MODELLING GROUNDWATER RECHARGE ESTIMATION USING MODIFIED SOIL MOISTURE BALANCE APPROACH IN OTUKPO BASIN

BY

ADAUDU, Ignatius Idoko
MEng/SEET/2017/6834

DEPARTMENT OF CIVIL ENGINEERING
FEDERAL UNIVERSITY OF TECHNOLOGY, MINNA

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ABSTRACT

In this study, groundwater recharge in Otukpo basin has been estimated using a modified daily soil moisture balance based on a single soil water store for a climate classified as tropical with distinct dry and wet seasons in the Middle Belt part of Nigeria. Soil properties like field capacity, permanent wilting point, readily available water, actual and potential evapotranspiration, soil moisture deficit were all estimated and deployed in the model which algorithm was developed using Python programming language, hence the name modified soil moisture balance model. Runoff is estimated using runoff matrix and runoff coefficients which depend on rainfall intensity and soil moisture deficits. A new component, near surface storage, is used to represent continuing evapotranspiration on days following heavy rainfall even though the soil moisture deficit is high. Groundwater recharge is estimated for cassava and yam which are commonly cultivated vegetable crops in the study area. Meteorological data for the periods of 2008 to 2018 were used in the model analysis. The model recorded annual groundwater recharge which varied from 333.35 mm in 2009 water year (just 20.01% of annual rainfall for the year) to 38.119 mm in 2017 water year which is 3.6% of annual rainfall for the year). The highest annual rainfall depth was also observed in the year 2009 as 1665.4 mm, with the lowest annual rainfall depth, 1062.4 mm also observed in the year 2017. The annual runoff ranged from 322.04 mm in the year 2015, a 32.16 % of annual rainfall for the year to 935.56 mm in the year 2008 a 58.17 % of annual rainfall for the year. The lowest actual evapotranspiration AE was also observed in 2017 as against the highest in 2012. The AE ranged from 583.84 mm in 2017 to 721.39 mm in 2012. The model gave a simplified method of groundwater recharge estimation as well as runoff depth coupled with rainfall-runoff relationship.

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LIST OF ABBREVIATIONS AND SYMBOLS

TDR	Time Domain Reflectory
WTD	Water Table Fluctuation
DHB	Distributed Hydrological Budget
HB	Hydrological Budget
DW	Change In Social Moisture Storage
W_1	Weight of Container
W_2	Weight of Container with Wet Soil
W_3	Weight of Container with Dry Soil
δ_b	Bulk Density
θ_{fc}	Field Capacity
θ_{wp}	Permanent Wilting Point
RD	Depth of Root Zone
ETo	Evapotranspiration
T_{mean}	Mean Temperature
T_{max}	Maximum Temperature
T_{min}	Minimum Temperature
Ra	Extra Terrestrial Solar Radiation
MSMB	Modified Soil Moisture Balance Model
SMD	Soil Moisture Deficit
AWE	Available Water
Zr	Maximum Root Depth
TAW	Total Available Water
NSS	Near Surface Storage
FC	Field Capacity
WP	Wilting Point

PE	Potential Evapotraspiration
AE	Actual Evapotranspiration
Kc	Crop Coefficient
FAO	Food and Agriculture Organization
NAF	Nigerian Airforce
NIMET	Nigerian Meteorological Agency
TAC	Tactical Air Command

CHAPTER ONE

1.0 INTRODUCTION

1.1 Background to the Study

Groundwater recharge is defined as water that infiltrates through the sub surface to the zone of saturation beneath the water table (Reese and Risser, 2010). It results in the increase of ground water storage and contributes to groundwater flow (Idowu, 2010). Groundwater recharge is a hydrologic process, where water moves downward from surface water to groundwater.

Groundwater is the primary source of water for domestic and agricultural water supplies throughout the tropics and much of sub-Saharan Africa (Doll *et al*., 2012) Efforts to meet projected increase in freshwater demand over the next few decades across sub-Saharan Africa depend on the development of the groundwater resource which in many environments is the only perennial source of freshwater. Groundwater is the capital source of freshwater for nearly half of earth’s population for irrigation and domestic water needs (Kunkel and Wendland, 2002). Groundwater is identified as a renewable water resource for supporting agricultural, industrial, environmental and municipal domestic water demands .The estimatio[illegible]ound water recharge is the key to understanding the groundwater reservoir and forecasting it’s potential accessibility and sustainability even though other elements have to be taken into accounts for example ,social, economic and hydrogeological considerations (Bogena, 2005).

Recharge is the primary method through which water enters an aquifer. This process usually occurs in the Vadoze zone, below plant roots and is often expressed as a flux to the water table surface.

According to Najjar (1999) groundwater recharge also encompasses water moving away from the water table farther into the saturates zone. Recharge occurs both naturally through the water cycle and through anthropogenic processes in other words, artificial groundwater recharge where rain water and or reclaimed water is rooted to the subsurface.

Groundwater recharge happens when a part of precipitation on the ground surface infiltrates through the soil and the reaches the water table. Groundwater recharge can be known as water moving from the land surface to the unsaturated zone. When water reaches the water table, it can go out of the ground water to the surface water which is called discharge. The amount of recharge in humid region is usually high because the region receives large amount of rainfall, have favourable surface conditions for infiltration and a less susceptible to the influences of high temperatures and evapotranspiration (Reese and Risser, 2010). For example Azeez (1972) reported that a substantial rate of groundwater recharge occurs in the regolith overburden in the basement complex of Southwestern Nigeria. The estimation of the rate of natural groundwater recharge is a pre-requisite for efficient groundwater resource management (Kumar and Seethapathi, 2002). Hence, recharge estimate is peculiarly important in regions with large demands for groundwater supplies, where such resources are the key to economic development. While the estimation of recharge by whatever method is normally subject to large uncertainties and errors.

For many years, hydrologists have been trying to estimate natural recharge rates to aquifer system. In order to estimate the potential long term safe yield of an aquifer system however is related not so much to the proportion of the discharge that groundwater extraction centers are able to capture.

Groundwater recharge is an important component of the water balance and evaluation of groundwater resources largely depends on it (Dages *et al*., 2009). This estimation of the rate of natural ground water recharge is a basic pre-requisite for efficient groundwater resources management strategy that will ensure the protection of groundwater resources not only from climate change, but also from other stresses (Hennon, 2005).

Water movements in top soils determine the rate of recharge to the groundwater rate of plant transpiration, soil evaporation and run-off (Walker, 2002). Therefore an accurate description of unsaturated zone water movement and accurate methods for determination of parameter and input data are essentials to derive proper estimate of ground water recharge. Practicing hydrologists, typically make the best estimates of recharge possible by the use of methods that are relatively straight forward in their application and require only commonly available hydrologic data. Arnold (2007) reported that groundwater recharge is the process by which water percolates down the soil and reaches the water table either by natural or artificial methods to replenish the aquifer with water table from the land surface. Most of the time groundwater is determined to a large extent on an imbalance at the land surface between precipitation and evaporative demand. In arid and semi-arid regions, the search for water which are under increasing stress from the growing human population poses a great challenge due to its scarcity (Corpo, 2010).

Groundwater as a dynamic system is located beneath the earth surface and moves under the control of many factors which are influenced by forces that are dependent on hydrogeology, hydrology and climatology. Recharge as one of the factors controlling the situation and fluctuates of groundwater is an important parameter that needs to be assessed more fully. Recharge occurring in small and large scales spatially and temporally is influenced by several factors such as meteorology soil characteristics, geology, surface cover, slope and depth of the groundwater level. Groundwater recharge estimation from precipitation is an integral part of hydrology and hydrogeology. Although precipitation is the most important source of groundwater recharge, the accuracy of currently attainable techniques for measuring recharge are not completely acceptable (Adeleke *et al.*, 2015).

In measuring groundwater recharge, indirect methods are employed since there are no universally accepted standard methods. However methods for estimating groundwater recharges have been classified into three groups namely: physical model, obtained when the recharge is calculated from the base flow; chemical model method, which is used when the measurements of water soluble substances are considered; and numerical model methods such as HELP, RORA, PART, WEAP, WTF, PUISE and HYSEP, which can be used in the estimation of groundwater recharges for various climates such as in arid, semi-arid and tropical regions albeit each with its own advantages and disadvantages.

1.2 Statement of the Research Problem

Groundwater recharge is a fundamental importance to meet the rapidly increasing agricultural, industrial and domestic water supply requirement within the Otukpo basin. This resource is almost the only key to economic development in the area and hence the estimation of groundwater is a necessity for the efficient and sustainable groundwater resource managements. Gehrels (2000) concluded that the method of estimating actual evapotranspiration and charges in soil water storage determines the accuracy of the water balance. However due to lack of basic understanding of the spatial and variability of hydrological processes, water management is becoming a major challenge. The groundwater recharge estimation and causes of groundwater level fluctuations in the Otukpo basin are not well understood due to limited knowledge of the soil water flow through the thick unsaturated zone and of the actual evapotranspiration from the area.

1.3 Aim and Objectives of the Study

The aim of this study is to estimate the groundwater recharge of Otukpo basin using the modified soil moisture balanced model alongside the hydrologic data.

The specific objectives of this study are to;

i. Determine the relationships between hydrological parameters (Rainfall, Runoff, SMD, Recharge and ETo) in Otukpo Basin, and

ii. Estimate the groundwater recharge within Otukpo Basin using modified hydrological model

1.4 Justification of the Study

The role of groundwater, with recharge estimation as a critical parameter for determining its sustainable use is becoming increasingly important in the emerging integrated water resource management. A proper understanding of estimating recharge as a result of modeling is crucial to assessing groundwater availability efficiently. The study would provide a better understanding of groundwater recharge estimation in the Otukpo basin and would also provide detail of how much groundwater that is available.

1.5 Scope of the Study

The scope of this work was limited to the estimation of the groundwater recharge within Otukpo basin using modelling written, using Python Programming Language.11 years Rainfall data of the study area was used in the modelling (2008-2018). Hargreaves equation was used for estimating evapotranspiration ETo.

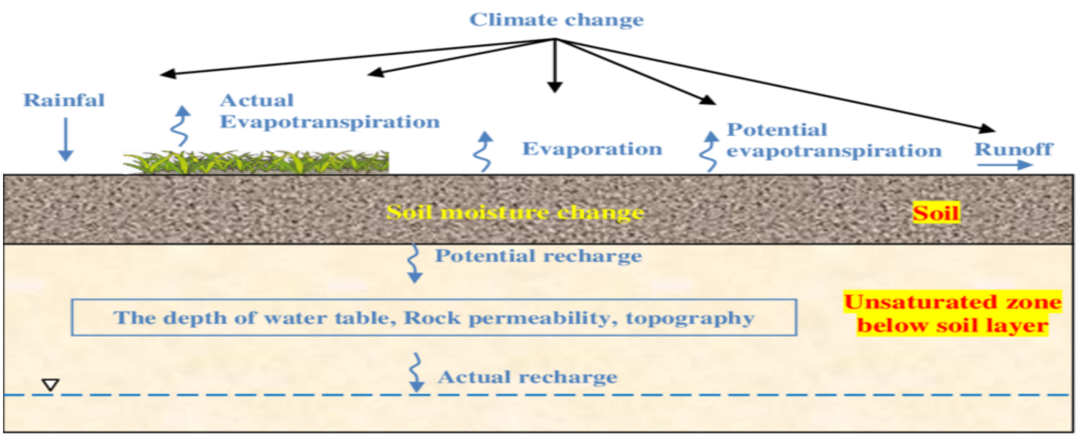
CHAPTER TWO

2.0 LITERATURE REVIEW

2.1 Concepts of Groundwater Recharge

Groundwater recharge can be defined as the amount of water added to the groundwater reservoir in excess of soil moisture deficits and evapotranspiration by direct percolation through the vadose zone. It is the resultant variable weather conditions, root water uptake, processes of soil water flow and vadose zone properties (Gehrels, 2000).

There are various sources of recharge to a groundwater system. Direct (precipitation) recharge in which water is added to the groundwater reservoir in excess to soil moisture deficit and evapotranspiration. Indirect recharge is that type where water percolates to, the water through the beds of surface water courses (Volt, 2000). Fulton (2007) puts it that, quantitative understanding of the process of groundwater recharge is fundamental



to the sustainable management of groundwater resources in such a way that the amount of recharge dictates the amount of water extracted sustainably from the aquifers. Recharge has a great importance to assess the impact of climate changes on groundwater resources and aquifer vulnerability to contaminants (Lox, 2012).

Figure 2.1: Ground water recharge processes
(Source: Eilers *et al.*, 2007)

Groundwater recharge is one of the most significant aspect of the hydrological system, though it is also the most difficult to determine (Nyagwabo, 2006). According to Xu and Beckham (2003) the downward flow of water through the unsaturated zone is the most significant mode of recharge in arid and semi-arid regions as upper Berg River catchments. Groundwater recharge occurs naturally from sources such as precipitation, rivers, canals, lakes as well as irrigation and urbanization as man induced activities (Lerner, 2000). This can be seen in Figure 2.1.

There are three mechanisms of natural groundwater recharge, such as direct, indirect and localized. According to Lloyd (1986), each type of ground water recharge mechanism is more prevalent in some climatic conditions than others.

