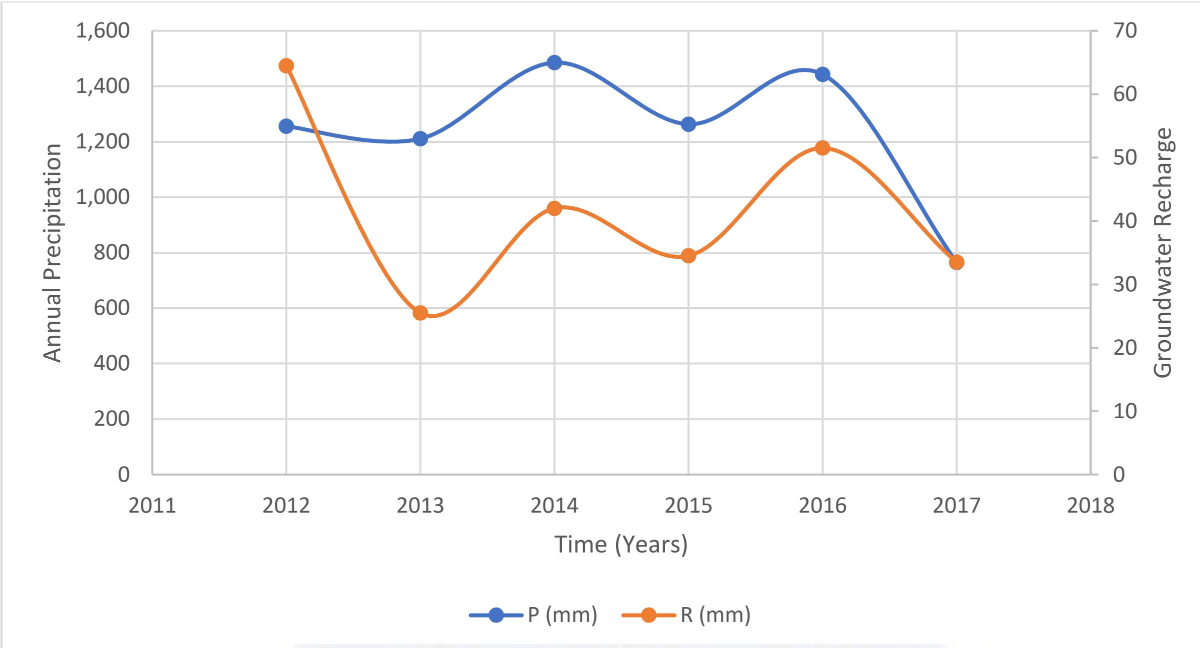


Figure 4.18: Annual variations in precipitation and groundwater recharge in borehole NMW10S (2012–2017)

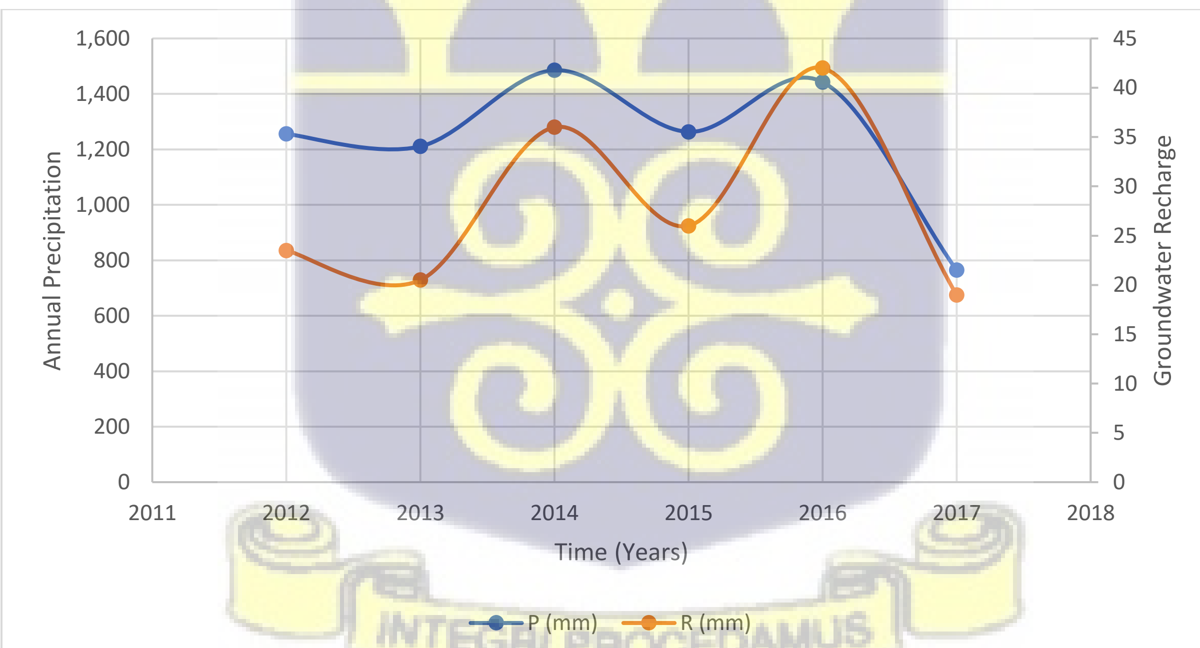


Figure 4.19: Annual variations in precipitation and groundwater recharge in borehole NMW12S (2012–2017)

Figure 4.20: Annual variations in precipitation and groundwater recharge in borehole NMW17S (2012–2017)

Figure 4.16 to 4.20 shows the recharge and rainfall yearly variation for Sy = 0.05. Although with smaller amounts, recharge follows rainfall changes in most boreholes: Since years with excessive rains have the tendency to enhance surface runoff, the peaks are less noticeable. When changes in groundwater levels are evaluated on a yearly basis, it is clear that all points demonstrate a net reduction in groundwater levels during the length of the research. In five of the six study years, below-mean precipitation deviations may have contributed to the reduction. Future research should determine which precipitation conditions give the most and least significant recharge to the aquifer, as well as the system's ability to provide temporal buffering (storage) to mitigate the impacts of drought and over-abstraction.

Groundwater storage and amount can be determined by analyzing inter-annual changes in groundwater levels. The research sites, for example, could reveal the system's resilience to provide the resource, notwithstanding the reductions of groundwater level caused by series of negative precipitation deviations over a long time. The sites, on the other hand, may illustrate what precipitation conditions cause the system to recover, based on the hypothesis that recharge is fueled by rainfall just once every ten years. Variations in patterns of precipitation are future climatic forecasts for many tropical and semi-tropical places; therefore, a complete study of recharge patterns in relation to the onset/distribution of precipitation and the frequency of extreme events will become increasingly relevant.

4.6.2 Water table fluctuation method over the six years period

Table 4.9 displays the outcomes of recharge of groundwater from the WTF method averaged over the six years period. With a specific yield of 0.05, the groundwater recharge estimate for borehole MW8 was 75.5 mm, while in borehole NMW8S, groundwater recharge estimated was 60 mm, representing 6.1% and 4.8% of the total rainfall received, respectively. Groundwater recharge in borehole NMW10S was projected to be 97.5 mm, or 7.8% of total rainfall received. Groundwater recharge estimates for borehole NMW12S was 40 mm, while in borehole NMW17S, groundwater recharge estimated was 250 mm representing 3.2% and 22.6% of the total rainfall received, respectively.

Table 4.9: Groundwater recharge from WTF method

Boreholes	Specific yield	Groundwater recharge (mm/year)	Groundwater recharge (%)
MW8	0.05	75.5	6.1
NMW8S	0.05	60	4.8
NMW10S	0.05	97.5	7.8
NMW12S	0.05	40	3.2
NMW17S	0.05	250	22.6

The WTF method can estimate groundwater recharge when water enters the water table at a faster pace than it leaves. Water levels increase as a result of this condition. Even though the borehole hydrograph indicates that levels of water are dropping, recharge can still occur. This means that the recharge rate is slower than when water moves out from the water table. If the flow of water distant from the water table equaled the steady recharge rate, the water level would not change, and the WTF model would rule out the possibility of a recharge (Healy and Cook, 2002). The hydrographs (Figures 4.11, 4.12, 4.13, 4.14, and 4.15) indicate that the graphs of rainfall amount and groundwater level fit each other closely, demonstrating a consistent association between groundwater levels and rainfall amount. As a result, the quantity of water recharged into the groundwater system is derived from the amount of rainfall collected during that rainfall event.

The values of recharged acquired in this research are plausible and were close and consistent to groundwater recharge outcomes acquired with the Water table fluctuation (WTF) and other techniques in earlier hydrogeological investigations conducted in similar geology in many arid or semi-arid regions in Africa (e.g., Ayenew et al., 2006; Darko and Krasny, 2003; Friesen et al., 2005; Martin, 2005; Nyagwambo, 2006; Obuobie, 2008; Sandwidi, 2007), despite the fact that the results of this study were slightly higher compared to some of the past research mentioned, it was anticipated that there would be some uncertainty. This is because the specific yield results utilized in this research were acquired from literature and not tested for the specific aquifers in the study area. Using specific yield outcomes obtained for the aquifers within the research region can improve the study's results' credibility.

