

# Estimation of evapotranspiration losses in the vadose zone using stable isotopes and chloride mass balance method

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**Abstract** This study used stable isotope and chloride data of porewater to evaluate water fluxes through the vadose zone and thereby estimate evapotranspiration water losses in parts of the Nabogo catchment of the White Volta Basin in Ghana. The overall objective was to develop a framework so that the recharge regime can be properly conceptualized for numerical hydrological modeling. Unsaturated zone material was sampled at regular intervals of 50 cm to a maximum depth of 300 cm in four different locations in the study area. Rainwater, groundwater and surface water were simultaneously sampled and analyzed for their  $\delta^{18}\text{O}$  and  $\delta^2\text{H}$  characteristics. Porewater, extracted from the unsaturated zone material was analyzed for the  $\delta^{18}\text{O}$  and  $\delta^2\text{H}$  and chloride content and profiles were created to gauge the isotopic evolution of precipitation and estimate evaporative losses at each interval of the sampling. The chloride mass balance technique was used to estimate the fraction of infiltrating water remaining at each interval of the sampling. Transpiration losses through the entire profile were then estimated. This study finds that the vertical infiltration of water through the vadose zone is dominated by piston flow and a mixture of piston and preferential flows. In the shallow subsurface (0.0–3.0 m), evaporative losses estimated from stable isotope data fall in the range of 29.3–52.4 % (322.3–576.4 mm/year) of the annual precipitation, with an average of 40 % (or 440 mm/year). Estimated vadose zone recharge at the maximum depth of

sampling ranges between 11.1 and 185 mm/year with an average of 32.9 mm/year, representing 1.1, 18.5, and 3.29 % of the annual precipitation, respectively. Estimated transpiration losses within this interval range between 29.1 % (290 mm/year) and 69.5 % (695 mm/year), with an average of 54.7 % (547 mm/year) of the annual precipitation. Transpiration losses appear to increase down the profile and apparently account for a significant percentage of water losses in the vadose zone. A significant proportion of the original precipitation is lost within the upper 300 cm (3 m) of the vadose zone. The Water Table Fluctuations method was independently used to estimate saturated zone groundwater recharge and indicates that recharge rates range between 64.65 and 151.2 mm/year with an average of 102.5 mm/year which, respectively represent 5.9, 13.7, and 9.3 % of the average annual precipitation in the area. The apparently higher estimates from the water table fluctuations method may arise from uncertainties in the specific yield values used for the vadose zone material.

**Keywords** Chloride mass balance · Evaporation · Groundwater · Isotopes · Porewater · Recharge · Transpiration

## Introduction

The use of natural tracers in hydrological processes research is conventional practice and has been copiously documented in the contemporary scientific literature. Natural and artificial tracers have been useful in determining water pathways, evolution, and age (e.g. Salem et al. 2004; Flusche et al. 2005; Mahlknecht et al. 2006; Négrel et al. 2007; Al-Gamal 2011). Isotope tracers in particular provide a unique tool for assessing groundwater recharge and

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evolutionary processes (e.g. Ortega-Guerrero 2003; Mukherjee et al. 2006), groundwater salinization (e.g. Kim et al. 2003) and groundwater interaction with surface flows (e.g. Maduabuchi et al. 2006). The stable isotope composition of porewater can reveal information about evaporation processes (e.g. Braud et al. 2009), transpiration (Wenninger et al. 2010) and downward infiltration processes (Gazis and Feng 2004). Earlier foundational researches on the characteristics of porewater stable isotopes suggest that the impacts of evaporation are highest near the surface of the soil and reduce with depth (Zimmermann et al. 1967; Barnes and Allison 1984; Brooks et al. 2010). For instance Zimmermann et al. (1967) indicate that deuterium is enriched near the surface, and depletes exponentially down the soil profile. However, the variations in stable isotope composition of porewater are mostly interrupted by other processes such as mixing with newer rainwater. Song et al. (2011) proposed that the isotope characteristics of porewater should be regarded as being the result of the impacts of evaporation and the cumulative effects of several precipitation events. This is because the entire process of infiltration of rainwater through the unsaturated zone material and the eventual percolation of the same through the saturated zone are complicated processes with several intervening processes (Cheng et al. 2014).