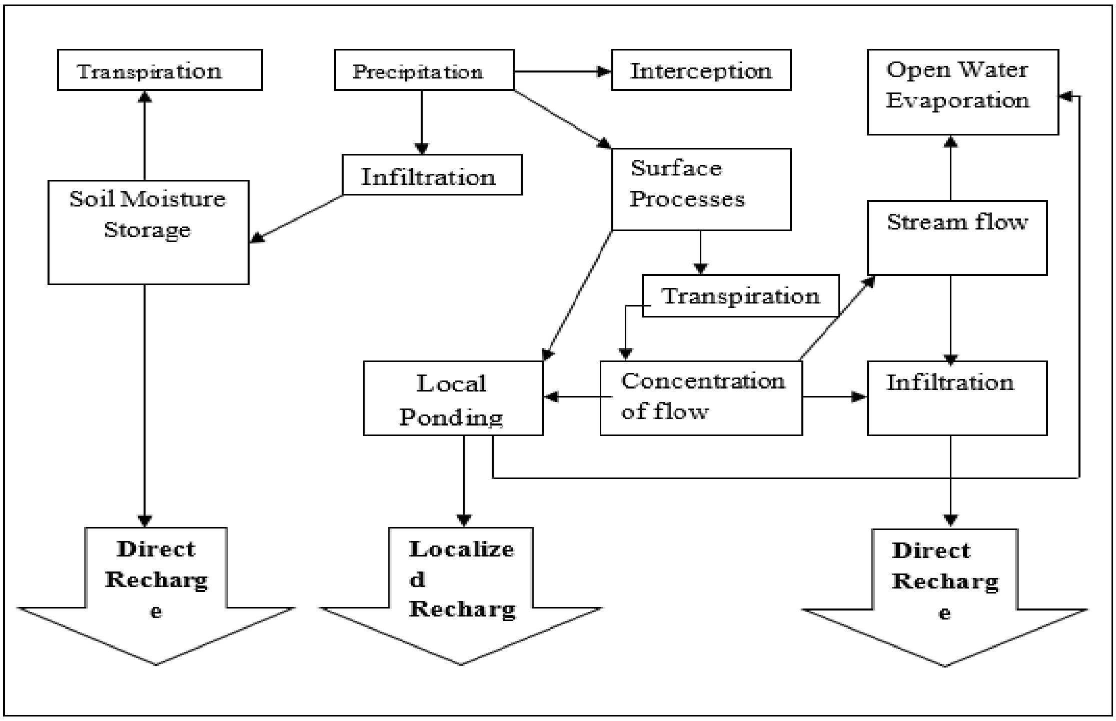


Figure 2.2: Elements of recharge in a semi-arid area
(Source: Xu and Beekman, 2019*)*

2.2 Factors Affecting Groundwater

Several factors control groundwater recharge, such as rate and duration of rainfall, the antecedent moisture, condition of the soil profile, geology, soil properties, the depth of water table and aquifer properties, vegetation and land use, topography and land form. The amount of recharge or infiltration at a specific site depends on the amount of precipitation evaporated back into the atmosphere, the amount of water transmitted from natural vegetation to the air, site topography and vegetation (Blaise, 2007). A simple illustration of this is shown in Figure 2.2

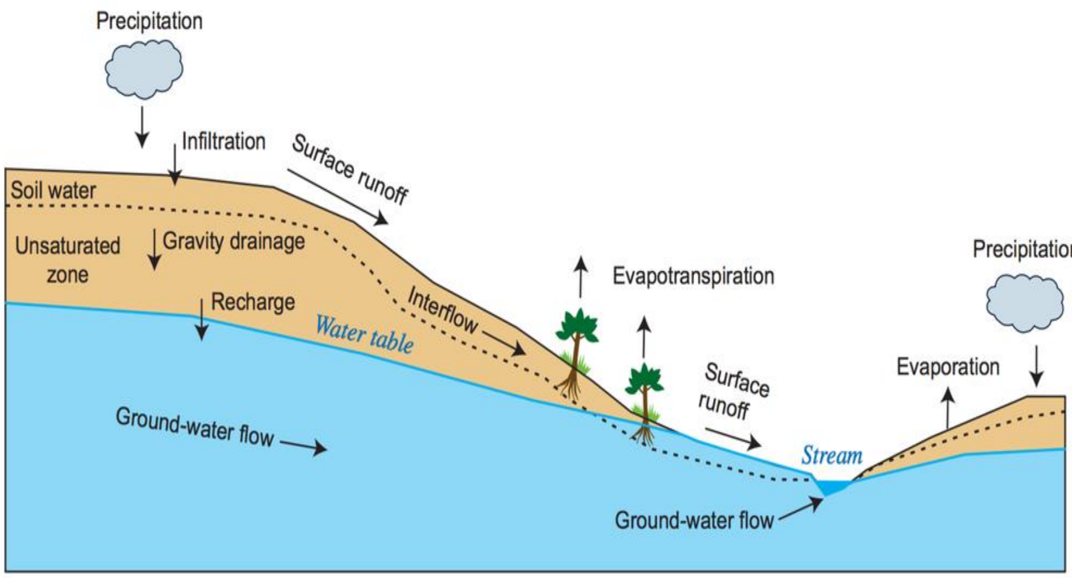


Figure 2.3: Hydrologic cycle
(Source: Najjar, 1999)

Vegetation cover influences the recharge processes. Albhansi (2012) observed that recharge has increased systematically with years after de-forestation in some regions of Africa.

2.2.1 Climate change and variability

Increased variability in precipitation and more extreme weather events caused by climate change can lead to longer periods or droughts and floods, which directly affects availability and dependency on groundwater. Global groundwater resources are threatened by consequences of climate change and human activities. Changes in global climate are expected to affect the hydrological cycle altering groundwater recharge to aquifers and surface water levels with other associated impacts on the ecosystem (Vogel, 1996). The hydrologic cycle is shown in Figure 2.3 over the next 100 years, the full impact that climate change is having on groundwater recharge will become apparent in half of the world's aquifers. Climatic variability and change influences groundwater systems both directly through replenishment by recharge and indirectly through changes in ground water use (McCallum *et al*., 2020). Understanding climate variability and change is vital for society and ecosystems, particularly with regards to complex changes affecting the availability and sustainability of groundwater resources (Dragoni and Sukhija, 2008). The potential effects of climate change variability and change on water resources are well recognized globally and have been identified as a major issue facing the availability of groundwater resources in the United States (Alley, 1999). Basically, climate change and variability affects groundwater recharge rate, depth to water table and water levels in aquifers (Chen *et al.,* 2004).

2.2.2 Rainfall variation

The rainfall is the main recharge of groundwater reservoirs. Any changes in the rainfall quantity and storm pattern can affect the recharge quantity since it has direct impact on the rate of infiltration. In the event of degrease in rainfall, the recharge also reduces (Blaise, 2007). Heavy intensity of rainfall causes more runoff and less infiltration reducing the recharge quantity and depletion (Bogena, 2005). Hanell (2011) observed that groundwater recharge with respect to rainfall varied as a result of difference in hydrologic soil properties, vegetation and rainfall intensity.

Groundwater recharge in a region depends mainly on the rainfall change during major recharge season. In temperate climates, an increase in rainfall is generally foreseen during the winter season, where most recharge occurs. The natural and artificial recharge of ground water is shown in Figure 2.4. During the hotter summer, however there might be increased evapotranspiration in particular if the groundwater table is close to the land surface (Brouyere *et al*., 2004). The sustainability of current and future groundwater abstraction relies on groundwater recharge but the conversion of rainfall into recharge remains poorly understood (Valerie, 2000).

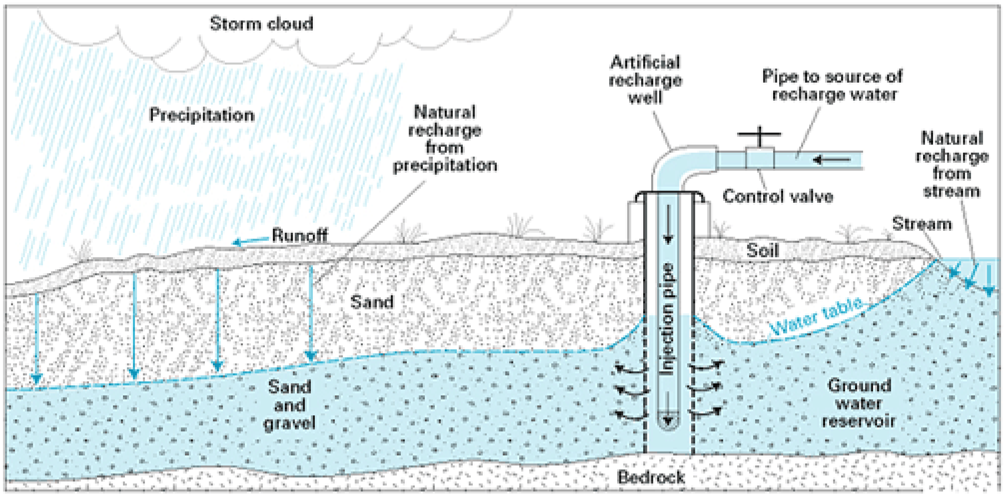


Figure 2.4: Natural and artificial recharge of groundwater
(Source: Harry, 2000)

2.2.3 Evapotranspiration

Evapotranspiration occurs in the zones where groundwater is too close to the surface. These zones are known as wetlands water may not be emanating or the terrain may not be wet, but the table is located high enough for plants to be consumed by groundwater.

2.2.4 Vegetation cover

Woody plant encroachment could alter soil infiltration rates, soil water storage, transpiration, interception and sub-surface pathways to affect groundwater recharge (Akiti, 2003). Strong correlation between groundwater, vegetation and atmosphere processes has been recognized in semi-arid and sub humid ecosystems where woodland encroachment has been wildly reported (Hoffman *et al*., 2011). This illustration can be seen in Figure 2.5.

Vegetation can potentially have a strong influence on the water budget and hydrological processes in vegetated ecosystems due to canopy rainfall interception and the partitioning of rainfall into through fall and stream flow as well as root water uptake from the vadose zone or groundwater table (Junilang, 2014). Vegetation exerts a strong control over the hydrological cycle including groundwater recharge which provides water for many human and natural communities (John *et al*., 2012). A sample of this can be found in Figure 2.6.

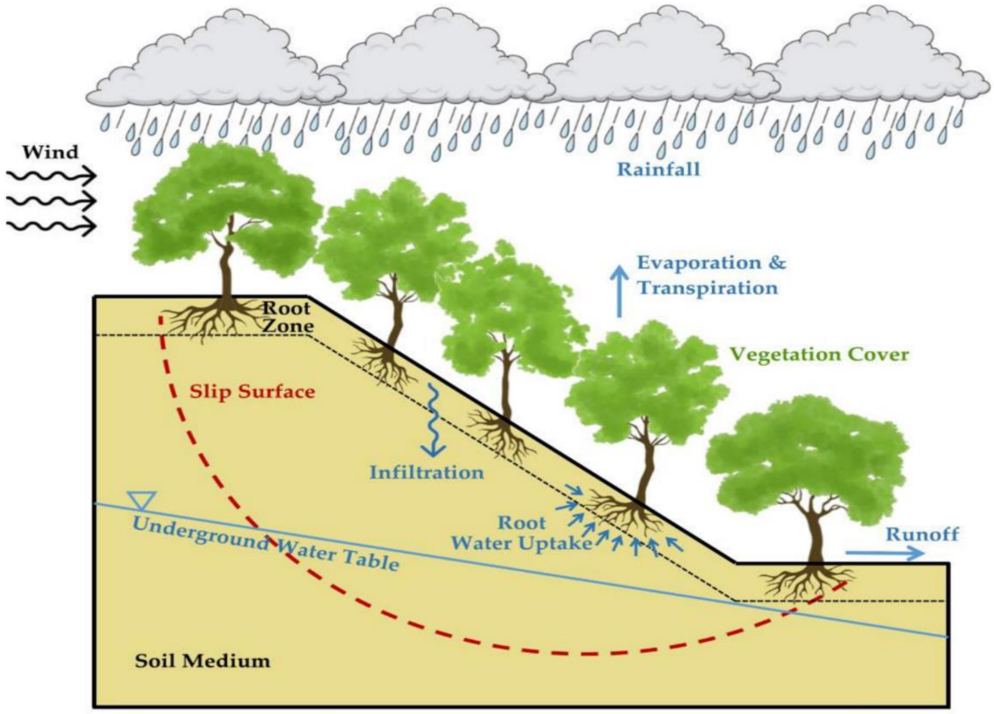


Figure 2.5: Schematic of vegetated slope stability analysis
*(*Source: Helbora, 2008)

Woody encroachments also form preferential flow paths via continuous root channels and enhanced organic matter and biological activity, thus increasing downward fluxes. While more emphasis has been placed on woody plant encroachment into grassland or savanna type vegetation, woody understory encroachment into woodlands or forest can alter the physiological state of the trees and modify over story function and therefore groundwater recharge (Grep, 2010).

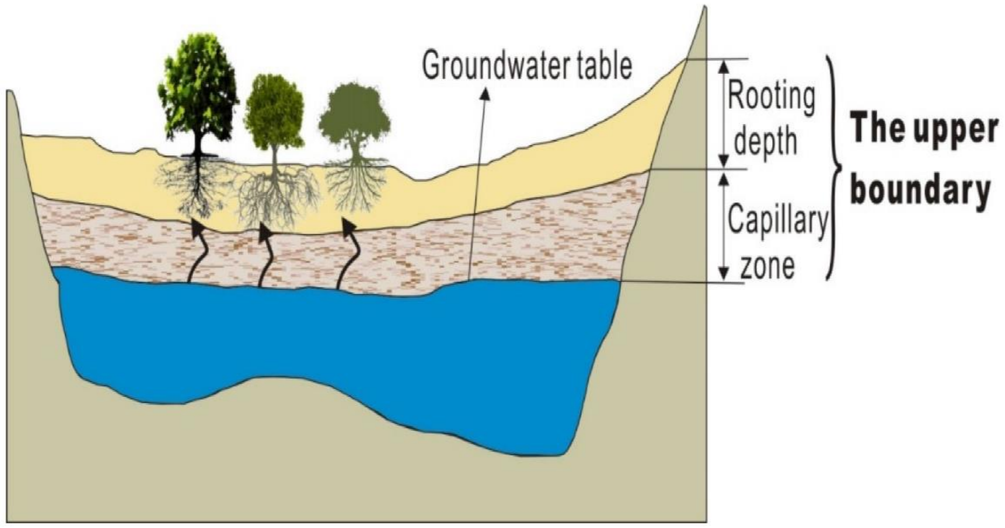


Figure 2.6: Diagram showing the conceptual upper groundwater boundary crucial to stability of present vegetation system
(Source: Fries *et al.*, 2008)

2.3 Groundwater Recharge Estimation Methods

Groundwater recharge estimation is best calculated as an iterative process, because data is always limited and circumstances vary both in time and space (Walker, 2002). Xu and Beekman (2019) shows three steps that should be considered when selecting a method for estimating ground water recharge. The first step is to define the groundwater system in terms of the geological structures and the resultant flow mechanisms. Second, the complete water balance must account for all water that does not become recharge and the underlying groundwater recharge processes clear. Third, the estimate must consider the time scale for the recharge process. Lewen (2008) states that estimates that are based on the summation of shorter-term steps are better than those based on longer term steps for the same duration. Estimating the rate of aquifer replenishment is the most difficult of all measures in the evaluation of groundwater resources.