The recharge estimations achieved using the WTF technique in this investigation are comparable to those obtained with the RIB model. An analysis of spatially approximated groundwater recharge estimates derived with the RIB and WTF methodologies in the research area demonstrates that the two techniques agree on possible maximum and lowest recharge zones in general. Recharge estimates calculated were 2.9 % to 21.4 % and 3.2 % and 22.6 % using the rainwater infiltration breakthrough model and water table fluctuation method, respectively. This slight variation in the results acquired using both rainfall infiltration breakthrough and water table fluctuation may be a result of the differences in defining parameters and methodologies. The RIB model takes into consideration the lag time and length of related rainfall events, which is dependent on a variety of conditions, such as the thick unsaturated zone, the texture of the soil, the vegetation's nature and size, and the geological properties of the aquifer. In the RIB model, the difference between consecutive departures is referred to as recharge. The WTF technique connects variations in observed water-level elevation to differences in the amount of water stored in the aquifer. The product of the rise in the water table and the specific yield is referred to as recharge.

The specific yield should represent the aquifer system of the catchment of interest for reliable and consistent calculations of groundwater recharge using the WTF and RIB methods. Laboratory values of specific yield are appropriate for detecting the impact of abstraction of an aquifer in the long run. However, the laboratory results of Sy are probably too big for groundwater recharge estimation using the WTF and RIB techniques (Sun et al., 2013). The current study's specific yield was taken from literature and then compared to the results acquired using the straight envelope line of the precipitation–groundwater rise points of each registered rainfall–recharge event, as well as specific yields obtained from other studies (Obuobie et al., 2012). The value that represents specific yield might have been substantially more reliable in the current study if the specific yield had been measured for the entire study area rather than just one point. Groundwater recharge estimates are limited by the specific yield, which is not indicative of the entire area. On the other hand, the results gained provide insight into how to use the procedure and produce outcomes that can be used to inform the additional investigation.

One of the conditions for successful use of the WTF method and RIB approach is to select boreholes that are representative of the entire research catchment. As a result, the utilization of more wells for collecting groundwater samples in recharge studies is more important. Wells chosen must be representative of the entire study region. Diouf et al., (2013) employed 30 drilled wells, boreholes, and piezometers to obtain groundwater level data in their investigation. Data from wells that are representative of the entire area are used to improve groundwater recharge estimates. The groundwater recharge estimations' dependability and actuality are hampered by the use of only a few boreholes. Future research should include more boreholes for groundwater level data, according to the findings of this study.

CHAPTER FIVE

CONCLUSIONS AND RECOMMENDATIONS

5.1 Introduction

This chapter presents the findings and contributions of the research. This section draws inferences from all results obtained from the present study. From the interpretation made on the estimation on groundwater recharge using the Rainfall infiltration breakthrough model and Water level fluctuation method, as well as scenarios to forecast groundwater level and recharge changes under climate change, namely rainfall reductions. A conclusion will be drawn in light of the objectives of the study. Thus, deductions will be given on groundwater recharge within the study area.

5.2 Conclusions

This work presents a conceptual analysis of groundwater recharge estimation based on available monthly water level measurements and rainfall. The focus of the research was to develop an improved groundwater recharge estimate from rainfall and groundwater level in the Birimian Province in Southwestern Ghana that can be applied and modified for improved water management in the future. For the purpose of this work, groundwater recharge models, rainfall infiltration breakthrough, and water table fluctuation were employed to estimate the amount of groundwater recharge. For the development of the recharge models, all available data were integrated and processed.

The specific yield data acquired from the literature were compared to those determined using the linear regression model as quality assurance. This confirmed the analysis's dependability, i.e., the association between rainfall and groundwater level. The value of Sy = 0.06 was obtained from the envelope straight line of the precipitation–groundwater rise points of each registered rainfall–recharge event.

The RIB model and WTF method were applied in five existing observation wells. The groundwater recharge as a percentage of monthly rainfall was estimated for each of the five observation wells. With the specific yield of 0.05, groundwater recharge was estimated using the RIB method, ranging between 2.9 % and 21.4 % of MAP at the monthly scale. Using the WTF model, with the specific yield of 0.05, annual groundwater recharge was estimated, ranging between 3.2 % and 22.6 % of the mean annual rainfall. The results revealed that the sensitivity of recharge calculated using the RIB model is higher for boreholes NMW17S and NMW10S than for boreholes MW8, NMW8S, and NMW12S, based on these analyses. Groundwater recharge estimates derived from these two techniques are in a similar range and are also in agreement with groundwater recharge estimates derived from previous studies conducted in the same area. The results of these methods show that they are applicable when all of the assumptions required for their successful application are met.

The analysis of water table fluctuation in response to rainfall from 2012 to 2017 for all boreholes showed that the only time the water level rises is when rainfall event has occurred. The analysis showed that the area receives high rainfall in the major raining season (between late March and July), and the minor season occurs between September and November. The results also showed that the two to four months between the beginning of the rainy season and the start of groundwater level rise could be thought of as a period of soil replenishment caused by moisture shortages from the previous dry season. In the examined area, the lag indicates that there are impacts of threshold and a non-linear correlation between recharge and rainfall. Furthermore, the lag indicates that most of the boreholes in the studied area do not recharge quickly. Results indicated that recharge follows rainfall changes in most boreholes, although with smaller amounts: the peaks are less pronounced since years with excessive rainfall have the tendency to enhance surface runoff.

The prediction of the climate scenario showed that less rainfall had no influence on groundwater levels over the modeled timeframe. The recharge rate to precipitation ratio, on the other hand, did not alter considerably; It was somewhat higher than the baseline (average recharge percentage is 3.73 %, 4.20 %, and 4.80 % of corresponding rainfall with 10 %, 20 %, and 30 % rainfall reduction, respectively).

5.3 Recommendations

The following are recommended for future work on groundwater recharge estimation.

- The specific yield should be reflective of the aquifer system of interest for valid and consistent predictions of groundwater recharge utilizing the water table fluctuations and rainfall infiltration breakthrough models. Using specific yield values calculated from the aquifers in the study area can improve the reliability of the study results.
- The fraction of withdrawal through pumping wells that can act to recharge the groundwater table should be incorporated in the calibration of future research.
- The research should be extended to cover the entire area, and long-term regular and periodic groundwater data collection and monitoring are needed for simulating a model. The time frame of this project and the lack of input data did not allow to investigate this modeling as much as it could be, but it provided a good basis for further studies.

6.0 REFERENCES

Abid, K., Zouari, K., Abidi, B., (2010). Identification and characterization of hydrogeological relays of continental intercalaire aquifer of southern Tunisia. *Carbonates and Evaporites*, *25*(1), 65-75.

Acheampong SY. (1996). Geochemical evolution of the shallow groundwater system in the southern Voltaian Sedimentary Basin of Ghana. Ph.D., University of Nevada, Reno.

Acheampong, S.Y., Hess, J.W., (2000). Origin of the shallow groundwater system in the southern Voltaian Sedimentary Basin of Ghana: an isotopic approach. *Hydrogeology Journal 233*: 37–53.

Adams. S., Titus. R., Xu. Y., (2004). Groundwater recharge assessment of the basement aquifers of central Namaqualand. Report to the Water Research Commission. WRC Report No. 1093/1/04. ISBN: 1-77005-214-3.

Agyekum, W. A., (2004). Groundwater resources of Ghana with the focus on international shared aquifer boundaries. UNESCO-ISARM International Workshop – Managing shared aquifer resources in Africa. B. Appelgreen. Tripoli, Lybia, June 2002. (No. 8). IHP-VI Series on groundwater no.8. Lybia.

Ahmadi, T., Ziaei, A.N., Davari, K., Rasoulzadeh, A., Faridhoseini, A.R., Izadi A., (2012). Estimation of groundwater recharge using various methods in Neishaboor Plain, Iran, Iran. In 5th international groundwater symposium, Kuwait, (9-15). in: Hadi & Copty (Eds.), Taylor & Francis Group: Groundwater Modeling and Management under Uncertainty.

Ahmadi, T., Ziaei, A.N., Rasoulzadeh, A., Davary, K., Esmaili, K., Izady, A., (2014). Mapping groundwater recharge areas using CRD and RIB methods in the semi-arid Neishaboor Plain, Iran. *Arabian Journal of Geosciences* *8*(5): 2921–2935. https://doi.org/10.1007/s12517-014- 1321-2.

Aishlin. P. S., (2006). Groundwater Recharge Estimation Using Chloride Mass Balance Dry Creek Experimental Watershed. Masters of Science in Hydrologic Sciences Boise State University. USA.

Albhaisi, M., Brendonck, L., Batelaan, O., (2013). Predicted impacts of land-use change on groundwater recharge of the upper Berg catchment, South Africa. *Water SA* Vol. 39 No. 2.

Allison, G.B., Barnes, C.J., Hughes, M.W., Leaney, F.W.J., (1984). Effect of climate and vegetation on oxygen-18 and deuterium profiles in soils. Isotope *Hydrology*. Proc. Symp. Vienna, IAEA, 105-123.

Allison, G.B., Cook, P.G., Barnett, S.R., Walker, J.R., Jolly, I.D., and Hughes, M.W. (1990). Land clearance and river salinization in the western Murray Basin, Australia. *J Hydrol*. 9: 1-20.

Allison, G.B., Gee, G.W., Tyler, S.W., (1994). Vadose-zone Techniques for estimating groundwater recharge in arid and semi-arid regions. *Soil sci. Soc. Am. J.*, 58, 6-14.