Two main processes have been proposed for the analyses of the evolution of rainwater through the soil profile (Cheng et al. 2014): piston flow and preferential flow. According to Zimmermann et al. (1966), piston flow occurs when porewater moves in layered form with older water being progressively pushed deeper by newer water. In preferential flow on the other hand, newer water bypasses older water through crevices created in the wake of insect burrows and other processes in the soil zone, so that newer, relatively depleted signatures can be observed at deeper horizons (Cheng et al. 2014) while relatively enriched older waters remain in the soil matrices in higher horizons. These two mechanisms of flow have been used to describe stable isotope profiles of porewater in the literature (e.g. Gazis and Feng 2004), and based on them groundwater recharge processes through the vertical infiltration and percolation of precipitation have been evaluated in several places. By critically examining the variations in the signatures of porewater down soil profiles, one can conveniently indicate the most prominent flow system prevailing in the area.

This current investigation aims at evaluating the processes of direct groundwater recharge through infiltration and percolation of rainwater through the unsaturated zone and thus assists in conceptualizing the role of the unsaturated zone for large scale numerical hydrological modeling of the White Volta basin. Specifically, this study estimates

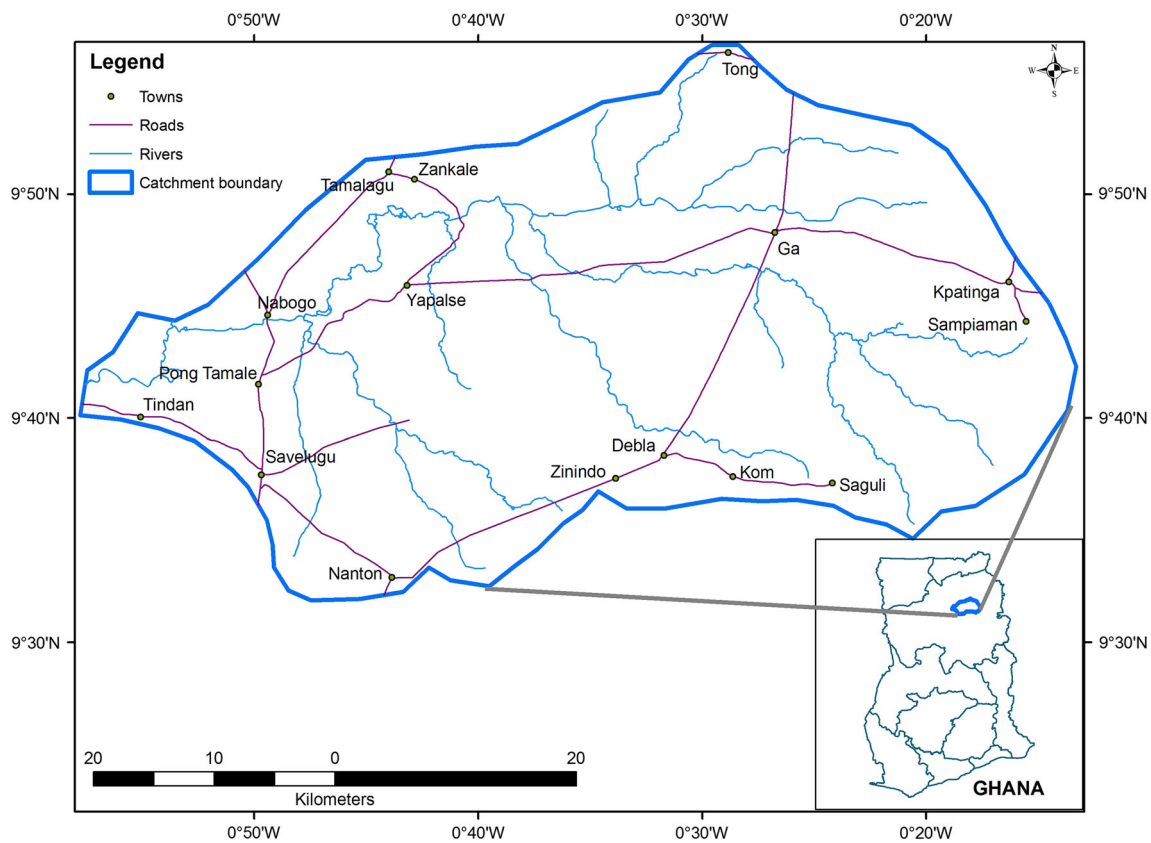
evaporation and transpiration losses of infiltrating rainwater through the unsaturated zone, and therefore provides an avenue for adequately representing the role of evapotranspiration in the water resources assessments and evaluations within the White Volta Basin and its sub-catchments. It uses stable isotope and chloride data of porewater from the unsaturated zone material to estimate changes in the fraction of the original rainwater remaining at predefined intervals through the unsaturated zone using the chloride mass balance technique.