Estimation of groundwater recharge requires modeling of the interaction between all of the important processes in the hydrological cycle such as infiltration, surface run-off, evapotranspiration and groundwater level (Jyrkama and Sykes, 2007). Many methods are available for quantifying groundwater recharge as there are different sources and processes of recharge and each of the method has its own limitation in terms of applicability and reliability (Scanlon *et al.,* 2002). Large scale vegetation determines the amount of net rainfall, infiltration rate, deep drainage and the available storage capacity of the groundwater system. Any change in vegetation say from forest to grassland can have a large effect on recharge (Topaz, 2000).

Vegetation influences recharge through interception and transpiration. The amount of stored water that can be removed by vegetation depends mainly on the rooting depth. Shallow rooted grasses will remove less water than deeper rooted shrubs and trees (Jyrkama and Sykes, 2007).It is well known that the degree of water saturation of the root zone determines the distribution of hydraulic conductivity and as a result the percolation to the groundwater table. It also influences the water uptake by roots and thus the actual evapotranspiration (Berendrecht, 2004).

The process of groundwater recharge is not only influenced by the spatial and temporal variability in the major climate variables, but is also dependent on the spatial distribution of land surface properties and the depth and hydraulic properties of the underlying soils (Simmer, 2003).

Table 2.1: Appropriate techniques for estimating groundwater recharge in regions with arid, semi-arid and humid climates

Hydrologic zone	Groundwater recharge techniques/methods	Estimation
	Arid and semi-arid climate	Humid climate
Surface water	Channel water budget Seepage meters Heat tracers isotropic Tracers water shed modeling	Channel water budget Seepage meters Base flow discharge Isotropic tracers Watershed modeling
Unsaturated zone	Lysimeters Zero-flux plane Darcy's law Tracers [historical(^{36}Cl,^{3}H) environmental (Cl)] Numerical modeling	Lysimeters Zero-flux plane Darcy's law Tracers (applied) Numerical modeling
Saturated zone	——— ——— Tracers [historical (CFCs ^{3}H / ^{3}He), environmental (Cl,^{14}c)] Numerical modeling	Water table fluctuation Darcy's law Tracers [historical (CFCs, ^{3}H / ^{3}He)], Numerical modeling

(Source: Scanlon *et al.*, 2002)

Christoph *et al.* (2011) introduced a new approach for investigation of the unsaturated zone through a combined use of laboratory and field technique in arid environments and this shown in Table 2.1. This technique uses direct push technique to get undisturbed soil samples, extraction of pore water for isotope analysis and application of Time Domain Reflectometry (TDR) to determine moisture content. Combination of these techniques resulted in a better estimation of present and historic ground water recharge.

Ahmadi *et al*. (2013) used water balance principle (rainfall –ground water level relationship) based approach to estimate ground water recharge. These methods are Water Table Fluctuation (WTD), Distributed Hydrological Budget (DHB), and Hydrological Budget (HB). These methods are useful, easy to use, cost effective, simple, requiring few data such as ground water level measurements, rainfall, aquifer properties and ground water extraction datasets. Use of these methods helps to provide irrigation return flow percentage and contribution of precipitation to natural ground water recharge.

According to Leon (2013) the most important methods available for estimating ground water recharge can be categorized as follow, direct measurements, water-balance methods, hydrological models and tracer methods.

2.3.1 Direct measurement-lysimeter

A lysimeter is a device consisting of an in situ weighable column or volume of soil for which the inflow and outflow water can be measured and changes in storage can be monitored by weighing (The technique is used to determine evaporation in a natural environment by measuring the other water balance components, but as it is mentioned, measuring recharge using this method at reasonable spatial scale is difficult (Truze, 2016). The different lysimeters are shown 2.7, 2.8 and 2.9 respectively.

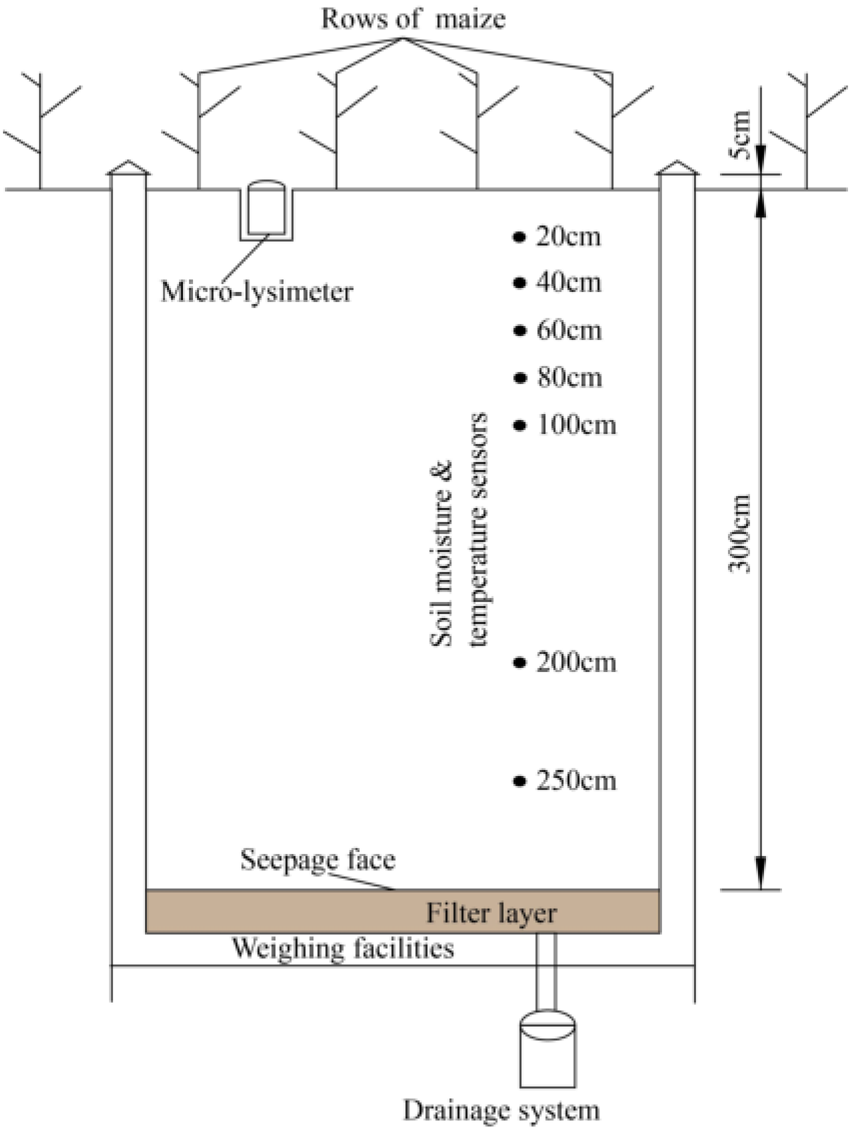


Figure 2.7: Lysimeter structure
(Source: Zeng, 2003)

Figure 2.8: Polyethylene (PE-HD) lysimeter station with four lysimeters in a cloverleaf arrangement with an entering hatch (center position) during the installation process (*Source:* Nelson, 1998)

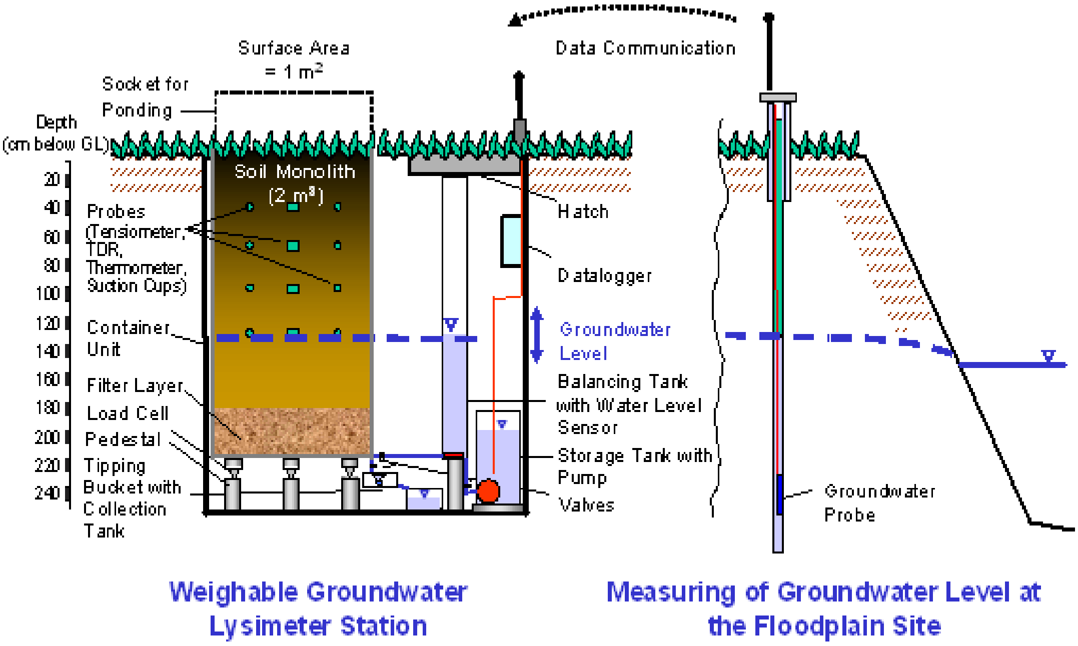


Figure 2.9: Schematic of weighable groundwater lysimeter with a groundwater control system and radio data transmission (Source: Hory, 2000)

2.3.2 Water balance method

The basis of the water balance method of estimating groundwater recharge is that the soil is free draining when the moisture content of the soil reaches a limiting value called the field capacity.

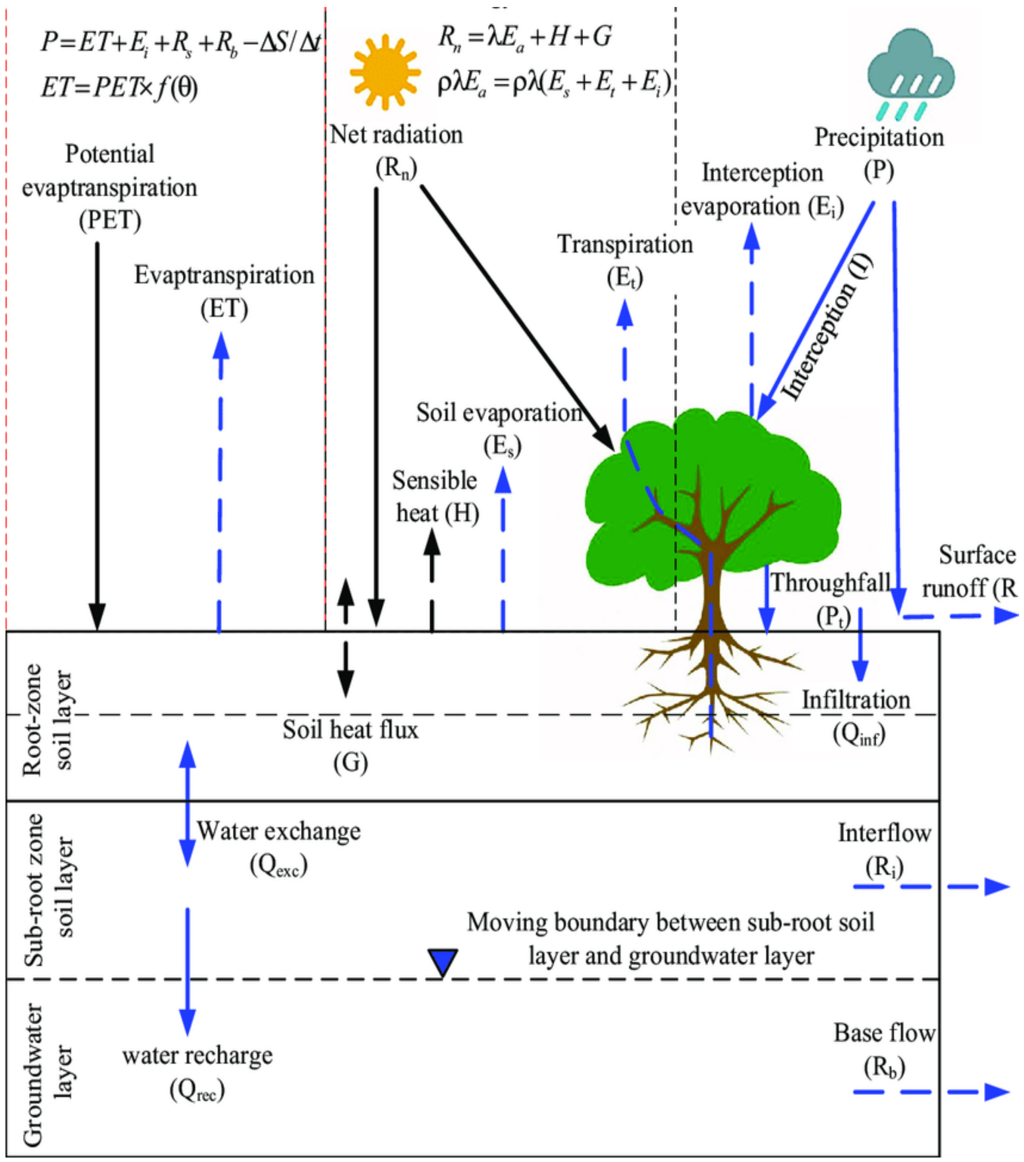


Figure 2.10: Schematic Diagram of Single Cell Water Balance in the Hydro-Informatic Modelling Systems and Penman-monteith Leuning (Source*:* Gibson, 2000)

To determine when the soil reaches the critical condition of field capacity, it is necessary to simulate soil moisture conditions throughout the year. This involves the representation of the relevant properties of the soil and the capacity of crops to collect moisture from the soil and to transpire water to the atmosphere as illustrated in Figure 2.10 . If no crops are growing or if there is only partial crop cover, bare soil evaporation must be considered (Melyn, 2010). Bare soil evaporation is important both in semi-arid locations to represent soil moisture conditions at the end of the dry season and in temperate climates where recharge occurs in winter when evaporation is usually the major loss from the soil (Klopp, 2009). Transpiration and evaporation often occur at less than the potential rate due to crop stress arising from limited soil moisture availability. The input to the soil water balance is infiltration which equals the daily precipitation minus interception or run-off (Greenbec, 2013).

Soil water balance is written as $R = P – D – ET –\Delta W$

Where:

R = Recharge [L]

P = Precipitation [L]

D = Net run-off [L]

ET = Actual evapotranspiration [L]

ΔW = Change in soil moisture storage [L]   (Derick, 2009).