AMEC Geomatrix, Inc., (2010). Draft Baseline Groundwater Investigation 2008-2010 Technical Report, Akyem Project Ghana, West Africa. Submitted to Newmont Golden Ridge Ltd.

Anon., (2006). "Birim North District Profile" http://ghanadistricts.gov.ghdistrict/? Accessed: February 22, 2016.

Antwi, F., (2010). An Assessment of The Impacts of Newmont Gold Ghana Limited (NGGL) on the socio-economic development after six years of operation in the Birim North District. Master of Arts. Kwame Nkrumah University of Science and Technology, Kumasi.

Atta-Darkwa, T., Kyei-Baffour, N., Ofori, E., Mensah, E. & Agyare, W. A., (2013). Quantification Of Groundwater Recharge in The River Oda Catchment Using the Water table Fluctuation Method. *G.J. E.D.T.*, *Vol. 2*(1) 2013:96-103.

Attiogbe. F., Nkansah. A., (2017). The Impact of Mining on the Water Resources in Ghana: Newmont Case Study at Birim North District (New Abirem). *Energy and Environment Research;* Vol. 7, No. 2; ISSN 1927-0569.

Attua, E. M., Annan, S.T., Nyame, F., (2014). Water quality analysis of rivers used as drinking sources in artisanal gold mining communities of the Akyem-Abuakwa area: A multivariate statistical approach. *Ghana Journal of Geography* Vol. 6, 2014 Pages 24 – 41.

Atule, G.K., (2007). "Petrology, Geochemistry and Ore Mineralogy of the Yaayaso Gold Deposit, New Abirem, Ghana," Unpublished MSc. Thesis Report, University of Mines and Technology, Tarkwa, Ghana, pp. 23.

Ayenew, T., Alemayehu, T., Kebede, S., (2006). Environmental Isotopes and Hydrochemical Studies as Applied to Water Resources Assessment in Central and Southern Afar and Adjacent Highlands. Ethiopian Science and Technology Agency. Unpublished Report of the International Atomic Energy Agency Project (ETH/8/008). Addis Ababa, Ethiopia, p. 128.

Baalousha H: (2005). Using CRD method for quantification of groundwater recharge in the Gaza strip, Palestine. *J Environ Geol* 48:889– 900. doi:10.1007/s00254-005-0027-x.

Banoeng-Yakubo, B., Danso, S., and Tumbulto, J. (2005). Assessment of Pollution Status and Vulnerability of Water supply Aquifers in Keta, Ghana. UNEP/UNESCO/UNHABITAT/ECA.

Bates, B., Kundzewicz, Z., Wu, S., & Palutikof, J. (2008). Climate change and water. Technical paper of the intergovernmental panel on climate change. Climate Change Policy with a Renewed Environmental Ethics, 21, 85–101.

Bates, B.C., Kundzewicz, Z.W., Wu, S., Palutikof, J.P. 2008. Climate change and water. (Technical paper of the Intergovernmental Panel on Climate Change.).

Bazuhair, A.S., Wood, W.W., (1995). Chloride mass-balance method for estimating groundwater recharge in arid areas: an example from western Saudi Arabia. *Journal of Hydrology* 186: 153–159.

Beekman, H.E., Xu, Y., (2003). Review of groundwater recharge estimation in arid and semi-arid Southern Africa. In: Xu, Y., Beekman, H.E. (Eds.), Groundwater Recharge Estimation in Southern Africa. UNESCO IHP Series No. 64, UNESCO Paris. ISBN 92-9220-000-3.

Bhattacharya, S. K., Froehlich, K., Aggarwal, P. K. and Kulkarni, K. M. (2003). Isotopic variation in Indian monsoon precipitation. Records from Bombay and New Delhi. *Geophysical Research Letters*. 30: 2285–2288.

Birim North District Planning Coordinating Unit (2006). Birim North District medium-term development plan—2006–2009: the profile of the Birim North District. Birim North District Assembly, New Abirem, pp 5–57.

Boerner, A., Weaver, C., (2012). Nebraska recharge estimation from chloride mass balance. Nebraska.

Bogaard T, Guglielmi Y, Marc V, Emblanch C, Bertrand C, Mudry J. 2007. Hydrogeochemistry in landslide research: a review. *Bulletin de la Societe Geologique de France 178*: 113–126.

Bredenkamp, D.B., Botha, L.T., Van Tonder, G.J., Van Rensburg, H.J., (1995). Manual on quantitative estimation of groundwater recharge and aquifer storativity, *Water Research Commission*, Pretoria, pp. 363.

Brielmann, H., Griebler, C., Schmidt, S.I., Michel, R., Lueders, T., (2009). Effects of thermal energy discharge on shallow groundwater ecosystems. *FEMS Microbiol. Ecol.* 68 (3), 273–286. https://doi.org/10.1111/j.1574-6941.2009.00674.x.

Brunner, P., Bauer, P., Eugster, M., Kinzelbach, W., (2004). Using remote sensing to regionalize local precipitation recharge rates obtained from the chloride method. *J. Hydrol.*, *294* (4). pp. 241-250.

Butterworth. J.A., Schulze. R.E., Simmonds. L.P., Moriarty. P., Mugabe. F., (199). hydrological processes and water resources management in a dryland environment IV: long term groundwater level fluctuations due to variation in rainfall. *Hydrology and Earth System Sciences*, 3(3), 353-361.

Carrier, M., Lefebvre, R., Racicot, J., and Asare, E. (2008a). Groundwater recharge assessment in northern Ghana using soil moisture balance and chloride mass balance, 1437–1444.

Cartwright, I., Hall, S., Tweed, S., Leblanc, M., (2009). Geochemical and isotopic constraints on the interaction between saline lakes and groundwater in southeast Australia. *Hydrogeology Journal*, *17*(8), 1991-2004.

Chen, Z., Nie, Z., Zhang, G., Wan, L. and Shen, J. (2006). Environmental Isotopic Study on the Recharge and Residence time of Groundwater in the Heihe River Basin, northwestern China *Hydrogeology Journal*, *14*(8), 1635-1651.

Clear Creek Consultants Inc., (2010). Newmont Golden Ridge Limited Proposed Akyem Gold Mine Project- Ghana, Africa Meteorological Monitoring Program Data Summary and Review: the Year 2009. Prepared for Newmont Golden Ridge Resources Ltd. January 2010.

Craig H., (1961). Isotopic variations in meteoric waters. *Science*, 133: 1702-1703.

Crosbie, R.S., Binning, P., Kalma. J.D., (2005). A time-series approach to inferring groundwater recharge using the water table fluctuation method. *Water Resour. Res*. 41 1–9.

Crosbie, R.S., Pickett, T., Mpelasoka, F.S., Hodgson, G., (2013). An assessment of the climate change impacts on groundwater recharge at a continental scale using a probabilistic approach with an ensemble of GCMs. *Clim. Change* 117 (1), 41–53. https://doi.org/10.1007/s10584-012-0558-6.

Dafny, E., Burg, A., Gvirtzman, H., (2006). Deduction of groundwater flow regime in a basaltic aquifer using geochemical and isotopic data: the Golan Heights, Israel case study. *Journal of Hydrology* 330: 506–524.

Darko, P.K., Krasny, J., (2003). "Regional transmissivity distribution and groundwater potential in hard rocks of Ghana" *In J*. Krásný, Z. Hrkal, and J. Bruthans, eds. Proceedings of IAH international conference on groundwater in fractured rocks. Prague, Czech Republic, 15.-19. Sept. 2003. IHP-VI, Series on Groundwater, No. 7 *Prag*. pp. 45–46.

Datta, P.S., (1999). Groundwater Situation in New Delhi: Red Alert, Nuclear Research Laboratory, IARI, New Delhi-12.

De Vries J, Simmers I., (2002). Groundwater recharge: an overview of processes and challenges. *J Hydrogeol* 10:5–17. doi:10.1007/ s10040-001-0171-7.

Delin G. N., Healy, R.W., Lorenz, D. L. and Nimmo, J. R., (2006). Comparison of local-to regional scale estimates of ground-water recharge in Minnesota, USA. *Journal of Hydrology* 334(1-2, 20): 231-249.

Demlie, M., Ayenew, T. and Wohnlich, S., (2007). Comprehensive hydrological and hydrogeological study of topographically closed lakes in highland Ethiopia: The case of Hayq and Ardibo. *Journal of Hydrology*, *339*(3-4), 145-158.

Demlie, M., Wohnlich, S. and Ayenew, T., (2008). Major ion hydrochemistry and environmental isotope signatures as a tool in assessing groundwater occurrence and its dynamics in a fractured volcanic aquifer system located within a heavily urbanized catchment, central Ethiopia. *Journal of Hydrology, 353*(1-2), 175-188.

Diouf, O. C., Faye, S. C., Diedhiou, M., Kaba, M., Faye, S., Gaye, C. B., and Wohnlich, S., (2013). Combined uses of water-table fluctuation (WTF), chloride mass balance (CMB), and environmental isotopes methods to investigate groundwater recharge in the Thiaroye sandy aquifer (Dakar, Senegal). *African Journal of Environmental Science and Technology, 6*, 425-437.

Duriez. S., (2005). On the Use of Groundwater Contaminant Transport Modelling in Risk Assessments. Master's Thesis in Applied Environmental Measurement Techniques. Chalmers University of Technology, Göteborg, Sweden.