Groundwater recharge is an important parameter for accurate assessment of groundwater resources potential and for evaluating optimal groundwater abstraction rates. The use of stable isotope tracers in studies of hydrological processes in Ghana is not new (e.g. Akiti 1980; Pelig-Ba 2009; Adomako et al. 2010; Yidana and Koffie 2014). For instance, Adomako et al. (2010) used time series soil profile isotope data monitored in parts of the Densu Basin to estimate possible groundwater recharge rates in the basin. Their study provided an understanding on the use of stable isotopes for evaluating hydrological processes in such a setting. The methodology applied in this current investigation is different from that adopted by Adomako et al. (2010) as the objectives are slightly different. The methodology applied is similar to that of Peng et al. (2011), Dogramaci et al. (2012, 2015), and Skrzypek et al. (2015), and is a derivative of the Gibson and Reid (2010) methodology for evaluating unsaturated zone processes. Evaluating groundwater recharge processes through the analyses of stable isotope data provides a convenient way of estimating vertical groundwater inputs and the impacts of evapotranspiration on groundwater budgets through time.

## The study area

### Location, geography, and vegetation

The study area (Fig. 1) is one of the nine sub-catchments of the White Volta in Northern Ghana. The area lies within the tropical Continental or Interior Savannah Climatic Zone. It experiences a single rainfall season, which starts from May and ends in October. Historical data (Dickson and Benneh 1995) suggest that the annual rainfall in the area peaks between July and August and declines sharply between October and November. However, over the past few years, the patterns of rainfall in the area and the rest of the northern parts of the country have been erratic and difficult to predict. Sometimes the rainy season does not set in until July and August. The area experiences occasional storms, which have implications for base soil erosion depending on its frequency and intensity especially when



**Fig. 1** A map of the study area showing some of the communities sampled and the drainage network

they occur at the end of the dry season. Mean annual rainfall ranges between 950 mm and 1200 mm (Dickson and Benneh 1995). The dry season is characterized by Hamattan winds. These winds, which blow across the Sahara desert, are warm and dry causing significant daily temperatures and causing the soil to lose moisture rapidly. Maximum day temperatures are recorded between March and April of about 45 °C while minimum night temperatures of about 12 °C have been recorded in December-January. Relative humidity is high during the rainy season (65–85 %) but may fall to as low as 20 % during the dry season. Daily sunshine duration is about 7 h except in the rainy days of June to September when it ranges from 6.9 to 4.7 h (Dickson and Benneh 1995).

The drainage network in the terrain is dendritic and is quite dense especially in the rainy (wet) season when some parts of the basin are flooded due to heavy rains in the northern parts that extend to Burkina Faso. The network of streams is largely made of tributaries of the White Volta which comprises the Nabogo, Sissili and the Kulpawn rivers. Some of the streams are ephemeral and dry out during the dry season due to high temperatures and low humidities. The natural vegetation of the district is classified as the Guinea Savannah Woodland and is composed of

short trees of varying sizes and density, growing over a dispersed cover of perennial grasses and shrubs (Dickson and Benneh 1995). The prevailing weather conditions, the relief features and soil texture foster water logged conditions in the rainy season and cakes up in the dry season. Consistent with these conditions, characteristically hardy tree vegetation capable of withstanding long dry spells are prevalent in the area. They have long tap roots which reach down to the shallow aquifer zone and are therefore important in assessing the impacts of evapotranspiration on groundwater resources availability and sustainability in the terrain.

Detailed accounts of the vegetation types in Ghana are contained in the Bagamsah (2005). The original vegetation cover of the area was one of moist Guinea Savannah. However, through years of extensive farming, bush burning, and active deforestation, the original vegetation has been reduced to shrubs, grasses, and scattered trees. The most dominant tree types are the drought resistant savannah type trees including the neem tree, shearnut trees, baobab trees amongst others. These tree types by their nature, have elaborate and long root systems which go several meters through the vadose zone and into the saturated zone in places. The remnant vegetation cover rarely

forms a close canopy as the trees are scattered, and the area is dominated by grasses and shrubs. The traditional savannah woodland with light canopy which characterized the greater part of the area has gradually been replaced with a type of open savannah with scattered trees that rarely form closed canopy (WRC 2006).