2.3.3 Hydrological models

Different types of models are available for determining recharge: one-dimensional semi-distributed numerical models such SWAP, one-dimensional lumped parametric models such as EARTH and three-dimensional fully distributed numerical groundwater flow models such as MODFLOW (Creazy, 2009). According to Bran (2011), the advantage of the hydrological models is that the impact of transferring water between competing sectors can be simulated and the effects of man-induced scenarios on regional hydrology can be studied.

The unsaturated zone physically based numerical models such as SWAP solve the unsaturated zone rate flow equation called the Richards equation for porous media (Clare, 2007). In contrast to the lumped parametric water balance models, numerical models allow detailed evaluation of the effects on groundwater recharge of Vadoze zone hydraulic properties and their spatial variabilities. These methods are based on soil profile partitioning with a number of homogeneous layers with their own characteristics hydraulic properties. They simulate the transformation of precipitation into flow taking into account all the intermediate processes such as evapotranspiration, interception, infiltration and run-off. They are therefore able to estimate recharge at many points and at many times.

For simulating recharge, boundary and initial conditions must be imposed on the models together with hydraulic soil and vegetation properties (Anselm, 2011). Parametric models such as EARTH use a numerical or analytical relationship between precipitation and recharge. These models have been developed to deal with conceptual recharge situations that cannot be encompassed by existing numerical models (Clifford, 2010). Arnold (2007) stated that parametric models such as EARTH can be used both in porous and hard rock formations.

Model such as the MODFLOW model uses the fully distributed three-dimensional numerical groundwater flow to estimate groundwater recharge by adjusting the recharge input value in the model until groundwater levels calculated by the model match the aquifers measured water level (Frank, 2008).

The uncertainty associated with transmissivity is larger than the uncertainty associated with recharge of groundwater; thus the accuracy of the estimated recharge may be low (Mac, 2004).

2.3.4 Tracer methods

There are three kinds of tracers. However, the most commonly used in this field are the environmental tracers (Mez, 2003). Tracers are dissolved substances introduced into the large scale water cycle either by nature or by men over long periods (Kiz, 2010). Tracers are able to trace water movement over long periods. In contrast to artificially applied tracers which show water movement over small spatial and temporal scales, the most important tracer is chloride (Greg, 2007). Historical tracers (3H and 36 Cl) resulting from human activities or historical events in the past were used to estimate the groundwater recharge by the location of tracers concentration peak, assuming the piston flow for soil water (Jun, 2015). An environmental tracer such as chloride was used to estimate the groundwater recharge by the mean concentration of chloride below the rooting zone supposing that the concentration of chloride keeps constant below the depth of the rooting zone (Freda, 2006).

Transport of environmental tracers through the unsaturated soil zone is a combination of advective and diffusion transport in both soil water and air. Water flow in the soil zone is very complex on a microscopic scale. Therefore it is practically impossible to estimate water velocities or fluxes accurately within individual pores. However as the tracers are permanently exchanged between infiltrating water and stagnant water, they reveal an average pore water velocity including the entire soil moisture (Bromley *et al.*, 1997). Main applied tracers include tritium and fluorescent dyes considering conservative tracers (Bull, 2011). Lin *et al.* (2000) and Ali (2017) used environmental tracers Cl, F and SO_4 to estimate groundwater recharge beneath irrigated farmland of North China plain. Coplen (1993) reported that another major technological growth area has been in the application of isotropic analysis to groundwater hydrology, wherein isotropic measurements are being used to help interpret and define groundwater flow paths, ages and leakages.

2.4 Cassava as a Common Plant in the Basin

Cassava is one of the world's most important food crops with annual global production at approximately 276 million metric tons in 2013.The top producing countries globally in 2013 were, Nigeria accounting for 19%, Thailand 11 %, Indonesia 9%, Brazil 8% and Democratic Republic of Congo 6% (Shola, 2009). Global demand for the commodity has been growing significantly between 2004 and 2013 because of its appeal as a food security crop for growing populations in emerging markets and the growing demand for industrially processed cassava products (Olayinka, 2016).

Cassava has traditionally played an important role as an irreplaceable food security crop in large parts of developing world (Bokanga, 1995). Whenever cassava is grown, it is primarily used as food .The exception to this rule is Thailand where 90 percent of the cassava produced is exported and the rest is used in industries (Nweke, 1994). A sample of cassava in Plate I , while harvested one is shown in Plate II

Plate I: Cassava Farm in Nigeria
(Source*:* Taiwo *et al.*, 2014)

Plate II: Cassava Roots
(Source: Obeya, 2016)

CHAPTER THREE

3.0 MATERIALS AND METHODS

3.1 Description of Study Area

The Otukpo basin is located in Benue State, North Central part of Nigeria, and this can be seen in Figure 3.1. It is bordered geographically by latitudes 7^{o} 12` 60.00`` N and Longitude $8^{o}$ 08` 60.00`` E. Climatically, the town belongs to the Kopper's Aw climate group and experiences, seasonal wet and dry seasons. The rain falls for seven months from April to October, while dry season sets in November and ends in March (Ologunorisa, 2006). Temperatures are constantly high averaging between 28^{o} – 32^{o}C and sometimes rising to 37^{o}C.

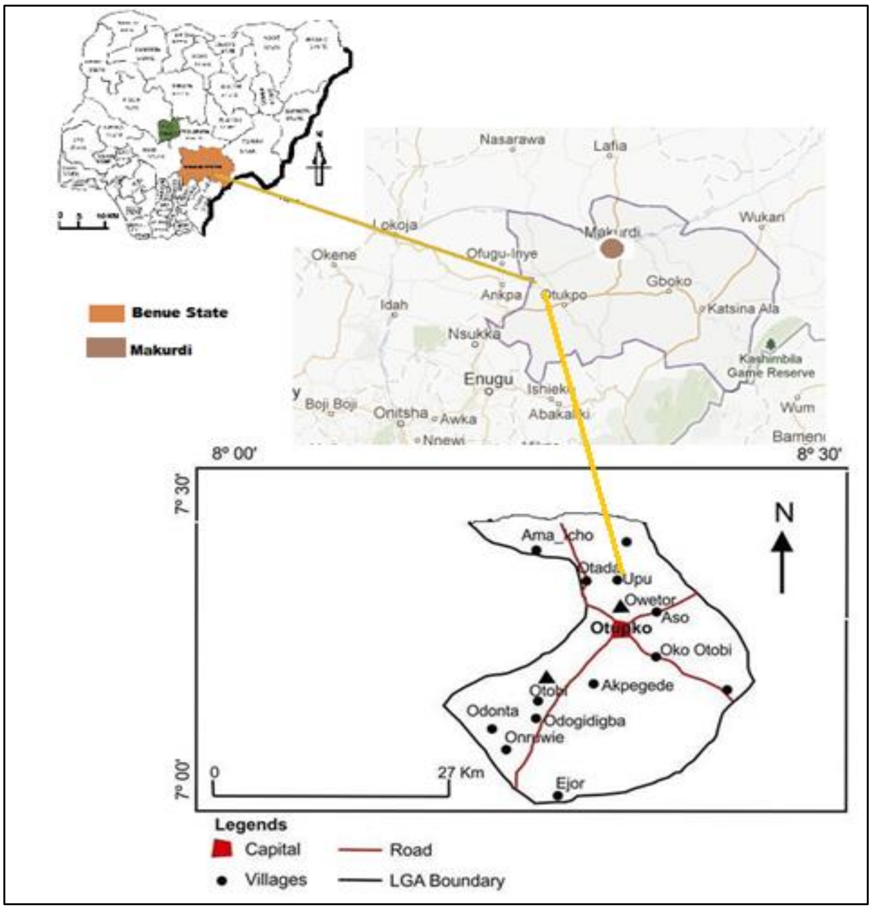


Figure 3.1: Map of Otukpo Basin showing the study area

3.2 Materials

1. Ten samples of soil

2. Metrological data including daily rainfall, minimum and maximum temperature.

3. Oven

4. Weighing balance

5. Cylindrical moulds

6. Metal rammer

7. Samba Model

Ten different soil samples obtained from different parts of Otukpo basin. These samples were kept in an air-tight container and transported to a soil and water laboratory where the samples were compacted and other properties such as moisture content, total available water, field capacity and wilting point were determined. The rainfall, temperature relative humidity, sunshine hours and radiation data were also obtained from the meteorological agency office located at the Nigeria Air Force Base, Makurdi. The equipment used for the determination of soil moisture content were $1000m^3$ cylindrical moulds, metal rammer, straight edge, weighing balance, moisture cans, scoops and oven.

3.3 Method of Experimentation

3.3.1 Moisture content determination

3000 g each of the soil samples air-dried were weighed afterwards, 3 % by weight of sample was also weighed and mixed thoroughly with a rammer that fell through a length of 300 mm. The soil was compacted in three layers with 27 blows on each layer to make sure that the blows are evenly distributed. The collar of the mould was removed and the soil was trimmed off evenly with straight edge, the cylinder would be cleaned properly and the mould with the sample would be weighed. Some quantities of the compacted soil sample were removed from the bottom and top for moisture content determination. The soil was broken by hand and another 3% by weight of soil sample added and mixed thoroughly, this was done repeatedly until the weight of soil and mould dropped. The containers with soil sample were weighed each and their weights recorded likewise.

The moisture content of the soil is the ratio of mass of soil which is expressed in percentage. Oven dry method and pycnometer method are commonly used to determine the moisture content of soil in the laboratory, but for the sake of the research work, the oven dry method was used because of accurate results. In the oven dry method the containers were well labeled for proper identification of the different soil samples from different zones in the Otukpo basin. These containers were kept in the oven for 24hours maintaining a temperature of $110^0 \pm 5^0$C. The containers were then removed from the oven and allowed to cool. After the determination of the weight of soil samples using the oven method, the moisture contents were then calculated using the formular below:

$$\text{Moisture content} = \left(\frac{w_2 - w_1}{w_3 - w_1}\right) \times 100 \qquad (3.1)$$

Where

w_1 = weight of container

w_2 = weight of container with wet soil

w_3 = weight of container with dry soil

3.3.2 Determination of bulk density

The bulk density of the different soil samples was also determined in the laboratory, using the mould and base with a volume of $1000m^3$ and the weights of the mould and base were recorded as W_1 in grams, and the weights of the mould, base and soil samples were recorded also as W_2 in grams,

Calculation:

The bulk density δ_b in Mg/m^3 of each compacted soil sample from the different points in the Otukpo basin was calculated.

Calculation:

$$\delta_b = \frac{W2 - W1}{1000} \tag{3.2}$$

Where:

W_1 = weight of the mould and base

W_2 = weight of the mould, base and soil samples

3.3.3 Determination of dry density

The dry bulk density is the weight of the dry soil divided by the total soil volume.

The soil samples in each mould was then placed in an oven at a temperature of 105^0c for 24hours and after then the soil sample in each mould was weighed after it must have cooled and the weights were recorded.

The dry bulk density would be calculated using the equation below:

$$\delta_b \quad = \quad \frac{100\ \delta b}{100 + M} \tag{3.3}$$

3.3.4 Determination of field capacity

The field capacity of each of the soil samples was determined in the laboratory and a pressure plate was used and a suction of -$^{1}/_{3}$ atmosphere to a saturated soil sample was applied.

Calculation

$$\theta_{fc} = \frac{\gamma b * x W f c}{\gamma w} \tag{3.4}$$

The field capacity is the amount of soil moisture water content held in the soil after excess water has drained away and the rate of downward movement has increased. This usually takes place 2-3 days after rain in pervious soils of uniform structure and textures (Frimpong, 2017).This was determined in the laboratory for each of the soil sample provided in the laboratory.

3.3.5 Determination of permanent wilting point

The permanent wilting point of the ten soil samples was determined from the moisture content corresponding to a pressure of -15 atmosphere from the pressure test that was carried out.

Volumetric moisture content at field capacity was determined also

Calculation $\theta_{wp} = \frac{\gamma_b\, W\, x\, Wp}{\gamma_w}$ (3.5)

Where:

W_{fc} and W_{wp} are the dry weight moisture fraction at each point.

3.3.6 Determination of the total available water

The total available water of each soil sample was determined and recorded in the laboratory. The total available water TAW is the difference between field capacity and wilting point moisture contents multiplied by the depth of the root zone.

Calculation:

Total available water = (θ_{fc} - θ_{wp}) RD, where RD is the depth of the root zone. (3.6)

3.4 Meteorological and Hydrological Data

Secondary data inputs for the study were acquired from the Nigerian Meteorological Agency (NIMET), located within the office of the Tactical Air Command (TAC) of the Nigerian Air Force (NAF) Makurdi, Benue State.

3.4.1 Rainfall data

The daily rainfall of Otukpo basin was obtained and it covered for eleven years from 2008-2018.

3.4.2 Maximum and minimum temperatures

The maximum and minimum temperature data of Otukpo basin would be obtained and be used to obtain the mean temperatures from the year 2008-2018.

3.5 Determination of Evapotranspiration

The evapotranspiration would be calculated using the Hargreaves equation shown below

$$ETO = 0.0023 * (Tmean + 17.8) * (Tmax – Tmin) 0.5*Ra. \quad (3.7)$$

Where;

T_{mean} = Mean temperature

T_{max} = Maximum temperature

T_{min} = Minimum temperature

ET_O = Evapotranspiration

Ra= Extra Terrestrial solar radiation for Makurdi and the value is 15.702 $MJm^{-2}day$ (Audu *et al*., 2013).

3.6 Modified Soil Moisture Balance

A simplified daily soil moisture balance model is used which is based on the methodology described by Eilers *et al.* (2007) which also lists the relevant algorithms; calculations can be performed using an Excel spreadsheet or any other program. But in this study, Python was used in writing a programe for the execution of the algorithm. Other programmes that could still be used include languages like FORTRAN, BASIC and JAVA. Python is a generic, interpreted scripting language, supporting object-oriented programming which was first released in 1991.

The representation of crops and soils using this approach is based on FAO guidelines (Allen *et al.*, 1998). The estimation of potential recharge estimation using a modified soil moisture balance model (MSMB) is based on the fact that the soil becomes free draining when the moisture content of the soil exceeds a limiting value called the field capacity when excess water then drains through the soil to become potential recharge. Therefore, in order to determine when the soil reaches this critical condition, estimating soil moisture conditions on a daily basis throughout the water year becomes crucial. This is achieved by representing the appropriate properties of the soil, and also the ability of crops to take up moisture from the soil and to transpire to the atmosphere. The conceptual and computational models of this approach are as shown in Figure 3.2.