Earman, S. & Dettinger, M., (2011). Monitoring networks for long-term recharge change in the mountains of California and Nevada. In California Energy Commission PIER Workshop Paper CEC-500-2008-006. Available from: http://www. energy.ca.gov/2008publications/CEC-500-2008-006/CEC500-2008-006.PDF

Epting, J., Michel, A., Affolter, A., Huggenberger, P., (2020). Climate change effects on groundwater recharge and temperatures in Swiss alluvial aquifers: *Journal of Hydrology X 11* – 100071. https://doi.org/10.1016/j.hydroa.2020.100071

Fabryka-Martin, J., Davis, S. N., and Elmore, D., (1987). Applications of 129I and 36Cl to hydrology. *Nucl. Instrum. Methods Phys. Res.*, B29, 361–371.

Fayer, M. J., Gee, G. W., Rockhold, M. L., Freshley, M. D., and Walters, T. B., (1996). Estimating Recharge Rates for a Groundwater Model Using a GIS. *Journal of Environmental Quality*, 25(3), 510–518.

Fontes J.C., Edmunds W.M., (1989). The use of environmental isotopes techniques in arid zone hydrology-a critical review. *Technical documents in hydrology*. UNESCO, Paris.

Friesen, J., Andreini, M., Andah, W., Amisigo, B., van de Giesen, N., (2005). Storage capacity and long-term water balance of the Volta Basin, West Africa. *IAHS Publication no*. 296, pp. 138–145.

Frimpong, M., Ewusi, A., Amoako, R., and Appianing, E. J. A., (2016), "Assessment of Groundwater Quality in the Akyem Mine Area of Newmont Golden Ridge Ltd (NGRL)," 4th UMaT Biennial International Mining and Mineral Conference, pp. ES 55-63.

Gaye, C.B., Edmunds, W.M., (1996). Groundwater recharge estimation using chloride, stable isotopes, and tritium in the sands of Northwestern Senegal. *Environ. Geol*. 27:246-251.

Ghafari H, Rasoulzadeh A, Raoof M, Esmeali A. (2018). Estimation of natural groundwater recharge using WTF method (case study: Ardabil Plain aquifer). *Journal of Civil and Environmental Engineering* 48(1): 43–52 (in Persian).

Ghana Business News (13th March 2010). Birim River is Losing its Importance. www.ghanabusinessnews.com Accessed 24th July 2012.

Ghasemizadeh. R. (2015). Modeling Groundwater Flow and Contaminant Transport in the North Coast Limestone karst aquifer system of Puerto Rico. Ph.D. Thesis in Civil and Environmental Engineering. Northeastern University Boston, Massachusetts.

Golden Ridge Resources Limited, (2005). Akyem Project Baseline Survey- Hydrogeology Report Chapter 4 DRAFT. Prepared by SGS Environment.

Golden Ridge Resources Limited, (2006). Akyem Environmental Impact Statement. May 2006.

Golder and Associates (Golder), (2004). Hydrogeologic Conditions at Akyem. Technical Memorandum to NGGL.

Golder, (2006). Akyem Pumping Tests. Technical Memorandum.

González-Trinidad. J., Pacheco-Guerrero. A., Júnez-Ferreira. H., Bautista-Capetillo. C., Hernández-Antonio. A. (2017). Identifying Groundwater Recharge Sites through Environmental Stable Isotopes in an Alluvial Aquifer. *Water*. 9, 569; doi:10.3390/w9080569.

Green, T.R., Taniguchi, M., Kooi, H., Gurdak, J.J., Allen, D.M., (2011). Beneath the surface of global change: Impacts of climate change on groundwater. *J. Hydrol.* 405 (3–4), 532–560. https://doi.org/10.1016/j. jhydrol.2011.05.002.

Gyau-Boakye, P. (2001). Sources of Rural Water Supply in Ghana. *Water International*, 26(1), 96–104. http://doi.org/10.1080/02508060108686689.

Hay, J.E., (1997). An investigation of groundwater recharge along the western flank of the southern Bridger Range, southwestern Montana. Thesis. Montana State University. Bozeman, MT.

Healy, R. W., Cook, P. G., (2002). Using groundwater levels to estimate recharge. *Hydrogeology Journal*, *10*(1), 91–109. http://doi.org/10.1007/s10040-001-0178-0.

Healy, R.W., (2010). Estimating groundwater recharge. Cambridge University Press, Cambridge.

Huggenberger, P., Epting, J., (2011). In: Urban Geology - Process-Oriented Concept for Adaptive and Integrated Resource Management. *Springer*, Basel. https://doi.org/ 10.1007/978-3-0348-0185-0.

Intergovernmental Panel on Climate Change (2007) Summary for policymakers. In: Climate Change 2007 The Physical Science Basis. Contribution of Working Group I to the Fourth

Assessment Report of the Intergovernmental Panel on Climate Change. Available from: http://water.state.co.us/ DWRIpub/DWR%20Presentations/kknox_090606.pdf.

Ireland T.J., P. Knoetze, D. Midot,, and G.K. Atule. (2001). Discovery of the Akim Gold Deposit, Ghana. Newgengold Gold.

Jesußek, A., Grandel, S., Dahmke, A., (2013). Impacts of subsurface heat storage on aquifer hydrogeochemistry. Environ. *Earth Sci. 69*, 1999–2012. https://doi.org/10.1007/ s12665-012-2037-9.

Johnston, R. M., & McCartney, M. (2010). Inventory of water storage types in the Blue Nile and Volta River basins. Retrieved from http://dx.doi.org/10.5337/2010.214.

Jorgensen, N.O., Banoeng-Yakubo, B.K., (2001). Environmental isotopes (18O, 2 H, and 87Sr/86Sr) as a tool in groundwater investigations in the Keta Basin. Ghana. *Hydrogeology Journal 9*: 190–201.

Kankam-Yeboah, K, Obuobie, E., & Amisigo, B. (2009). Climate Change Impacts on Water Resources in Ghana. Ghana National Commission for UNESCO, 65–69.

Kankam-Yeboah, K., Dapaah-Siakwan, S., Nishigaki, M., & Komatsu, M. (2003). The hydrogeological setting of Ghana and the potential for underground dams. **岡山大学環境理工学部研究報告**, 8(1), 39–52.

Kattan, Z. (2001). Use of hydrochemistry and environmental isotopes for evaluation of groundwater in the Paleogene limestone aquifer of the Ras Al-Ain area (Syrian Jazirah). *Journal of Environmental Geology*, 41, 128–144.

Kesse G. O., (1985) "The Mineral and Rock Resources of Ghana," A. A. Balkema Publisher, Rotterdam.

Kipfer, R., Livingstone, D.M. 2008. Water resources and climate change. In: News, E. (Ed.).

Kippo, J.V., 2012. DRAI-AHP Vulnerability Assessment of the Quarternary aquifer in the Keta Strip. Ph.D. Thesis, University of Ghana.

Krautstrunk, M. L. (2012). An Estimate of Groundwater Recharge in the Nabogo River Basin, Ghana Using Water Table Fluctuation Method and Chloride Mass Balance.

Kumar, C. P. (2015). Modeling of Groundwater Flow and Data Requirements. *International Journal of Modern Sciences and Engineering Technology (IJMSET), 2* (2): pp. 18-27.

Kurylyk, B.L., MacQuarrie, K.T.B., (2013). The uncertainty associated with estimating future groundwater recharge: A summary of recent research and an example from a small unconfined aquifer in a northern humid-continental climate. *J. Hydrol. 492*, 244–253. https://doi.org/10.1016/j.jhydrol.2013.03.043.

Kurylyk, B.L., MacQuarrie, K.T.B., McKenzie, J.M., (2014). Climate change impacts on groundwater and soil temperatures in cold and temperate regions: Implications, mathematical theory, and emerging simulation tools. *Earth Sci. Rev. 138*, 313–334. https://doi.org/10.1016/j.earscirev.2014.06.006.

Kwoyiga, L., & Stefan, C. (2019). Institutional Feasibility of Managed Aquifer Recharge in Northeast Ghana. Sustainability, *MDPI, 11*(2), 378.

Lewis, F.M., Walker, G.R., (2002). Assessing the potential for significant and episodic recharge in Southern Western Australia using rainfall data. *Hydrogeol. J*.10:229237.

Liggett, J.E., Allen D.M., (2009). Comparing approaches for modeling spatially distributed direct recharge in a semi-arid region (Okanagan Basin, Canada) *Hydrol. J., 18* (2) pp. 339-357.

Loheide II, S.P., Butler, J.J., Gorelick, S.M., (2005). Estimation of groundwater consumption by phreatophytes using diurnal water table fluctuations: A saturated-unsaturated flow assessment. *Water Resources Research 41*, 1 of 14.

Lutz, A., Minyila, S., Saga, B., Diarra, S., Apambire, B., Thomas, J., (2014). Fluctuation of Groundwater Levels and Recharge Patterns in Northern Ghana*: Climate*, *3*, 1-15; doi:10.3390/cli3010001

Lycopodium Knight Piesold, (2005). Akyem Project Summary of Geotechnical Site Investigation.

Majdabady. M.P., Rasoulzadeh. A., Kanoon. A., Ahmadzadeh. G., (2020). Estimation Of Groundwater Recharge Originating from Agricultural Irrigation and Rainfall in Shyramyn Plain, Iran. *Irrigation And Drainage*. *69*: 107–120.