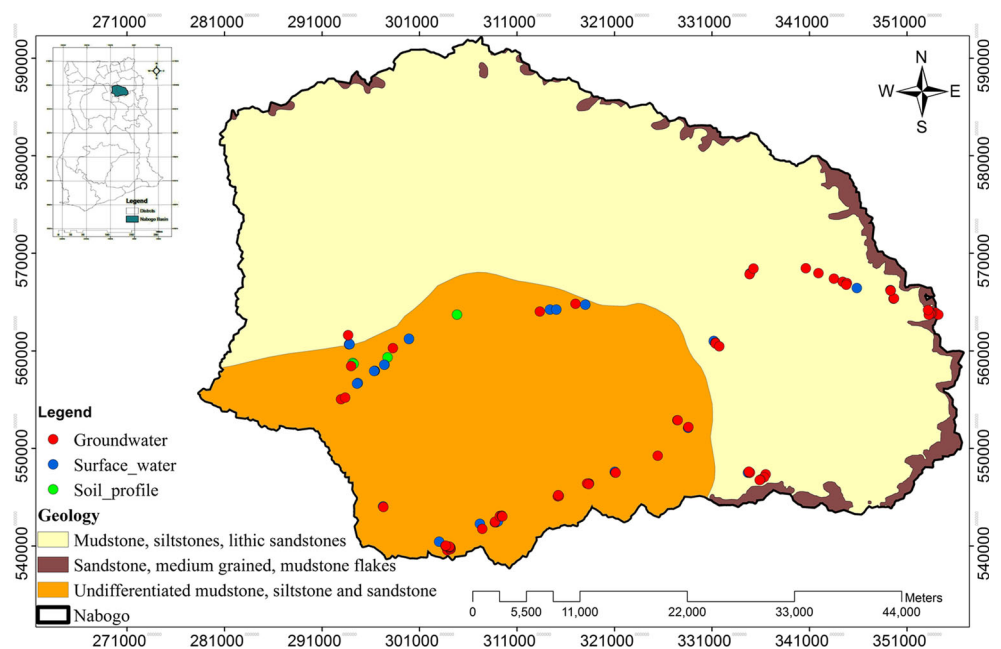
### Geological, hydrogeology, and Soil

The study area is underlain by rocks of the Upper and Middle Voltaian (Kesse 1985). However, in the study area, the mudstones and siltstones are the most predominant. The sandstones largely outcrop in the northern eastern fringes of the terrain (Fig. 2). The siltstones are abundant in the southern parts of the terrain and are highly consolidated due to high degrees of compression attending the Pan African orogenic event (Kesse 1985). Exposures of this formation are common on the sides of the major roads in the area and are also available as massive boulders in the area. Where these rocks undergo significant chemical and physical weathering, thick lateritic layers are common. These lateritic hardpans are common features of the lithology in the area and much of Northern Ghana. Where they are predominant, vertical infiltration and percolation of precipitation is very much restricted, leading to reduced recharge. Depending on the nature and composition of the parent rock material and the degree of chemical weathering, clayey to silty clay soils are common. The percentage of clay in the soils of the unsaturated zone and its spatial variability is believed to be responsible for the reduced recharge rates in parts of the Voltaian (Yidana and Koffie 2014).

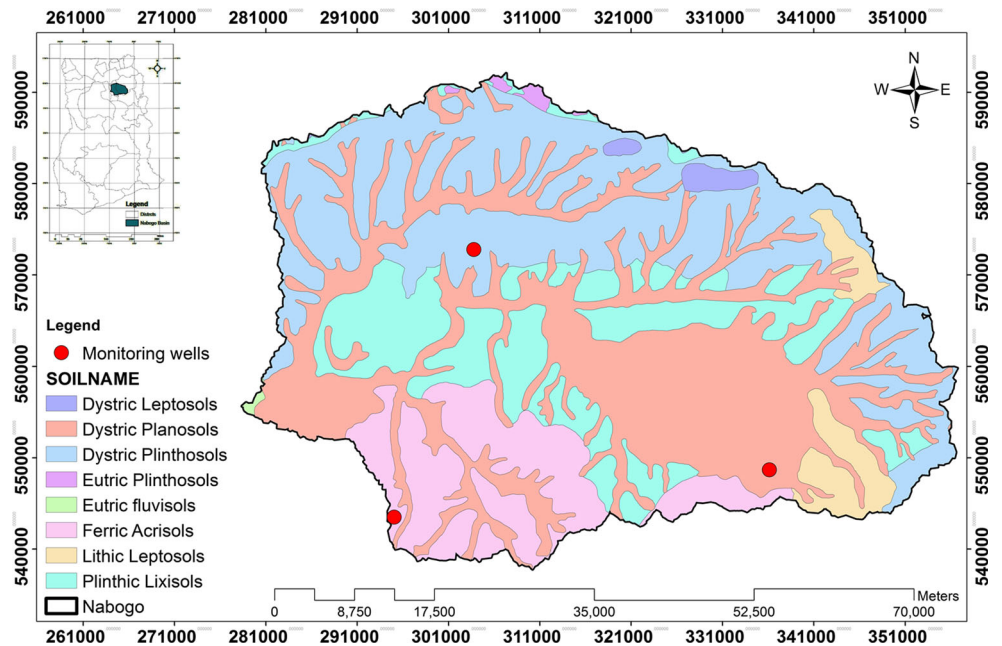
The hydrogeology and groundwater conditions of the Voltaian rocks are contained in previous reports and publications (e.g. Gill 1969; Acheampong 1996; Acheampong and Hess 1998; Yidana et al. 2008). As the rocks have been slightly metamorphosed, primary permeabilities are virtually absent and the groundwater potential and general hydrogeology of the rocks are based on the presence of secondary permeabilities in the form of fractures and weathered zones. Where the weathering is intense and the clay content is low, the rocks serve as prolific aquifers and are important in the local water economies of the largely rural populations in the area. Most of the shallow wells in the study area are completed in weathered zones of variable spatial extents.