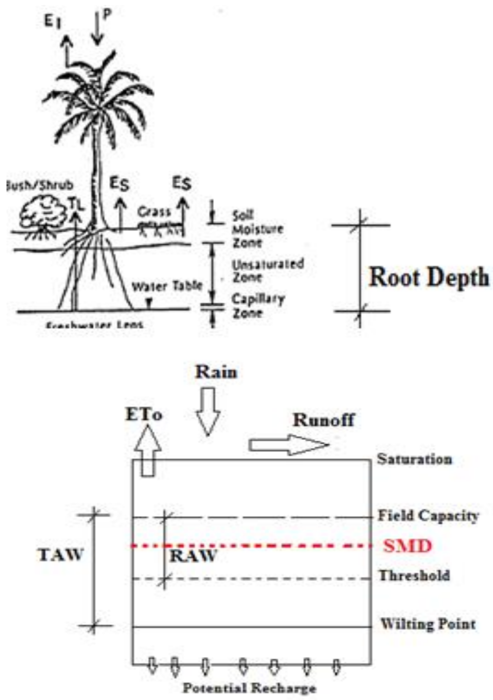


Figure 3.2: Conceptual and Computational Models of Soil Moisture Balance
(Source: Adesiji *et al.*, 2020)

Predominantly, the land use in the upland area of the study area is permanent grass with few trees; there are also vegetable plots around the areas where the soil samples for the laboratory analysis were collected. Input parameters for the soil moisture balance are highlighted in Table 3.1. The parameters are deduced from Allen *et al. (*1998) Eilers *et al. (*2007) and from farmers' information on planting and harvesting dates in the study areas. Soil in the uplands of the study area is well drained sandy clay loam, which, according to the laboratory results, was observed to have moisture content at field capacity of 0.55 m^3/m^3 and moisture content at wilting point of 0.23 m^3/m^3. The coefficient for near surface storage for grass is selected to be FRACSTOR = 0.70 based on studies in locations with similar soils. The crop parameters highlighted in Table 1 are selected based on the predominant crops in the study area.

Table 3.1: Crop and soil parameters for the soil moisture balance of oil palms study plots

Parameters/Year of cultivation	**2000**
CROP PARAMETERS:	
Maximum root depth (m)	0.50
*Depletion factor	0.70
Kc (initial)	0.15
Kc (development)	0.70
Kc (mild stage)	1.00
Kc (late)	1.00
SOIL PARAMETERS:	0.302
Bulk density (gcm^{-3})	0.55
VMC @ Saturation ($m^3\ m^{-3}$) $\boldsymbol{\theta_{sat}}$	0.55
VMC @ Field capacity ($m^3\ m^{-3}$)[θ_{sat} x $\frac{\gamma_b}{\gamma_w}$]	0.23
VMC @ Wilting Point ($m^3\ m^{-3}$) **[FC/2.4]**	35.5
Maximum TAW (mm)**[FC-WP]/900**	24.9
Maximum RAW (mm) **[TAW*0.7]**	58.3
Soil Moisture Deficit (mm)	0.70
*NSS Factor	0.50

* Depletion factor (Allen *et al.* 1998)

**NSS factor (Eilers *et al.,* (2007)

Actual evapotranspiration and potential recharge are calculated from daily rainfall data and the daily Penman-Monteith reference evapotranspiration of grass, ETo. Rainfall was recorded in the study area with a tipping bucket rain gauge. The CROPWAT model (Smith, 1992) was used to calculate the FAO adapted Penman-Monteith reference evapotranspiration for the study period. The crop potential evapotranspiration PE is calculated from ETo by multiplication with the crop coefficient *K*c. Crop coefficients for various crops are listed in Allen *et al.* (1998). The *K*c values vary during the crop period from initial stage, development stage, maturity and ripening stages; however, for grass, *K*c remains constant at 1.00. Values of *K*c for eggplant are listed in Table 3.1.

For the successful application of MSMB model, the structure below was used and followed with the input of the hydrological components;

(i) Daily rainfall and reference evapotranspiration (ET_o)

(ii) Use *SMD* at the driest season as initial soil moisture deficit - SMD

(iii) Compute runoff coefficient, using the runoff matrix

(iv) Compute the Runoff = Rainfall * Runoff coefficient

Obtain Runoff Coefficients through ‘trial and error’ approach

(v) Determine Available water for evaporation (AWE)

If SMDpr < 0, AWE = Rainfall – Runoff

AWE (Jan 3rd) = 47 - 19.74 = 27.3mm, This is when $SMD_{prev} < 0$

(vi) Compute crop coefficient *Kc* using information on planting date and crop duration

(vii) Potential evapotranspiration (PE) = *Kc* * ETo[*Kc* = 1.0 for mature oil palm]

(viii) Actual evaporation (AE) = PE, When SMD < TAW * *Zr*

Where *Zr* represents maximum root depth in *m* and

Zr = 0.9 m (as the oil palms are already mature)

(ix) Total available water, TAW is determined as:

TAW = [(FC-WP)*1000*Zr

(x) Readily available water, RAW = TAW * *ρ* (*ρ* is a depletion factor constant between 0.2 and 0.7, Allen *et al.*, 1998). Here 0.7 is used for peatland soil

(xi) Determine soil stress coefficient, *Ks* as follows:'

'SMD denotes soil moisture deficit at the end of day t, while SMDpr denotes previous day SMD.'

Rech denotes recharge at the end of day t, while Rechpr denotes previous day recharge

NSS is near surface storage at the end of day t and NSSpr is the previous day NSS

NSS factor is the storage fraction of near surface storage.

NSS = (AWE – AE) x 0.45, where 0.70 is a NSS constant (Eilers *et al.,* (2007)

SS (Jan 3rd) = (27.3 5.1) x 0.45 = 9.99 = 10 mm

Groundwater Recharge = [SMD_{pre} – 1] + NSS

Recharge only occurs when the SMD ≤ 0

CHAPTER FOUR

4.0 RESULTS AND DISCUSSION

4.1 Application of the Model for the Eleven (11) Years Period

Figure 4.1, shows the modified model GUI developed to estimate the groundwater recharge of Otukpo basin from 2008 to 2018.This particular model was developed using Phyton programming language. Cassava is a predominant crop in the basin and was therefore used to run the model.

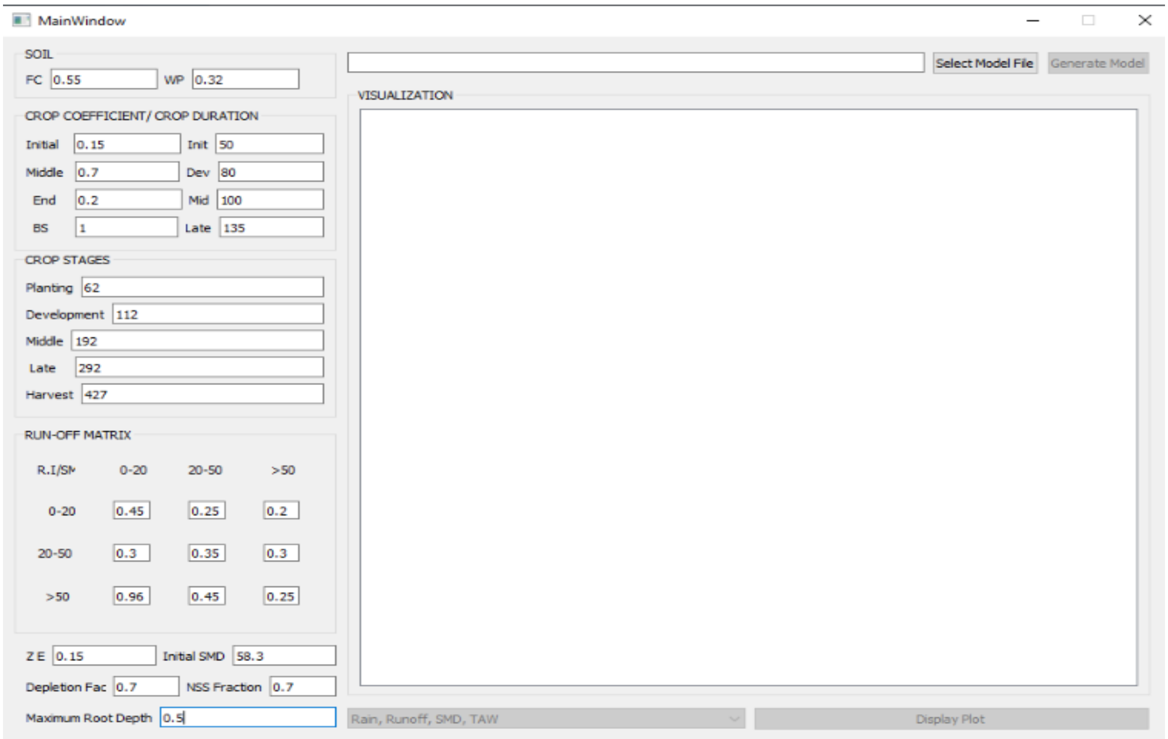


Figure 4.1: Modified soil moisture balance model (MSMB Model)

The total rainfall computed using the Modified Soil Moisture Balance Model (MSMB Model) for the year 2008 was 1608.2 mm and this can be seen in Figure 4.2 shown. The Total Available Water computed using the model for the year 2008, was 32741.525mm and a total Runoff of 898.362mm, while the total Soil Moisture Deficit was 16703.833mm.A Julian day of 365 days was used in the computation.

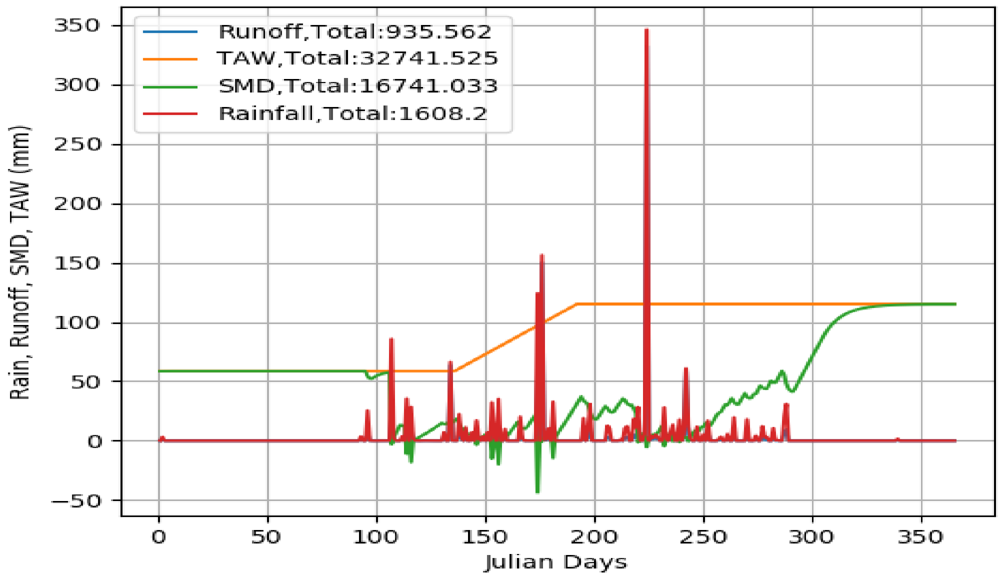


Figure 4.2: Total rainfall, total soil moisture deficit, total available water and total runoff for the year 2008

The model was able to calculate the monthly precipitation of the first year which was 2008. In the Figure 4.3 shown, month of August, 2008 recorded the highest precipitation of 591mm, and the months of July September and October had the same amount of 80 mm of precipitation. From the figure, precipitation became obvious in the month of April.

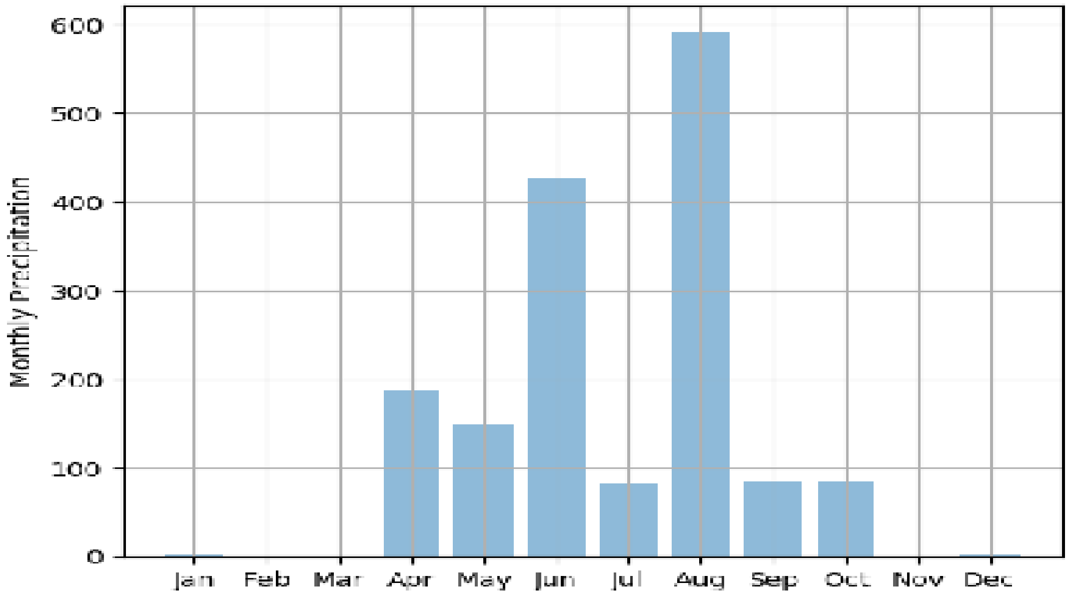


Figure 4.3: Monthly precipitation of 2008

From Figure 4.4, ETo of 5.66mm/day was earlier recorded from the 1st to the 59th Julian day and there was an increase to 6.07mm/day from the 60th to 90th Julian day. The highest ETo of 6.57mm/day was recorded from the 335th to 365th Julian day.4.14 mm/day was recorded as the lowest value of ETo recorded from 213th to 243th. Potential Evapotranspiration (PE) was highest with a value of 6.07 mm/day from 60th to 61st Julian day and the lowest value of PE was 0.786mm/day on the 112th Julian day. Actual Evapotranspiration AE at the earlier Julian day was recorded as 0 mm/day.