Malcom, R. & Soulsby, C. (2000) Modelling the potential impact of climate change on a shallow coastal aquifer in northern Scotland. *In:* Groundwater in the Celtic Regions: Studies in Hard Rock and Quaternary Hydrogeology (N. S. Robbins & B. D. R. Misstear, eds.). Special Publication 182. Geological Society Special Publication 182, London, pp. 191–204.

Mare´chal J.C., Dewandel, B., Ahmed S., Galeazzi L., Zaidi, F.K., (2006). Combined estimation of specific yield and natural recharge in a semi-arid groundwater basin with irrigated agriculture. *Journal of Hydrology 329*, 281– 293.

Marei, A., Khayat, S., Weise, S., Ghannam, S., Sbaih, M., and Geyer, S. (2010). Estimating groundwater recharge using the chloride mass-balance method in the West Bank, Palestine. *Hydrological Sciences Journal*, 55(5), 780–791. http://doi.org/10.1080/02626667.2010.491987.

Martin, N., (2005). Development of a water balance for the Atankwidi catchment, West Africa – A case study of groundwater recharge in a semi-arid climate. Doctoral thesis. University of Göttingen.

McCartney, M., Forkuor, G., Sood, A., Amisigo, B., Hattermann, F., Muthuwatta, L., (2012). The Water Resource Implications of Changing Climate in the Volta River Basin IWMI Research Report 146, International Water Management Institute (IWMI), Colombo, Sri Lanka. pp. 1-3.

Meixner, T., Manning, A.H., Stomestrom, D.A., Allen, D.M., (2016). Implications of projected climate change for groundwater recharge in the western United States. *J. Hydrol. 534*, 124–138. https://doi.org/ 10.1016/j.jhydrol.2015.12.027.

Menberg, K., Blum, P., Kurylyk, B.L., Bayer, P., (2014). Observed groundwater temperature response to recent climate change. *Hydrol. Earth Syst. Sci. 18* (11), 4453–4466. https://doi.org/10.5194/hess-18-4453.

Mensah, F. O., Alo, C., and Yidana, S. M. (2014). Evaluation of Groundwater Recharge Estimates in a Partially Metamorphosed Sedimentary Basin in a Tropical Environment: Application of Natural Tracers, 2014.

Ministry of Environment, Science, and Technology. (2013). Ghana National Climate Change Policy. Retrieved from http://www.un-page.org/files/public/ghanaclimatechangepolicy.pdf.

Moeck, C., Brunner, P., Hunkeler, D., (2016). The influence of model structure on groundwater recharge rates in climate-change impact studies. *Hydrogeol. J.* 24 (5), 1171–1184. https://doi.org/10.1007/s10040-016-1367-1.

Mook, W.G. 2006. Introduction to Isotope Hydrology: Stable and Radioactive Isotopes of Hydrogen, Oxygen, and Carbon, Taylor & Francis Group, London, UK.

Moon, S.K., Woo, N.C., Leeb, K.S., (2004). Statistical analysis of hydrographs and water table fluctuation to estimate groundwater recharge. *Journal of Hydrology 292*, 198–209.

Moore, J. E. (2002). Field Hydrogeology. Florida: Lewis Publishers.

Mul, M., Obuobie, E., Appoh, R., Kankam-Yeboah, K., Bekoe-Obeng, E., Amisigo, B., … McCartney, M. (2015). Water resources assessment of the Volta River Basin (Vol. 166). Colombo, Sri Lanka: International Water Management Institute (IWMI).

Mutoti. M. I., (2015). Estimating groundwater recharge using chloride mass balance in the upper Berg River catchment, South Africa. Masters of Science (MSc) in Environmental and Water Sciences. University of the Western Cape Western Cape, South Africa.

Nartey. V. K., Klake. R K., Hayford. E. K., Doamekpor. L. K., Appoh. R. K; (2011). Assessment of Mercury Pollution in Rivers and Streams around Artisanal Gold Mining Areas of the Birim North District of Ghana. *Journal of Environmental Protection. 2*, 1227-1239. doi:10.4236/jep.2011.29141.

Négrel, P.h., Lemiére, B., Machard de Grammont, H., Billaud, P., Sengupta, B., (2007). Hydrogeochemical processes, mixing and isotope tracing in hard rock aquifers and surface waters from the Subarnarekha River Basin, (east Singhbhum District, Jharkhand State, India). *Hydrogeology Journal 15*: 1535–1552.

Nemaxwi., P., Odiyo., J.O., Makungo., R. (2019). Estimation of groundwater recharge response from rainfall events in a semi-arid fractured aquifer: Case study of quaternary catchment A91H, Limpopo Province, South Africa. *Cogent Engineering. 6:* 1635815 https://doi.org/10.1080/23311916.2019.1635815.

Ng, G. H. C., McLaughlin, D., Entekhabi, D. and Scanlon, B. (2009). Using data assimilation to identify diffuse recharge mechanisms from chemical and physical data in the unsaturated zone, *Water Resources Research*: 45.

Nyagwambo, N.L., (2006). Groundwater recharge estimation and water resources assessment in a tropical crystalline basement aquifer. Ph.D. thesis. UNESCO-IHE. AA Balkema Publishers. The Delft University of Technology. The Netherlands.

Obuobie, E. (2008). Estimation of groundwater recharge in the context of future climate change in the White Volta River Basin, West Africa, (62), 164.

Obuobie, E., Diekkrueger, B., Agyekum, W., and Agodzo, S. (2012). Groundwater level monitoring and recharge estimation in the White Volta River basin of Ghana. *Journal of African Earth Sciences*, 71-72, 80–86. http://doi.org/10.1016/j.jafrearsci.2012.06.005.

Obuobie, E., Diekkrueger, B., and Reichert, B. (2010). Use of chloride mass balance method for estimating the groundwater recharge in northeastern Ghana. *International Journal of River Basin Management*, *8*(3-4), 245–253.

Ophori DU. 1999. Constraining permeabilities in a large-scale groundwater systemthrough model calibration. *Journal of Hydrology 224*: 1–20.

Owusu, E. H (2012). Natural Resources of Okyeman – an overview (2012). *West African Journal of Applied Ecology, 20* (3): 47-52.

Oyebande, L., & Odunuga, S. (2010). Climate change impact on water resources at the transboundary level in West Africa: The cases of the Senegal, Niger and Volta Basins. *Open Hydrology Journal*, 4(1), 163–172.

Pelig-Ba, K. B. (2009). Analysis of stable isotope contents of surface and underground water in two main geological formations in the Northern Region of Ghana. *West African Journal of Applied Ecology, 15*. http://doi.org/10.4314/wajae.v15i1.49430.

Peng, T.R., Huang, C.C., Wang, C.H., Liu, T.K., Lu, W.C., Chen, K.Y., (2012a). Using oxygen, hydrogen, and tritium isotopes to assess pond water's contribution to groundwater and local precipitation in the pediment tableland areas of northwestern Taiwan. *Journal of Hydrology* 450–451: 105–116.

Peng, T.R., Wang, C.H., Hsu, S.M., Chen, N.C., Su, T.W., Lee, J.F., (2012b). Use of stable water isotopes to assess sources and influences of slope groundwater on slope failure. *Hydrology Proceeding* 26: 345–355.

Peng, T.R., Wang, C.H., Hsu, S.M., Wang, G.S., Su, T.W., Lee, J.F., (2010a). Identification of groundwater sources of a local-scale creep slope: using environmental stable isotopes as tracers. *Journal of Hydrology 381*: 151–157.

Peng, T.R., Wang, C.H., Huang, C.C., Fei, L.Y., Chen, C.T.A., Hwong. J.L., (2010b). Stable isotopic characteristic of Taiwan's precipitation: a case study of Western Pacific monsoon region. *Earth and Planetary Science Letters 289*: 357–366.

Peng, T.R., Wang, C.H., Lai, T.C., Ho, S.K., (2007). Using hydrogen, oxygen, and tritium isotopes to identify the hydrological factors contributing to landslides in a mountainous area, central Taiwan. *Environmental Geology 52*: 1617–1629.

Petersen, R., M. (2012). A conceptual understanding of groundwater recharge processes and surface-water/ groundwater interactions in the Kruger national park. Master's degree. The University of the Western Cape, Bellville, South Africa.

Ping, J., Nichol, C., Wei, X., (2014). Quantification of groundwater recharge using the chloride mass balance method in semi-arid mountain terrain, South Interior British Columbia, Canada.

Plummer, L.N. (2003). Environmental tracers and how they are used to understand the aquifer. pp. 82-83; from James R. Bartolino and James C. Cole, eds., 2003, U.S. Geological Survey Circular 1222.

Possemiers, M., Huysmans, M., Batelaan, O., (2014). Influence of Aquifer Thermal Energy Storage on groundwater quality: A review illustrated by seven case studies from Belgium. *J. Hydrol. Reg. Stud*. 2, 20–34.

Qablawi, B., (2016). A Comparison of Four Methods to Estimate Groundwater Recharge for Northeastern South Dakota: Master of Science- South Dakota State University.

Rapti-Caputo, D., Martinelli, G., (2009). The geochemical and isotopic composition of aquifer systems in the deltaic region of the Po River plain (northern Italy). *Hydrogeology Journal 17*: 467–480.