The soils of Northern Ghana and the Volta Basin are discussed in Obeng (1967). The soil map of the study area is shown in Fig. 3. The area is underlain by mostly clay rich dystric planosols which are pervasive in the entire domain, imposing semi-confining conditions in places. Through the action of the main drainage network in the area, these soils appear to form a dendritic network due to erosion. In most parts of the study area, they are underlain by thick layers of the lateritic hardpan. The planosols are the most prevalent in the area, and due to their high clay content, waterlog conditions are common in most parts of the basin especially during the rainy/wet season. The plinthosols are largely encountered in the northern parts of the terrain, overlying the mudstones and siltstones. They are believed to be the weathered residue of the underlying rocks and contain variable quantities of clay and iron rich content and silica. It hardens in most places into impermeable ironstone

**Fig. 2** Geological map of the study area showing the sampled locations



**Fig. 3** Soil map of the area showing the locations of the monitored boreholes for the WTF estimations



imposing semi-confining conditions on the underlying subsurface. There are also the plinthic lexisols in the central parts of the domain. They contain variable quantities of clay and are suited for the cultivation of mainly tubers. The next most abundant soils in the area are the ferric acrisols which are prevalent in the south of the terrain. Due to the inherent clay content, the soils in the area present with low vertical hydraulic conductivities and therefore retard vertical infiltration of precipitation. The hydraulic conductivities of the soils are spatially variable due to the spatial variability in the clay content (Nyarko 2007).

**Materials and methods**

**Sampling and analysis**

A total of 125 water samples were collected during a field trip to the Nabogo area in December 2012 over a period of 1 week. Four sites were selected for sampling the unsaturated zone material at regular intervals of 0.50 m up to 3 m within the Nabogo sub-catchment of the White Volta Basin. A manual auger was used for the purpose of drilling out and sampling the material. About 500–1000 g of the bulk soil samples were collected in polyethylene bags, tightly sealed and packed in an ice chest to prevent moisture loss due to evaporation. Nineteen surface water samples were collected from perennial streams, rivers and ponds, for which the analysis of  $\delta^{18}\text{O}$  and  $\delta^2\text{H}$  was conducted. Also, 77 groundwater samples were taken (Fig. 2) in the area for the purpose of analyzing for their

stable isotope contents. In collecting water samples, standard protocols as described by the United States Geological Survey (USGS 2006) were strictly observed. The average depth of the boreholes sampled is 30 m. Groundwater samples were always collected after purging the well of stagnant water around the pump parts especially the nozzle. The typical physical parameters of pH, Electrical Conductivity (EC), Total Dissolved Solids (TDS), and Temperature were measured in situ in the case of water samples. Samples for the isotope analyses were collected in sterilized polyethylene bottles and filled to the very brim of the sampling container. Samples for isotope analysis were unfiltered. Particular care was taken to avoid trapping air bubbles in the bottles. The samples were stored away from light in a cool dry ice chest and then transported to the laboratory for analyses. Twelve rainwater samples were earlier taken in Tamale from three major events during the rainy season and towards the end of the rainy season, in 2012 to provide precipitation isotope data for this study. In addition, historical rainwater data for other parts of the Voltaian basin were acquired from the Department of Earth Science University of Ghana, to assist in developing the Local Meteoric Water Line, LMWL, for the entire area.

Porewater was extracted from the bulk soil samples for the analysis of  $\delta^{18}\text{O}$ , and  $\delta^2\text{H}$ . The detailed procedure is described by Araguáus-Araguáus et al. (1995) and Adomako et al. (2010). A vacuum extraction line consisting of three parallel units with cryogenic collection of the extracted soil water was used and the extracted water was stored in 30 mL glass bottles and tightly sealed to prevent evaporation.