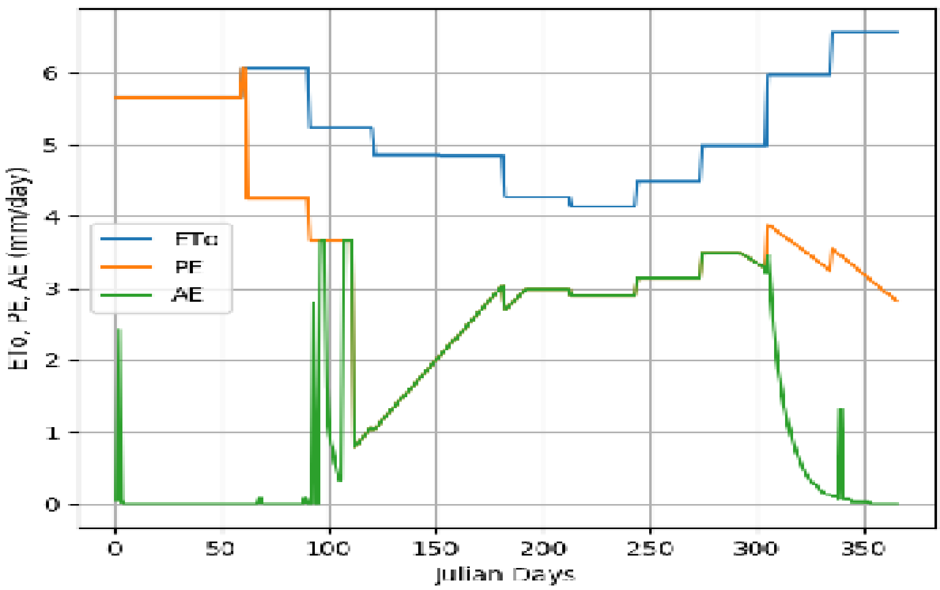


Figure 4.4: Reference evapotranspiration, potential evapotranspiration and actual evapotranspiration 2008

From Figure 4.5, the runoff recorded first in the year 2008 was 0.6mm and it was because of the early rain experienced, but the highest runoff value recorded was 332.16mm on the 224th Julian day which happened to be during a period of constant rainfall in the Otukpo Basin. Also, a recharge of 43.56997 mm was recorded on the 174th Julian day. In the year 2008, the total value of runoff recorded was 935.562mm and a total recharge of 142.566mm. It can be seen that there was more runoff than recharge in Otukpo basin.

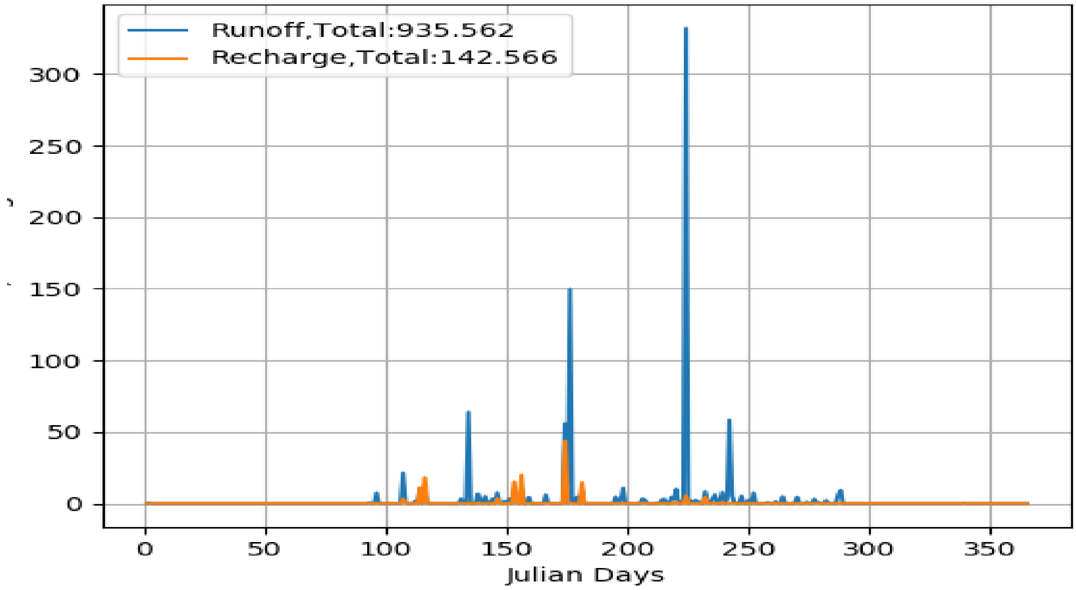


Figure 4.5: Runoff and recharge 2008

The total rainfall for the year was calculated to be 1665.4mm and this can be seen in Figure 4.6, and the year 2009 recorded a total soil moisture deficit of 14353.961mm.In Figure 4.6 too, the total available water calculated was 32741.525mm and the runoff for the year was 698.904mm.

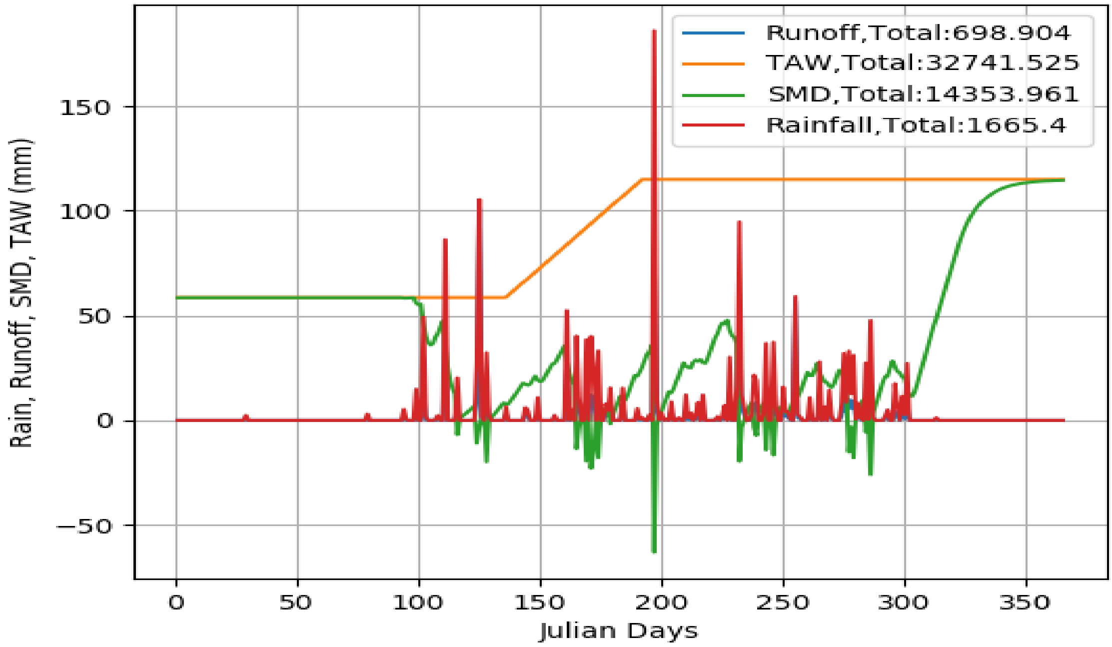


Figure 4.6: Total rainfall, total soil moisture deficit, total available water and total runoff for the year 2009

The monthly rainfall for the year 2009 was calculated and presented in Figure 4.7, the month of October recorded the highest rainfall value of 284.3mm and this is usually the peak of rainfall in Otukpo basin, and the month of August also recorded a very high value of rainfall of 275.8mm.Month of June also recorded a very high rainfall of 275.9mm slightly above the monthly rainfall in the month of August. There was a slight rainfall of 1.2 mm in November.

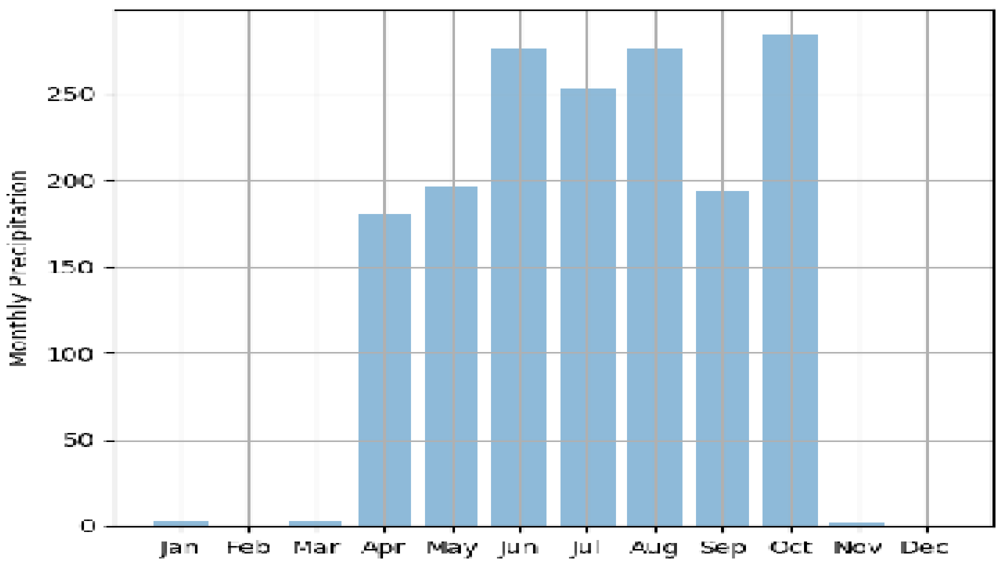


Figure 4.7: Monthly precipitation of 2009

The evapotranspiration ETo of 6.83 mm/day is the highest value recorded and this coincided from 335th to 365th Julian day in the year 2009.and evapotranspiration value is usually higher when the value of rainfall is very low or no rainfall. From Figure 4.8, 4.59 mm/day was recorded as the lowest evapotranspiration rate from the 152nd to 212th Julian day of the year 2009.

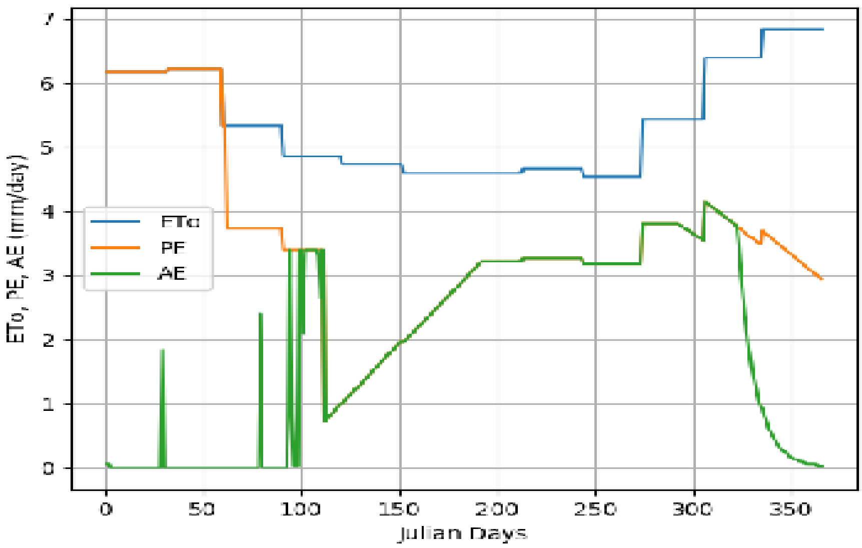


Figure 4.8: Reference evapotranspiration, potential evapotranspiration and actual evapotranspiration 2009

Figure 4.9 shows the total runoff and total recharge for the year 2009. Total recharge of 333.33 mm was recorded and a total runoff of 698.904mm.63.04739mm recharge was calculated and this is the highest recharge for the year 2009.The highest runoff calculated was 101.184mm and this occurred during the 125th Julian day.

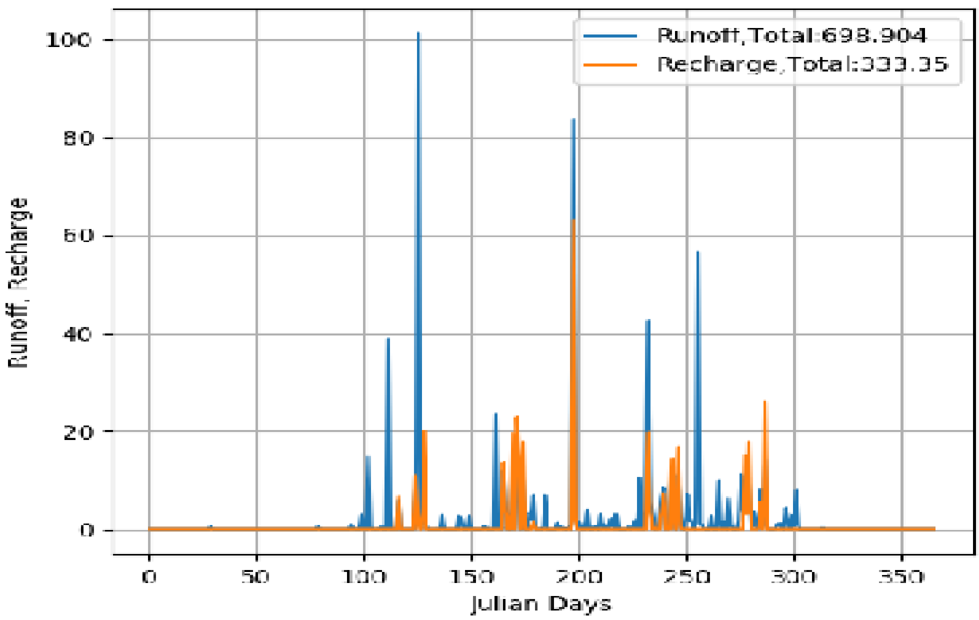


Figure 4.9: Runoff and recharge 2009

The total rainfall calculated for the year 2010, as shown in Figure 4.10 is 1211.4mm, and the soil moisture deficit calculated for the same year is 15732.608 mm. The total available water recorded is 32741.525mm, while the total runoff is 488.414mm

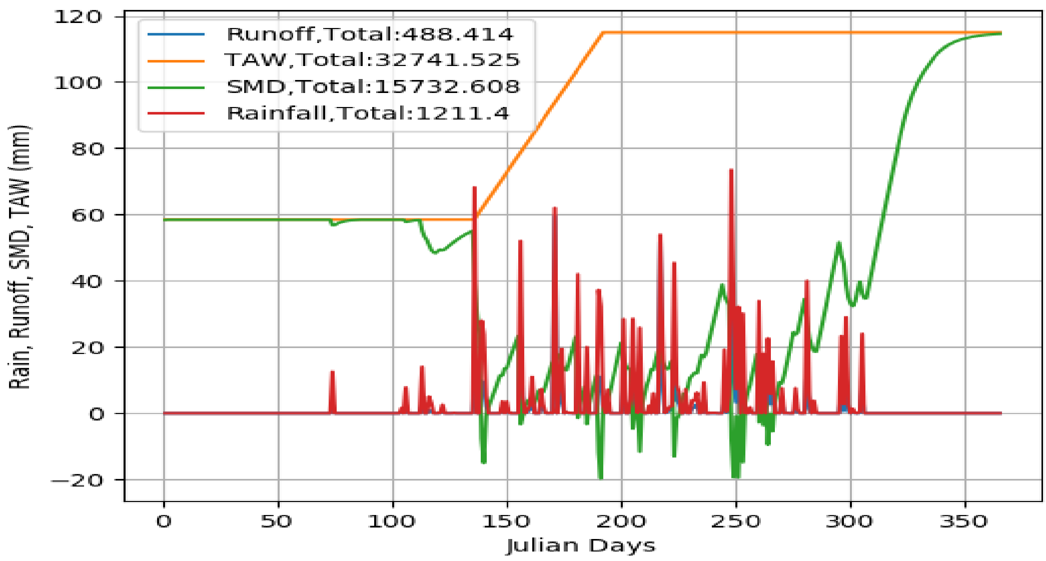


Figure 4.10: Total rainfall, total soil moisture deficit, total available water and total runoff for the year 2010

Month of September in Figure 4.11 recorded the highest rainfall of 305.9mm .The model did the rainfall calculation which gave the value for the monthly rainfall.