Risser, D.W., Gburek, W.J., and Folmar, G.J., (2005b). Comparison of methods for estimating groundwater recharge and base flow at a small watershed underlain by fractured bedrock in the Eastern United States. US Geological Survey Scientific Investigation Report 2005-5038.

Rozanski K. (1985). Deuterium and oxygen-18 in European groundwaters – links to atmospheric circulation in the past. *Chemical Geology. (Isot. Geoscience Section)* 52: 349–363.

Russell, C.E., Minor, T., (2002). Reconnaissance estimates of recharge based on an elevation-dependent chloride mass-balance approach, DOE/NV/11508-37, Publication No. 45164. Prepared for the US Department of Energy, National Nuclear Security Administration Nevada Operations Office. Desert Research Institute, Las Vegas, NV.

Saidi, S., Bouri, S. and Dhia, H., (2009). Groundwater vulnerability and risk mapping of the Hajeb-jelma aquifer (Central Tunisia) using a GIS-based DRASTIC model. *Environ. Earth Sci*. 59:1579–1588.

Salem ZE, Sakura Y, Mohammed Aslam MA. (2004). The use of temperature, stable isotopes, and water quality to determine the pattern and spatial extent of groundwater flow: Nagaoka area, Japan. *Hydrogeology Journal 12:* 563–575.

Sandwidi, W.J.P., (2007). Groundwater potential to supply population demand within the Kompienga dam basin in Burkina Faso. Ph.D. Thesis. Ecology and Development Series, No. 54. Cuvillier Verlag Göttingen, p. 160.

Scanlon, B. R., Healy, R. W., and Cook, P. G. (2002). Choosing appropriate techniques for quantifying groundwater recharge. Hydrogeology *Journal*, *10*(2), 347. http://doi.org/Doi 10.1007/S10040-002-0200-1.

Setiawan, B., Abdurahman, O., Puspita, N., Riawan, E., Iman, MI. (2011). Climate Risk and Adaptation Assessment in The South Sumatera Province: Water Sector. Ministry of Environment, AusAID, and GIZ.

Shahin, M., (2002). Hydrology and Water Resources of Africa, Water Science and Technology Library. Kluwer Academic Publisher, Dordrecht, The Netherlands, p. 659.

Sharda VN, Kurothe RS, Sena DR, Pande VC, Tiwari SP. (2006). Estimation of groundwater recharge from water storage structures in a semi-arid climate of India. *Journal of Hydrology 329*: 224–243.

Sharma, M. L. (1986). Measurement and prediction of natural groundwater recharge- An overview. *Journal of Hydrology*, 25(1).

Sharma, M.L., Hughes, M.W., (1985). Groundwater recharge estimation using chloride, deuterium, and oxygen-18 profiles in the deep coastal sands of western Australia. *Journal of Hydrology* 81: 93–109.

Shirahatti, M. S. Ranghswami, M. V., Manjunath, M. V., Sivasamy, R. and Santana, S. B., (2012). Using groundwater levels to estimate the recharge: a case study from hard rock area of South India. India Water Week 2012 – *Water, Energy, and Food Security*: 10-14.

Simmers, I., (1997). Recharge of phreatic aquifers in (semi-) arid areas. *International Association of Hydrogeologists (IAH) 19. A.A.* Belkema, Rotterdam. 277 pp.

Singh, J., Wapakaia, W.W. and Chebosi, P.K., (1984). Estimating Groundwater Recharge Based on the Infiltration Characteristics of Layered Soil. *IAHS.144*: 37-45.

Smerdon, B.D., (2017). A synopsis of climate change effects on groundwater recharge. *J. Hydrol.* 555, 125–128. https://doi.org/10.1016/j.jhydrol.2017.09.047.

Somaratne, N. and Smettem, K. R. J. (2014). Theory of the generalized chloride mass balance method for recharge estimation in groundwater basins characterized by point and diffuse recharge. *Hydrology and Earth System Sciences Discussions*. *11*: 307–332.

Sophocleous, M. (2004). Groundwater Recharge, in Groundwater. In and E. J. U. Luis Silveira, Stefan Wohnlich (Ed.), Encyclopedia of Life Support Systems (EOLSS), Developed under the Auspices of the UNESCO. Oxford, UK: Eolss Publishers.

Sophocleous, M. (2005). Groundwater recharge and sustainability in the High Plains aquifer in Kansas, USA. *Hydrogeology, 13*: 351–365. DOI 10.1007/ s10040-004-0385-6.

Sophocleous, M. and Perry, C. A. (1985). Experimental studies in natural groundwater – recharge dynamics: the analysis of observed recharge events. *J. Hydrol 8:* 297–332.

Sun, X, Xu, Y., Jovanovic, N.Z., Kapangaziwiri, E, Brendonck, L., Bugan RDH., (2013). Application of the rainfall infiltration breakthrough (RIB) model for groundwater recharge estimation in west coastal South Africa. *J Water S Afr 39*(2):221–230.

Szilagyi J., Zlotnik V.A., Gates. J.B., Jozsa J. (2011): Mapping mean annual groundwater recharge in the Nebraska Sand Hills, USA. *J Hydrogeol*. doi:10.1007/s10040-011-0769-3.

Tamura, Y., Sato, T., Ooe, M., Ishiguro, M., (1991). A procedure for tidal analysis with a Bayesian information criterion. Geophys. *J. Int*. 104 507–516.

Taylor R. G. and K.W.F. Howard. (1996). Groundwater Recharge in the Victoria Nile Basin of East Africa: Support for the Soil Moisture Balance Approach Using Stable Isotope Tracers and Flow Modeling. *Journal of Hydrology v180* p 31-53.

Taylor, R.G., Scanlon, B., Doll, P., Rodell, M., (2013). Groundwater and climate change. *Nat. Clim. Change 3* (4), 322–329. https://doi.org/10.1038/Nclimate1744.

Ting, C.S., Kerh, T. and Liao, C.J. (1998). Estimation of groundwater recharge using the chloride mass-balance method, Pingtung Plain, Taiwan. *Hydrogeology Journal. 6*: 282–292.

Toll, N.J., and Rasmussen, T.C., (2007) Removal of barometric pressure effects and earth tides from observed water levels. *Ground Water 45* 101–105.

Tshelane, P. F., Ramoba, M. E., Mutheiwana, S. C., Tleane, D. A., & Seakamela, T. P. (2014). Quarterly status report on groundwater level trends—Limpopo Region: Directorate: Water regulation and use, Department of Water and Sanitation.

Van der Sommen, J.J., Geirnaert, W., 1988. On the continuity of aquifer systems on the crystalline basement of Burkina Faso. In: Simmers, I. (Ed.), Estimation of Natural Groundwater Recharge. *Reidel Publishing Company*, Dordrecht, Netherlands, pp. 29–45.

Varni, M., Comas, R., Weinzettel, P., Dietrich, S., (2013). Application of the water table fluctuation method to characterize groundwater recharge in the Pampa plain, Argentina.

Von, A, J., Mass, K., Bakker, M and Petersen, J (2008) Modeling time series of groundwater head fluctuations subjected to multiple stresses. *Ground Water* 46 30–40.

Walton, W.C., (1970). Ground Water Resource Evaluation. McGraw- Hill, New York, 664 p.

Wood, W., Sanford, W., (1995). Chemical and isotopic methods for quantifying groundwater recharge in a regional, semiarid environment. *Ground Water* 33: 458–468.

Wu, Q., Wang, G., Zhang, W., Cui, H., and Zhang, W. (2016). Estimation of groundwater recharge using tracers and numerical modeling in the North China Plain. *Water (Switzerland). 8:* 1–19. doi:10.3390/w8080353.

Wu, W-Y., Lo, M-H., Wada, Y., Famiglietti, J.S., Reager, J.T., Yeh, P.J.-F., Ducharne, A., Yang, Z-L., (2020). Divergent effects of climate change on future groundwater availability in key mid-latitude aquifers: *Nature Communications* *11*:3710. https://doi.org/10.1038/s41467-020-17581-y.

Wu. J, Zhang R., Yang J: (1996) Analysis of rainfall–recharge relationships. *J Hydrol 177*:143–160.

Xu Y, van Tonder GJ (2001) Estimation of recharge using a revised CRD method. *J Water S Afr 27*(3):341–344.

Xu, Y., Beekman, H.E., (2003). Groundwater estimation in Southern Africa.UNESCO ISHP Series No. 64, UNESCO Paris.

Yidana, S. M. (2013). The Stable Isotope Characteristics of Groundwater in the Voltaian Basin – An Evaluation of the Role of Meteoric Recharge in the Basin. *Hydrogeology and Hydrologic Engineering*, *2*(2), 1–3. http://doi.org/10.4172/2325-9647.1000106.

Yidana, S. M., and Koffie, E. (2014). The groundwater recharge regime of some slightly metamorphosed Neoproterozoic sedimentary rocks: an application of natural environmental tracers. *Hydrological Processes, 14*.

Yidana, S. M., Fynn, O. F., Chegbeleh, L. P., Loh, Y., and Obeng, M. A. (2014). Analysis of recharge and groundwater flow in parts of a weathered aquifer system in Northern Ghana. *Journal of Applied Water Engineering and Research*, *2*(2), 91–104. http://doi.org/10.1080/23249676.2014.954009.

Zagana, F., Obeidat, M., Kuells, C.h., Udluft, P., (2007). Chloride, hydrochemical and isotope methods of groundwater recharge estimation in the eastern Mediterranean area: a case study in Jordan. *Hydrol. Process. 130*:2112-2123.