The Liquid Water Isotope Analyzer (LWIA-24d) was used to determine the isotope ( $\delta^{18}\text{O}$  and  $\delta^2\text{H}$ ) composition of extracted porewater. Standard procedures as described for sample preparation and analysis (Pelig-Ba 2009), were undertaken before the start of each analysis. This was done to ensure the quality and reproducibility of stable isotope data. Both  $^{18}\text{O}$  and  $^2\text{H}$  contents were determined relative to in-house standards that were calibrated using Vienna-Standard Mean Ocean Water (V-SMOW). The data were normalized and reported in the delta notation (Coplen 1988):

$$\delta[\text{‰}] = \left( \frac{R_S}{R_{V\text{-SMOW}}} - 1 \right) \times 1000 \quad (1)$$

where  $R_S$  represents either the  $^{18}\text{O}/^{16}\text{O}$  or the  $^2\text{H}/^1\text{H}$  ratio of the sample, and  $R_{V\text{-SMOW}}$  is the  $^{18}\text{O}/^{16}\text{O}$  or the  $^2\text{H}/^1\text{H}$  ratio of the V-SMOW. The analytical reproducibility is  $\pm 0.1$  and  $\pm 1.0$  ‰ for  $\delta^{18}\text{O}$  and  $\delta^2\text{H}$ .

In the determination of chloride ( $\text{Cl}^-$ ) concentration in porewater, a simple titration was employed in accordance with the procedure outlined by Skoog et al. (1996). The solid reagents used were  $\text{NaCl}$ ,  $\text{CaCO}_3$ ,  $\text{NaHCO}_3$ ,  $\text{K}_2\text{CrO}_4$  and  $\text{KNO}_3$ . A 5 %  $\text{K}_2\text{CrO}_4$  indicator was initially prepared by dissolving 1.0 g of  $\text{K}_2\text{CrO}_4$  in 20 mL distilled water. Details of the procedure are outlined in Skoog et al. (1996).

### Estimation of evaporation rates

Analysis of the effect of evaporation on the annual water budgets in the basin is an important aspect of sustainable water resources management for productive uses. In evaluating groundwater recharge from direct infiltration and percolation of rainwater, it is important to assess the fraction of precipitation that returns to the atmosphere in the process. Since the intervening material between the surface and the saturated zone is variable in thickness and content in space, the fraction of precipitation that returns to the atmosphere during infiltration is expected to be variable in space. Estimates were made on evaporation rates from surface flows in the area and recharging groundwater. The procedure used is akin to that proposed by Craig and Gordon (1965) and adopted in several recent researches. The fraction of precipitation lost to evaporation,  $f$ , is given by Eq. 2 (Gibson and Reid 2010; Dogramaci et al. 2012).

$$f = 1 - ((\delta_L - \delta^*) / (\delta_p - \delta^*))^{1/m} \quad (2)$$

$\delta_p$  is the isotopic signature of the source water for the streams and ponds or groundwater or porewater sampled;  $\delta_L$  is the isotopic composition of the surface water, groundwater or porewater whose evaporation rate is being estimated; the indices,  $m$  and  $\delta^*$  are, respectively given by

Eqs. 3 and 4 (Welhan and Fritz 1977; Gibson 2002; Dogramaci et al. 2012, 2015) for non-steady conditions.

$$m = \frac{(h - \varepsilon/1000)}{(1 - h + \varepsilon_k/1000)} \quad (3)$$

$$\delta^* = (h\delta_A + \varepsilon)/(h - \varepsilon/1000) \quad (4)$$

$h$ ,  $\delta_A$  are respectively the relative humidity and isotopic signature of ambient air/vapor. Gonfiantini (1986) defined the isotopic enrichment arising from kinetic isotope fractionation ( $\varepsilon_k$ ) and total isotope enrichment factor ( $\varepsilon$ ). Kinetic fractionation factors for  $^{18}\text{O}$  and  $^2\text{H}$  are, respectively defined by Eqs. 5 and 6 (Gonfiantini 1986).

$$\varepsilon_k^{18}\text{O} = 14.2(1 - h) \text{‰} \quad (5)$$

$$\varepsilon_k^2\text{H} = 12.5(1 - h) \text{‰} \quad (6)$$

The total fractionation factor is then given by Eq. 7 (Dogramaci et al. 2015).