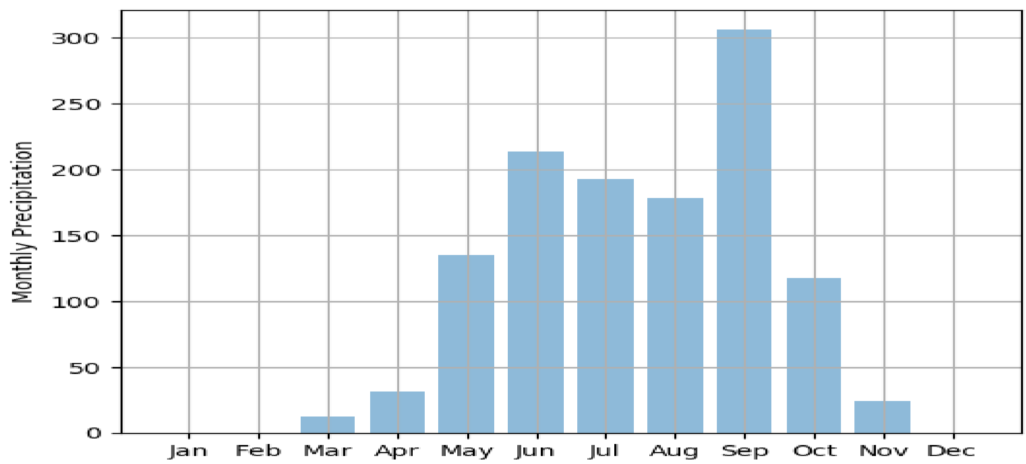


Figure 4.11: Monthly precipitation of 2010

Evapotranspiration ETo of 6.89 mm/day was calculated and this value is the highest value of ETo. 4.21 is the least value of Evapotranspiration recorded. The Potential Evapotranspiration PE, has the same highest value like the ETo and the least value of potential evapotranspiration recorded is 0.927131 mm/day. The actual evapotranspiration from Figure 4.12 recorded is 4.438mm/day which represents its peak value.

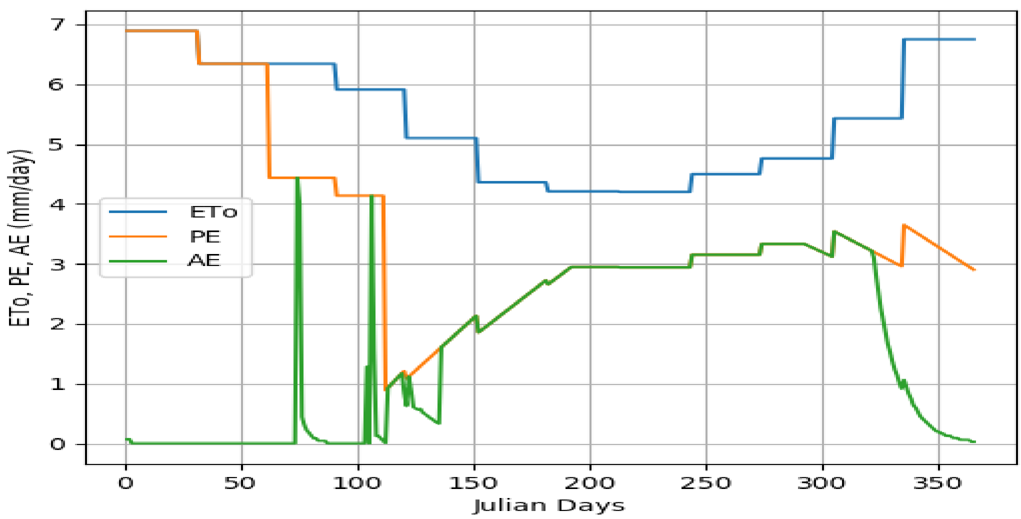


Figure 4.12: Reference evapotranspiration, potential evapotranspiration and actual evapotranspiration 2010

Total runoff of 488.414 mm was calculated by the model and a total recharge of 178.899mm was recorded for the year 2010.The highest recharge recorded is 19.69194 mm and this is on the 191st Julian day. In Figure 4.13, the total rainfall for the year 2017 was calculated as 1062.4 mm, and still in the same year the total available water was calculated to be 32741.525 mm. The total soil moisture deficit obtained was 17950.716 mm and the total runoff for the year 2017 was 496.985 mm.

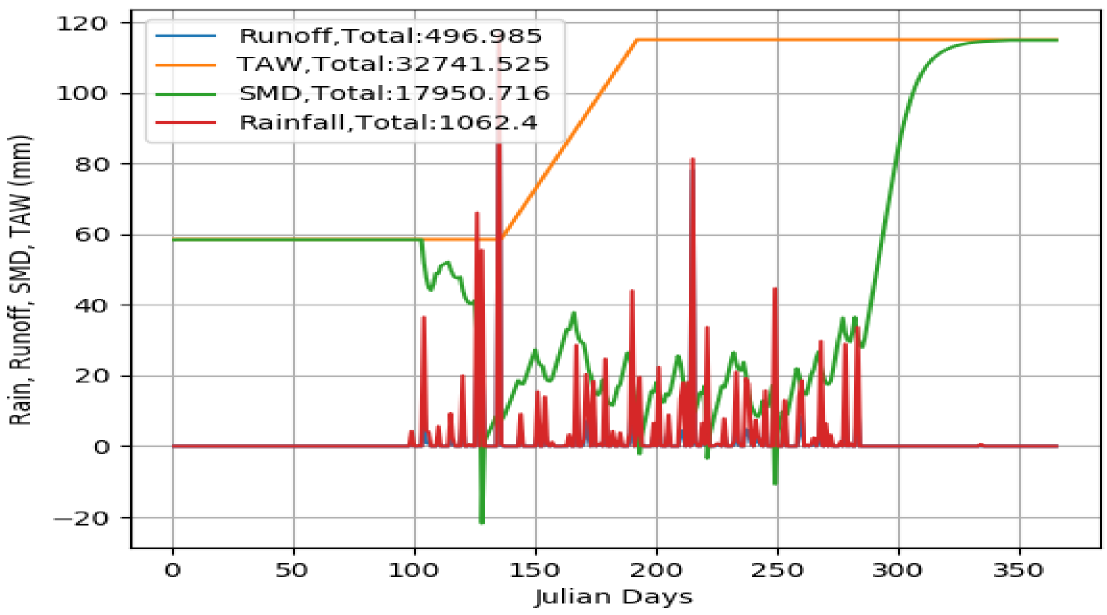


Figure 4.13: Total rainfall, total soil moisture deficit, total available water and total runoff for the year 2017

In Figure 4.14, maximum monthly rainfall was recorded in the month of May, unlike other years when the maximum rainfall would fall between July, September and October. The maximum monthly rainfall recorded is 263.80 mm and November also recorded the least amount of rainfall of 0.6 mm and normally it is very rare to have rainfall in November in Otukpo basin.

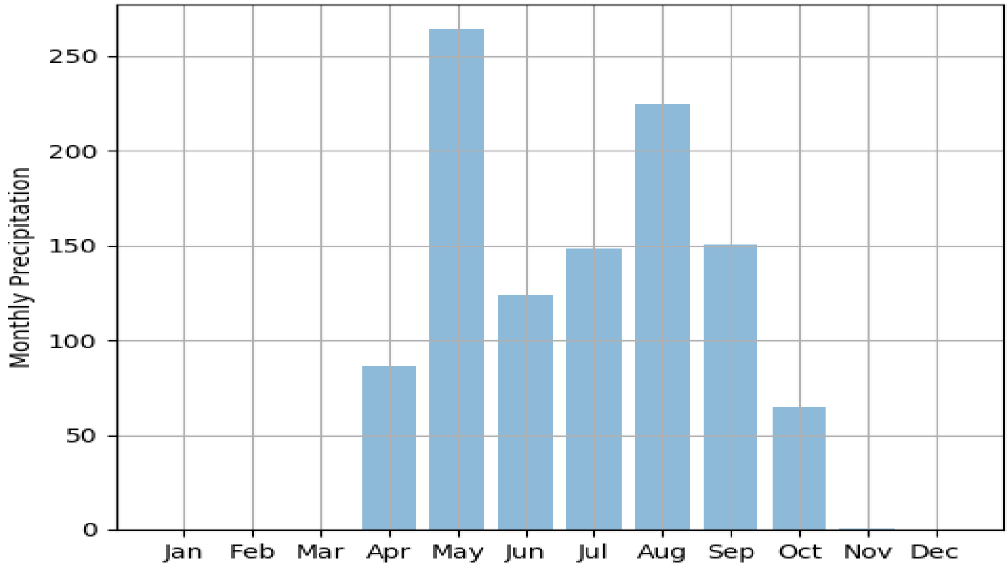


Figure 4.14: Monthly precipitation of 2017

Evapotranspiration ETo calculated has a maximum value of 7.17 mm/day and has a minimum value of 4.24 mm/day, Figure 4.15 clearly shows these values. The Potential evapotranspiration recorded has a maximum value of 7.17 mm/day which is the same as that of Evapotranspiration ETo, but the minimum value of potential evapotranspiration is 0.8625 mm/day and this is lesser than the minimum value of evapotranspiration. Actual Evapotranspiration calculated using the model has a maximum value of 4.025 mm/day and a minimum value of 0.0307 mm/day. The maximum and minimum values of the actual evapotranspiration are less than values of evapotranspiration and potential evapotranspiration.

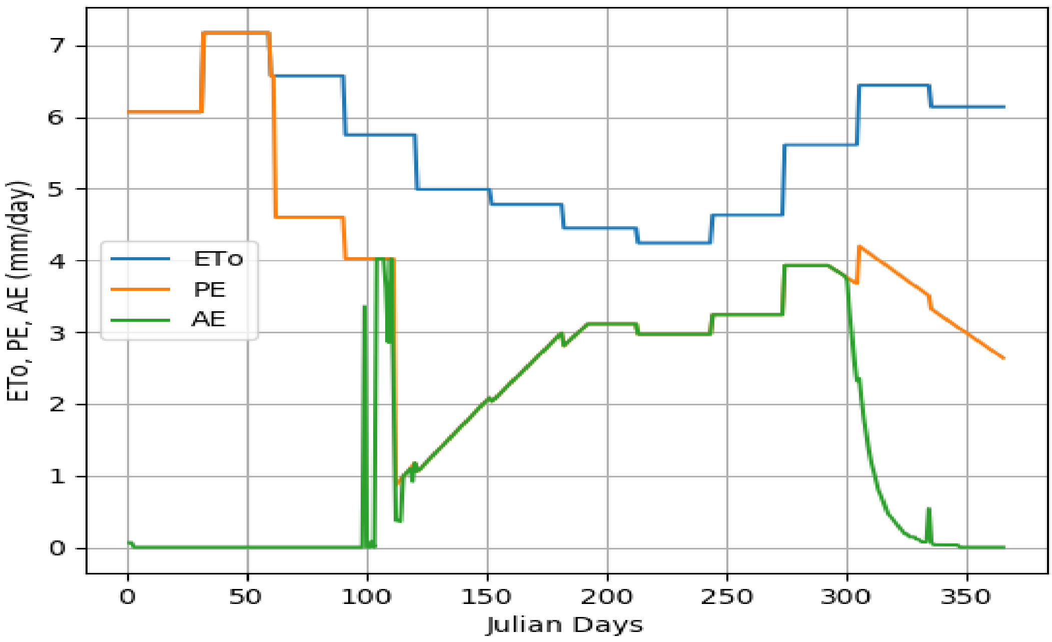


Figure 4.15: Reference evapotranspiration, potential evapotranspiration and actual evapotranspiration 2017

Figure 4.16 shows that 112.032 mm was the highest daily runoff calculated on the 135th Julian day for the year 2017.For the year 2017, the total runoff calculated was 496.985 mm. From Figure 4.16, a maximum daily recharge of 21.77404 mm was calculated for the year 2017 and 3.445 mm minimum daily recharge was calculated by the model. A total of 38.119 mm is the yearly recharge value for the year 2017.