Zhu, C., Winterle, J. R., and Love, E. I. (2003). Late Pleistocene and Holocene groundwater recharge from the chloride mass balance method and chlorine-36 data. *Water Resources Research*,39(7), 1182.

APPENDIX

Appendix 1: Time series of rainfall data and its corresponding hydraulic heads

Month	Rainfall (mm)	Hydraulic Head (m)
Jan-12	13.3	176.799
Feb-12	62.0	176.629
Mar-12	34.5	176.779
Apr-12	165.8	176.799
May-12	150.5	177.029
Jun-12	250.3	
Jul-12	49.3	177.489
Aug-12	55.3	177.109
Sep-12	215.1	177.079
Oct-12	140.3	177.429
Nov-12	45.4	177.404
Dec-12	74.3	177.364
Jan-13	0.0	176.939
Feb-13	5.5	176.559
Mar-13	134.0	176.609
Apr-13	69.5	176.429
May-13	238.0	
Jun-13	212.0	177.639
Jul-13	90.3	177.194
Aug-13	16.0	177.049
Sep-13	215.5	177.019
Oct-13	165.8	177.419
Nov-13	48.0	177.739
Dec-13	16.3	177.199
Jan-14	66.3	176.979
Feb-14	41.0	176.749
Mar-14	119.5	176.729
Apr-14	117.8	176.569
May-14	162.3	176.869
Jun-14	346.0	177.439
Jul-14	137.0	177.609
Aug-14	114.0	177.419
Sep-14	159.3	177.549
Oct-14	138.8	177.839
Nov-14	80.3	177.649
Dec-14	3.3	177.399

Jan-15	52.5	177.129
Feb-15	73.3	177.329
Mar-15	179.0	177.069
Apr-15	71.8	177.269
May-15	222.0	177.459
Jun-15	177.5	177.449
Jul-15	92.3	177.449
Aug-15	44.0	177.419
Sep-15	33.5	177.379
Oct-15	175.8	177.439
Nov-15	141.8	177.599
Dec-15	0.0	177.609
Jan-16	68.0	177.269
Feb-16	6.8	177.049
Mar-16	211.5	176.999
Apr-16	116.8	177.439
May-16	120.8	177.459
Jun-16	177.8	177.619
Jul-16	86.5	177.529
Aug-16	92.3	177.539
Sep-16	168.3	177.669
Oct-16	291.0	177.719
Nov-16	56.5	177.939
Dec-16	46.8	177.779
Jan-17	30.5	177.659
Feb-17	3.3	177.509
Mar-17	70.5	177.309
Apr-17	261.5	177.339
May-17	219.0	177.709
Jun-17	180.0	177.789

Month	Rainfall (mm)	Hydraulic Head (m)
Sep-12	215.1	197.357
Oct-12	140.3	197.477
Nov-12	45.4	197.607
Dec-12	74.3	197.547
Jan-13	0.0	197.337
Feb-13	5.5	197.057
Mar-13	134.0	195.227
Apr-13	69.5	194.287
May-13	238.0	
Jun-13	212.0	194.747
Jul-13	90.3	193.467
Aug-13	16.0	194.047
Sep-13	215.5	192.727
Oct-13	165.8	194.497
Nov-13	48.0	194.517
Dec-13	16.3	195.377
Jan-14	66.3	195.697
Feb-14	41.0	195.567
Mar-14	119.5	194.117
Apr-14	117.8	192.687
May-14	162.3	192.527
Jun-14	346.0	193.057
Jul-14	137.0	193.427
Aug-14	114.0	192.447
Sep-14	159.3	192.417
Oct-14	138.8	192.447
Nov-14	80.3	192.347
Dec-14	3.3	192.077
Jan-15	52.5	191.737
Feb-15	73.3	191.437
Mar-15	179.0	191.247
Apr-15	71.8	191.517
May-15	222.0	192.117
Jun-15	177.5	191.797
Jul-15	92.3	192.687
Aug-15	44.0	193.647
Sep-15	33.5	193.117
Oct-15	175.8	193.047
Nov-15	141.8	194.087
Dec-15	0.0	194.347

Jan-16	68.0	193.627
Feb-16	6.8	193.227
Mar-16	211.5	192.387
Apr-16	116.8	192.107
May-16	120.8	192.077
Jun-16	177.8	191.957
Jul-16	86.5	191.887
Aug-16	92.3	191.807
Sep-16	168.3	191.937
Oct-16	291.0	192.437
Nov-16	56.5	194.077
Dec-16	46.8	194.477
Jan-17	30.5	194.587
Feb-17	3.3	194.497
Mar-17	70.5	194.327
Apr-17	261.5	194.107
May-17	219.0	194.477
Jun-17	180.0	194.887

Month	Rainfall (mm)	Hydraulic Head (m)
Nov-11	39.3	169.617
Dec-11	0.0	169.487
Jan-12	13.3	169.297
Feb-12	62.0	169.177
Mar-12	34.5	169.247
Apr-12	165.8	169.157
May-12	150.5	169.157
Jun-12	250.3	
Jul-12	49.3	169.647
Aug-12	55.3	169.437
Sep-12	215.1	169.427
Oct-12	140.3	169.807
Nov-12	45.4	169.552
Dec-12	74.3	169.577
Jan-13	0.0	169.367
Feb-13	5.5	169.147
Mar-13	134.0	169.257
Apr-13	69.5	169.087
May-13	238.0	
Jun-13	212.0	
Jul-13	90.3	169.392
Aug-13	16.0	169.277
Sep-13	215.5	169.467
Oct-13	165.8	169.702
Nov-13	48.0	169.617
Dec-13	16.3	169.467
Jan-14	66.3	169.257
Feb-14	41.0	169.062
Mar-14	119.5	169.177
Apr-14	117.8	168.987
May-14	162.3	169.257
Jun-14	346.0	169.727
Jul-14	137.0	169.997
Aug-14	114.0	169.617
Sep-14	159.3	169.647
Oct-14	138.8	169.727
Nov-14	80.3	169.677
Dec-14	3.3	169.537
Jan-15	52.5	169.357

Feb-15	73.3	169.297
Mar-15	179.0	169.267
Apr-15	71.8	169.367
May-15	222.0	169.607
Jun-15	177.5	169.517
Jul-15	92.3	169.547
Aug-15	44.0	169.457
Sep-15	33.5	169.267
Oct-15	175.8	169.337
Nov-15	141.8	169.617
Dec-15	0.0	169.357
Jan-16	68.0	169.137
Feb-16	6.8	168.757
Mar-16	211.5	169.037
Apr-16	116.8	169.307
May-16	120.8	169.367
Jun-16	177.8	169.497
Jul-16	86.5	169.417
Aug-16	92.3	169.387
Sep-16	168.3	169.497
Oct-16	291.0	169.887
Nov-16	56.5	169.867
Dec-16	46.8	169.617
Jan-17	30.5	169.147
Feb-17	3.3	169.327
Mar-17	70.5	169.137
Apr-17	261.5	169.237
May-17	219.0	169.537
Jun-17	180.0	169.847

Month	Rainfall (mm)	Hydraulic Head (m)
Nov-11	39.3	178.136
Dec-11	0.0	178.166
Jan-12	13.3	178.316
Feb-12	62.0	178.086
Mar-12	34.5	177.996
Apr-12	165.8	177.926
May-12	150.5	177.906
Jun-12	250.3	
Jul-12	49.3	177.866
Aug-12	55.3	177.931
Sep-12	215.1	177.966
Oct-12	140.3	177.976
Nov-12	45.4	178.326
Dec-12	74.3	178.336
Jan-13	0.0	178.356
Feb-13	5.5	178.306
Mar-13	134.0	178.226
Apr-13	69.5	178.136
May-13	238.0	
Jun-13	212.0	178.006
Jul-13	90.3	178.056
Aug-13	16.0	178.076
Sep-13	215.5	178.086
Oct-13	165.8	178.226
Nov-13	48.0	178.336
Dec-13	16.3	178.416
Jan-14	66.3	178.506
Feb-14	41.0	178.406
Mar-14	119.5	178.326
Apr-14	117.8	178.256
May-14	162.3	178.086
Jun-14	346.0	178.286
Jul-14	137.0	178.266
Aug-14	114.0	178.436
Sep-14	159.3	178.176
Oct-14	138.8	178.716
Nov-14	80.3	178.766
Dec-14	3.3	178.806
Jan-15	52.5	178.806
Feb-15	73.3	178.726
Mar-15	179.0	178.666

Apr-15	71.8	178.516
May-15	222.0	178.436
Jun-15	177.5	178.426
Jul-15	92.3	178.406
Aug-15	44.0	178.406
Sep-15	33.5	178.346
Oct-15	175.8	178.286
Nov-15	141.8	178.326
Dec-15	0.0	178.346
Jan-16	68.0	178.286
Feb-16	6.8	178.181
Mar-16	211.5	178.016
Apr-16	116.8	178.106
May-16	120.8	177.846
Jun-16	177.8	
Jul-16	86.5	177.946
Aug-16	92.3	177.936
Sep-16	168.3	177.936
Oct-16	291.0	178.136
Nov-16	56.5	178.636
Dec-16	46.8	178.686
Jan-17	30.5	178.696
Feb-17	3.3	178.676
Mar-17	70.5	178.596
Apr-17	261.5	178.456
May-17	219.0	178.446
Jun-17	180.0	178.826