$$\varepsilon = \varepsilon_{\text{eq}} + \varepsilon_k \quad (7)$$

where  $\varepsilon_{\text{eq}} = 1000(1 - \alpha_{w-v}^{-1})$

$\alpha_{w-v}^{-1}$  is estimated for  $^{18}\text{O}$  and  $^2\text{H}$  respectively from Eqs. 8 and 9 (Leibundgut et al. 2009)

$$10^3 \ln \alpha_{w-v}^{18}\text{O} = -2.0667 - 0.4156(10^3/T) + 1.137(10^6/T^2) \quad (8)$$

$$10^3 \ln \alpha_{w-v}^2\text{H} = 52.612 - 76.248(10^3/T) + 24.844(10^6/T^2) \quad (9)$$

$\alpha_{w-v}$  is the equilibrium fractionation factor between vapor and water for oxygen and hydrogen isotopes, respectively. It is a function of temperature ( $T$ ) in Kelvins (Criss 1999) as indicated in Eqs. 8 and 9.

$\delta_A$  is a difficult parameter to measure. In most practical cases, it is estimated from other parameters. There are at least three options for determining  $\delta_A$ . The best approach is to direct measurement of  $\delta^{18}\text{O}$  and  $\delta^2\text{H}$  of ambient air. However, as indicated by Skrzypek et al. (2015), such measurements are a global challenge due to logistical constraints. It was therefore not possible to effect direct measurements in the current study. The second and third approaches involve estimation of these parameters from other parameters which are quite easier to determine. This study benefited from the latter approaches. Peng et al. (2011), Dogramaci et al. (2012, 2015), and Skrzypek et al. (2015) suggested a relationship between water vapor isotopic signature and rainwater ( $\delta_{\text{rain}}$ ) in equilibrium with it (Eq. 10). However, whereas Peng et al. (2011) suggest that the isotopic

signature of the first rains are most likely to reflect the signatures of rainwater in equilibrium with water vapor, Dogramaci et al. (2012, 2015) and Skrzypek et al. (2015) are of the belief that the best approach is to monitor the signatures of precipitation over a period of time and then take an average value over that period. The latter approach was considered more plausible and was therefore adopted in this study. Isotope signatures of rainwater sampled over the entire year in 1994, and rainwater data taken in the year 2012 were averaged to provide input for  $\delta_{rain}$  in Eq. 10.

$$\delta_A \cong \delta_{rain} - 10^3(\alpha_{w-v} - 1) \tag{10}$$

Average isotopic signatures of rainwater samples taken in the years 1994 and 2012 presented  $\delta^{18}O$  and  $\delta^2H$  values of  $-6.39 \text{ ‰}$  and  $-41.73 \text{ ‰}$ , respectively. These were used as inputs for  $\delta_{rain}$  in Eq. 10 to determine the  $\delta_A$  values for  $^{18}O$  and  $^2H$  respectively.

An average ambient temperature of  $28 \text{ }^\circ\text{C}$  ( $301 \text{ K}$ ) was determined from temperature records, and average relative humidity of  $65 \text{ ‰}$  ( $0.65$ ) was used to estimate  $\alpha_{w-v}^{-1}$  for  $^{18}O$  and  $^2H$  (Eqs. 9 and 10). As a result the estimated  $\delta^{18}O$  and  $\delta^2H$  of the vapor source of precipitation in the area are, respectively,  $-15.53$  and  $-118.04 \text{ ‰}$ . The isotopic signature of the source precipitation for porewater, stream water, and groundwater were, respectively determined by using the intercept of the local meteoric water line and the respective evaporation lines of porewater, stream water, and groundwater.

### Recharge Estimates through Chloride Mass Balance (CMB)

The chloride mass balance (CMB) method has been used extensively for estimating groundwater recharge in arid and semi-arid areas worldwide (Gee et al. 2004; Scanlon et al. 2002). Erikson and Khunakasem (1969) were among the first to estimate recharge using the CMB methodology. They applied it to estimate groundwater recharge on the coastal plain of Israel. Groundwater recharge estimates through the CMB method relies on measurements of porewater chloride content in the unsaturated zone or saturated zone chloride content, and average chloride concentration of precipitation over a period of time. The mathematical formulation of the CMB methodology is presented in Eq. 11 (Wood 1999).

$$q = \frac{P \times Cl_p}{Cl_{gw}} \tag{11}$$

where  $q$  is ground water recharge flux or recharge (mm/year),  $P$  is average annual depth of precipitation (mm/year),  $Cl_p$  is average precipitation weighted chloride concentrations (mg/L) and  $Cl_{gw}$  is the average chloride concentration in ground water (mg/L). In order to estimate the progressive changes in recharge through the soil profile, Eq. 11 was modified so that porewater chloride content  $Cl_{sw}$  was used in place of groundwater chloride content  $Cl_{gw}$ . In this case, the average chloride concentration in groundwater ( $Cl_{gw}$ ) was replaced with the average chloride concentration in porewater ( $Cl_{sw}$ ) (Eq. 12) (Ting et al. 1998). In this case, the computed recharge at every stage down the profile was diffuse recharge.