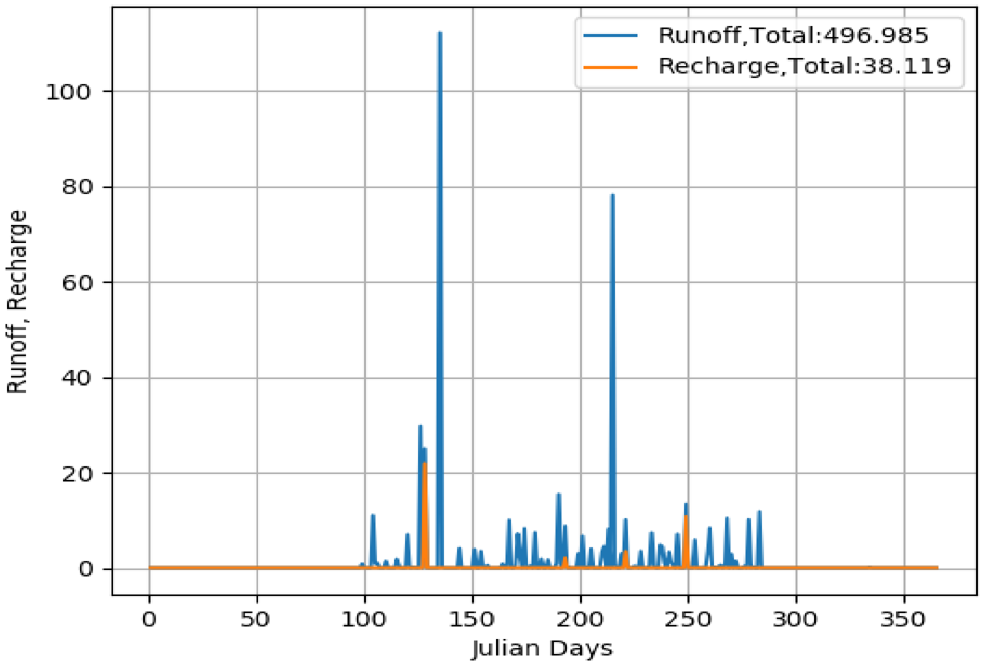


Figure 4.16: Runoff and recharge 2017

In Figure 4.17, the total rainfall for the year 2018 was calculated as 1298.9mm, and still in the same year the total available water was calculated to be 32741.525mm. The total soil moisture deficit obtained was 15458.642 mm and the total runoff for the year 2018 was 529.343 mm.

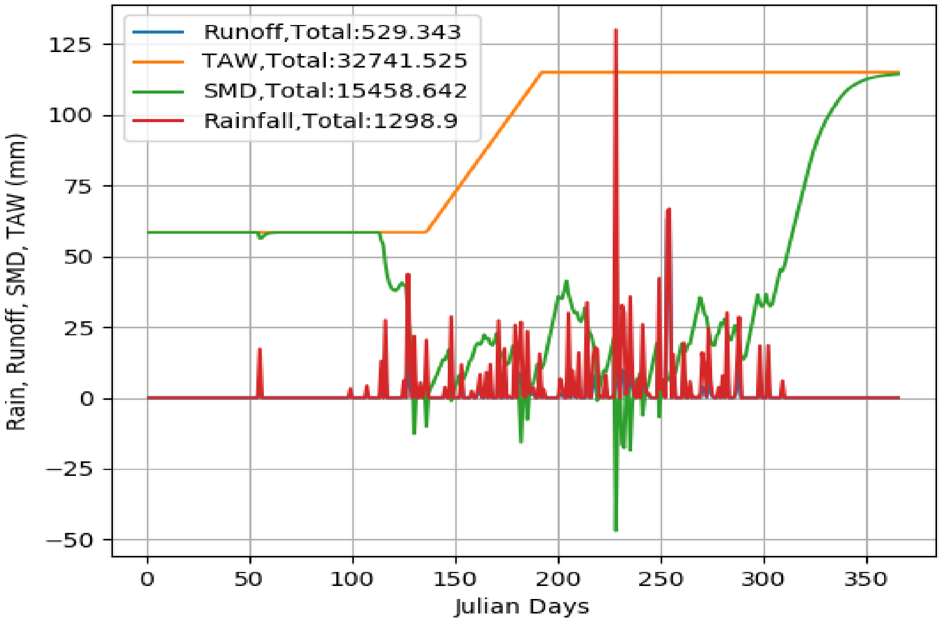


Figure 4.17: Total rainfall, total soil moisture deficit, total available water and total runoff for the year 2018

In 2018, first rainfall was experienced in the month of February with a total monthly rainfall of 17.2 mm, and month of March did not experience any form of rainfall. August recorded the highest amount of rainfall of 386.2 mm. November 2018 from Figure 4.18 recorded the lowest amount of rainfall of 6mm.

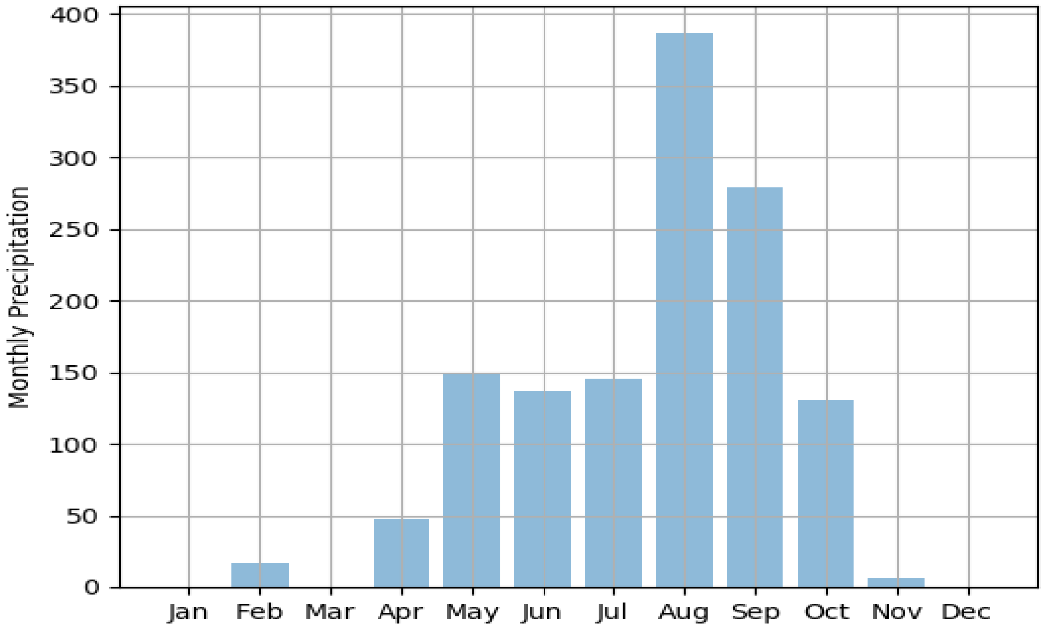


Figure 4.18: Monthly precipitation of 2018

Evapotranspiration ETo calculated has a maximum value of 6.93 mm/day and has a minimum value of 4.69 mm/day, Figure 4.19 clearly shows these values. The Potential evapotranspiration PE recorded has a maximum value of 6.93 mm/day which is the same as that of Evapotranspiration ETo, but the minimum value of potential evapotranspiration is 0.8835 mm/day and this is lesser than the minimum value of evapotranspiration. Actual Evapotranspiration AE calculated using the model has a maximum value of 6.76 mm/day and a minimum value of 0.056625 mm/day. The maximum and minimum values of the actual evapotranspiration are less than values of evapotranspiration and potential evapotranspiration.

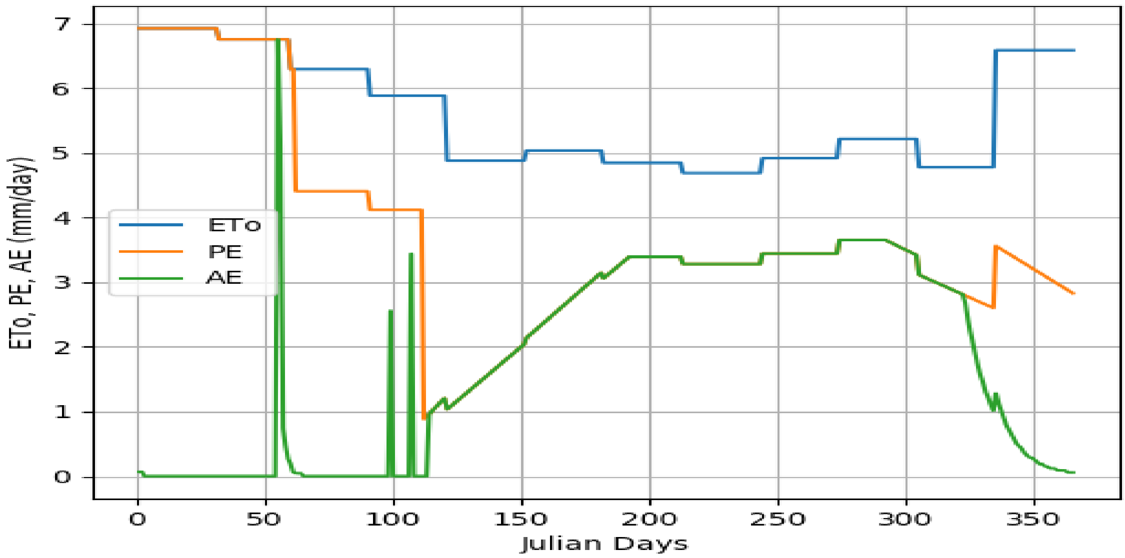


Figure 4.19: Reference evapotranspiration, potential evapotranspiration and actual evapotranspiration 2018

Figure 4.20 shows that 64.128 mm was the highest daily runoff calculated on the 254th Julian day for the year 2018.For the year 2018, the total runoff calculated was 529.343 mm. From the figure shown, a maximum daily recharge of 47.013 mm was calculated for the year 2018 and 0.94155mm minimum daily recharge was calculated by the model. A total of 165.046 mm is the yearly recharge value for the year 2018

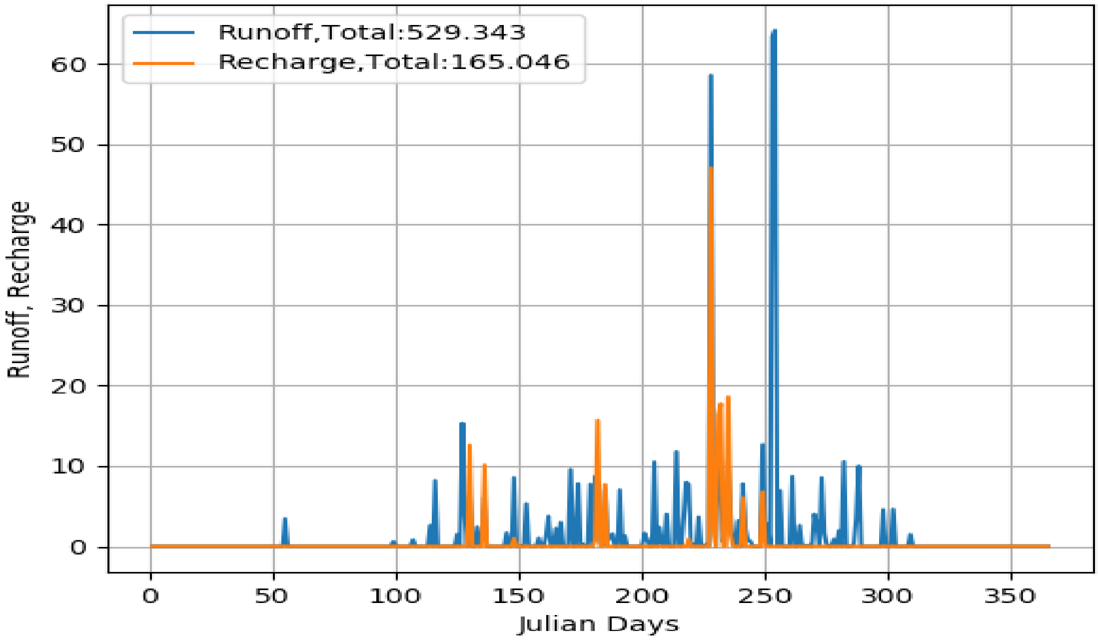


Figure 4.20: Runoff and recharge 2018

In Table 4.1, it can be seen that from the year 2008 to 2018,the rainfall was not stable in terms of intensity and this also can be found in the recharge calculated using the model. Year 2009 recorded the highest recharge of 333.35 mm and this coincided with the highest rainfall of 1665.4mm from 2008 to 2018.The lowest recharge of 39.119 mm was recorded in 2017.

Table 4.1: Annual hydrological parameters in Otukpo basin from 2008 to 2018

Year	Rainfall (mm)	Runoff (mm)	Recharge (mm)	Actual Evapotranspiration (mm/year)
2008	1608.2	935.56	142.57	586.61
2009	1665.4	698.904	333.35	689.45
2010	1211.4	488.414	178.899	600.34
2011	1449.2	785.705	90.924	628.95
2012	1493.9	640.34	188.269	721.39
2013	1287.9	541.544	137.852	664.78
2014	1248.7	497.64	154.579	652.50
2015	1001.3	322.04	91.699	644.04
2016	1379.9	598.44	190.44	647.56
2017	1062.4	496.985	38.12	583.84
2018	1298.9	529.34	165.05	660.59

CHAPTER FIVE

5.0 CONCLUSION AND RECOMMENDATIONS

5.1 Conclusion

Potential recharge has been estimated for a climate that belongs to the Kopper's Aw climate group and defined as "tropical with distinct dry seasons" using a daily Modified Soil Moisture Balance Model based on a single soil water store. Reliable estimates can only be obtained if all the physically important processes are represented satisfactorily. Soil and crop properties are determined and simulated in the model using crop coefficients and total and readily available water. Runoff coefficients are based on the current soil moisture deficit and the magnitude of the daily rainfall. Field records of runoff are required so that, by a trial-and-error procedure of adjusting the runoff coefficients, improved simulation of the runoff can be achieved. Near surface storage should be included in the model to represent the

5.2 Recommendations

Modified soil moisture balance model used in this study estimates daily runoff, SMD, and groundwater recharge. The estimation of these output parameters aided in the estimation of groundwater quantity of the Otukpo basin. The model gave the on-point irrigation requirement estimation, which is influenced by rainfall input in the model. In other words, rainfall plays a major role in the model; Soil moisture deficit , runoff, Total available water , Readily available water , and groundwater recharge are all dependent of rainfall. To further enhance the usefulness and efficacy of this modified moisture balance model, it is therefore recommended that further studies is carried out on using the model to predict or estimate the groundwater quantity for some other crops within or outside the basin for future water resources planning of the basins. This will also help in ascertaining the adequate availability of soil moisture for these crops' use and possibility of planning for irrigation regime for the next couple of years. This recommendation became necessary owing to the quantity of water being consumed daily by these crops and the need to sustain adequate provision of soil moisture for overall use of the vegetation in the region.

5.3 Contribution to Knowledge

The study has now made it possible to estimate daily groundwater recharge in any basin unlike other previous models that are monthly and annually based. The study also has been able to establish novel approach of estimating groundwater recharge in Otukpo basin to improve the agricultural yields in the basin especially Cassava, which is the predominant crop in the region. The model will also help in ascertaining the adequate availability of soil moisture for crops yield within the basin.

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