Month	Rainfall (mm)	Hydraulic Head (m)
Nov-11	39.3	193.951
Dec-11	0.0	193.811
Jan-12	13.3	192.311
Feb-12	62.0	193.371
Mar-12	34.5	193.201
Apr-12	165.8	192.221
May-12	150.5	192.986
Jun-12	250.3	
Jul-12	49.3	193.111
Aug-12	55.3	193.101
Sep-12	215.1	193.011
Oct-12	140.3	193.091
Nov-12	45.4	193.511
Dec-12	74.3	193.151
Jan-13	0.0	192.981
Feb-13	5.5	192.801
Mar-13	134.0	192.631
Apr-13	69.5	192.471
May-13	238.0	
Jun-13	212.0	192.561
Jul-13	90.3	192.611
Aug-13	16.0	192.591
Sep-13	215.5	192.496
Oct-13	165.8	192.856
Nov-13	48.0	192.721
Dec-13	16.3	192.701
Jan-14	66.3	192.561
Feb-14	41.0	192.321
Mar-14	119.5	192.231
Apr-14	117.8	192.051
May-14	162.3	192.031
Jun-14	346.0	192.371
Jul-14	137.0	192.491
Aug-14	114.0	192.551
Sep-14	159.3	192.631
Oct-14	138.8	192.781
Nov-14	80.3	192.871
Dec-14	3.3	192.821
Jan-15	52.5	192.871
Feb-15	73.3	192.391
Mar-15	179.0	192.301

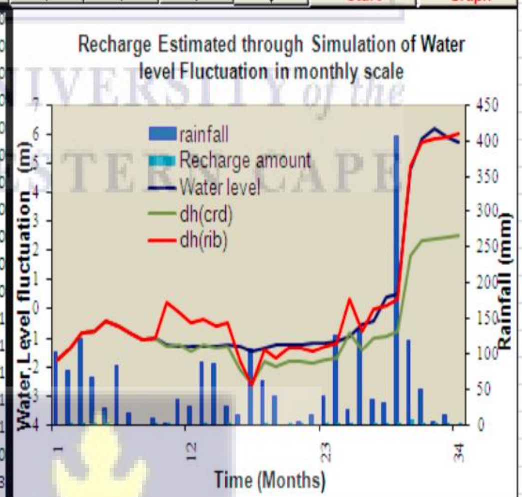
Apr-15	71.8	192.181
May-15	222.0	192.261
Jun-15	177.5	192.401
Jul-15	92.3	192.441
Aug-15	44.0	192.561
Sep-15	33.5	192.431
Oct-15	175.8	192.331
Nov-15	141.8	192.531
Dec-15	0.0	192.561
Jan-16	68.0	192.681
Feb-16	6.8	192.271
Mar-16	211.5	192.051
Apr-16	116.8	192.051
May-16	120.8	192.161
Jun-16	177.8	192.001
Jul-16	86.5	192.121
Aug-16	92.3	192.111
Sep-16	168.3	192.071
Oct-16	291.0	192.321
Nov-16	56.5	192.951
Dec-16	46.8	193.031
Jan-17	30.5	192.981
Feb-17	3.3	192.761
Mar-17	70.5	192.541
Apr-17	261.5	192.311
May-17	219.0	192.941
Jun-17	180.0	192.971

Appendix 2: The user interface of RIB program

RIB interface.XLS [Compatibility Mode] - Microsoft Excel

	A	B	C	D	E	F	G	H	I	J	K	L	M	N	O	P
1	RIB	Rav	$Q_{other}(m^3/m)$	$Q_{pumpage}(m^3/m)$	$WL_{av}(m)$	K_{crd}	R_{rf}	Re(%)	r (%)	S_y	A(km²)	$Q_{out}(m^3/d)$	lag_months	Length_M	Sum_Min	Sum_Min (RIB)
2	34	58.66	#DIV/0!	4426.76	0.00	1.00	41.67	1.2397	1.67	1.90E-03	21.00	0.00E+00	-1	12	38.34131516	5.169014204
3															Start	Graph

	A	B	C	D	E	F	G	H	I	J
3	Monthi	rain(mm)	Q_{other}	$Q_{pumpage}$	dh_obs	dh(crd)	dh(rib)	Re(rib)	0.00E+00	
4	1	103.1		3900	-1.75	-1.75	-1.75	0.00	0.00E+00	0.00E+00
5	2	75.5		3780	-1.36	-1.36	-1.36	0.75	0.00E+00	0.00E+00
6	3	120.7		4380	-0.84	-0.84	-0.84	0.97	0.00E+00	0.00E+00
7	4	66.5		3690	-0.76	-0.76	-0.76	0.16	0.00E+00	0.00E+00
8	5	24.3		3870	-0.42	-0.42	-0.42	0.64	0.00E+00	0.00E+00
9	6	84.5		4080	-0.61	-0.61	-0.61	0.00	0.00E+00	0.00E+00
10	7	17.5		4920	-0.83	-0.83	-0.83	0.00	0.00E+00	0.00E+00
11	8	0		4530	-1.03	-1.03	-1.03	0.00	0.00E+00	0.00E+00
12	9	10.2		4920	-1.00	-1.00	-1.00	0.04	0.00E+00	0.00E+00
13	10	2.7		3630	-1.23	-1.28	0.25	2.39	1.90E-03	2.20E+00
14	11	35.1		4680	-1.26	-1.26	-0.11	0.00	2.42E-05	1.32E+00
15	12	25.8		5250	-1.28	-1.47	-0.50	0.00	3.32E-02	6.22E-01
16	13	89.2		3810	-1.28	-1.23	-0.38	0.22	2.50E-03	8.17E-01
17	14	86		3450	-1.29	-1.35	-0.58	0.00	3.53E-03	5.11E-01
18	15	26.2		4710	-1.22	-1.27	-0.45	0.24	2.35E-03	5.87E-01
19	16	15.3		3300	-1.31	-2.09	-1.84	0.00	6.15E-01	2.83E-01
20	17	108.4		4710	-1.46	-2.55	-2.64	0.00	1.18E+00	1.40E+00
21	18	61.8		6060	-1.34	-1.82	-1.41	2.34	2.29E-01	5.80E-03
22	19	41.3		4050	-1.23	-2.01	-1.70	0.00	6.01E-01	2.25E-01
23	20	0		5160	-1.23	-1.80	-1.37	0.63	3.21E-01	1.93E-02
24	21	4.4		4560	-1.26	-1.80	-1.36	0.03	2.93E-01	1.04E-02
25	22	14.7		5100	-1.15	-1.85	-1.46	0.00	4.98E-01	9.78E-02
26	23	39.4		4560	-1.16	-1.74	-1.27	0.37	3.35E-01	1.02E-02
27	24	126		4080	-1.14	-1.70	-1.19	0.15	3.23E-01	2.82E-03
28	25	21.1		2970	-0.96	-0.82	0.35	2.92	1.92E-02	1.69E+00
29	26	133		5340	-0.51	-1.43	-0.74	0.00	8.34E-01	4.97E-02
30	27	36.4		3900	-0.41	-1.01	0.01	1.41	3.51E-01	1.76E-01
31	28	31.4		4410	0.38	-0.92	0.15	0.27	1.70E+00	5.58E-02
32	29	405.5		5550	0.51	-0.78	0.36	0.41	1.66E+00	2.12E-02
33	30	118.7		2970	4.78	1.84	4.89	8.61	8.66E+00	1.20E-02



Appendix 3: Input/output data in the rainfall infiltration breakthrough (RIB) model

Input/ output	Symbol	Units	Definition	Type of data
Input	Day_i	-	Days of simulation	Time series (from 1 to *n*)
	rain	mm	Daily rainfall	Time series
	Q_{oth}	m^3 per time[a]	Source/sink of water other than through abstraction	Time series (positive value means groundwater sink, negative value means groundwater source)
	Q_p	m^3 per time[a]	Abstraction of groundwater	Time series
	dh_obs	m	Observed water level fluctuations (current WL– WL_{AV})	Time series
	S_y	-	Specific yield	One value
	A	km^2	Surface area of watershed	One value
	Q_{out}	$m^3\ d^{-1}$	Constant volume of groundwater sink (e.g. baseflow)	One value
	lag_D	time[a]	Time delay between rainfall events and recharge	One value
	Length_D	time[a]	Parameter that characterises rain sequences and antecedent conditions	One value
Output	R_{AV}	mm per time[a]	Average rainfall	One value
	WLF_{AV}	m	Average groundwater level fluctuations	One value
	K_{crd}	-	Parameter of the CRD[b] method	One value
	$R_{ref}(P_t)$	Mm per time[a]	Threshold value representing aquifer boundary conditions	One value corresponding to P_t (ranging from 0 for a closed aquifer to R_{AV} for an open aquifer system)
	r	-	Parameter of the RIB[c] method	One value
	Re(%)	-	Ratio of recharge to rainfall	One value
	dh(crd)	m	Calculated WLF with the CRD[b] method	Time series
	dh(rib)	m	Calculated WLF with the RIB[c] method	Time series
	Re(rib)	mm per time[a]	Calculated recharge	Time series

[a] *Day, month or year*
[b] *Cumulative rainfall departure method*
[c] *Rainfall infiltration breakthrough method*

*1 month at monthly basis is equal to 30/31 days at daily basis