$$q = \frac{P \times Cl_p}{Cl_{sw}} \tag{12}$$

The chloride concentration in porewater ( $Cl_{sw}$ ) at every drilled hole was determined and  $Cl_p$  was determined from average rainwater chloride concentration analyzed by scientists at the Ghana Atomic Energy Commission for the Northern part of Ghana for a number of rainy events. The average depth of precipitation,  $P$ , for the study area was obtained from raw data provided by the Ghana Meteorological Agency (GMA) for a period of 40 years on request. The application of CMB method in estimating recharge is based on some key assumptions: (1) water movement in the vadose zone is downward under the influence of gravity and flow is one-dimensional (piston flow); (2) precipitation has been constant throughout time; (3) chloride concentration in precipitation has been constant throughout time; (4) precipitation is the sole source of chloride entering the ground (Allison and Hughes 1978); (5) precipitation is the sole source of recharge (Scanlon 2002); and (6) land-use patterns are stable. In addition, it must be implied that all of the precipitation infiltrates the soil (Wood 1999; Dettinger 1989). Practically, it is difficult to meet most of the assumptions under this methodology as stated. For instance the assumption that all the precipitation infiltrates the soil is far reaching and hardly applies anywhere. However, where surface runoff is low as is the case in the study area, this assumption is approximated. Changing levels of industrialization with corresponding changes in atmospheric chemistry will certainly have an effect on atmospheric chloride content and the chloride content of precipitation through time. In the study area, this is not a challenge as the level industrialization through time has not been significant to effect such changes in atmospheric chloride content as to invalidate the relevant assumptions. Oboubie (2008) undertook samples of precipitation events over a period of time in an area north of the current study area. The analyses did indicate an inverse relationship

between rainfall amounts and chloride deposition. Oboubie (2008) observed that chloride deposition in the area generally varies between 0.2 and 2.1 mg/L with an aerial average of about 0.8 mg/L. Similar observations were made by other researchers in other parts of the White Volta Basin (e.g. Nkotagu (1996) and Sandwidi (2007)). It has therefore been noted that chloride deposition in the region, compared to other areas of similar climatic conditions, is quite stable. For the estimation of vadose zone recharge (the fraction of infiltrating water remaining at every interval of the profile sampling) in the current study, an average rainwater chloride concentration of 1.39 mg/L was used. This was obtained by taking an average of the rainwater chloride content of the rain events sampled in 2012, and historical rainwater data from Ghana Atomic Energy Commission. However, the main challenge in the current study was the inadequate quantification of the proportion of the groundwater chloride which is attributed to the lithology in order to effect appropriate corrections to groundwater chloride content. To minimize the effect of the local lithology on the groundwater chloride in this study, areas with obviously very high chloride values were excluded, so that groundwater recharge in those areas were not overly underestimated. Conversely, in areas where groundwater chloride content was obviously too low compared to precipitation chloride content, such data were not used in the estimation. These methods were applied to minimize the errors associated with the far reaching assumptions of this methodology.

The CMB technique has been used largely in the tropics and in arid climates for estimating groundwater recharge rates, and has been noted to produce reliable estimates to provide inputs into decision support systems for evaluating groundwater resources sustenance and efficient management. For instance Ting et al. (1998) used the CMB methodology to estimate diffuse groundwater recharge in parts of Taiwan. They used porewater (soil moisture) chloride concentrations in place of groundwater chloride content. In such a case, the estimates of recharge were regarded as approximate values as they represent water still in transit in the unsaturated zone. The methodology employed in their study was similar to the approach used in this current investigation. However, the difference is that in the current investigation, the estimates are made for different depth intervals to track the percentage losses of water in transit through the unsaturated zone. Carrier et al. (2008) used a similar approach over parts of Northern Ghana to estimate groundwater recharge rates. In their study, porewater samples from the unsaturated zone were analyzed for chloride content and then applied the CMB methodology to estimate diffuse recharge rates. Tyner et al. (1999) had earlier applied a similar approach to estimate groundwater recharge rates under areas where KCl was used to fertilize crops. Other

successful applications of the CMB method in estimating recharge and water fluxes include Aishlin (2006), Ping et al. (2014), Shaghravani et al. (2015) amongst others.

### Recharge estimates through water table fluctuation (WTF)

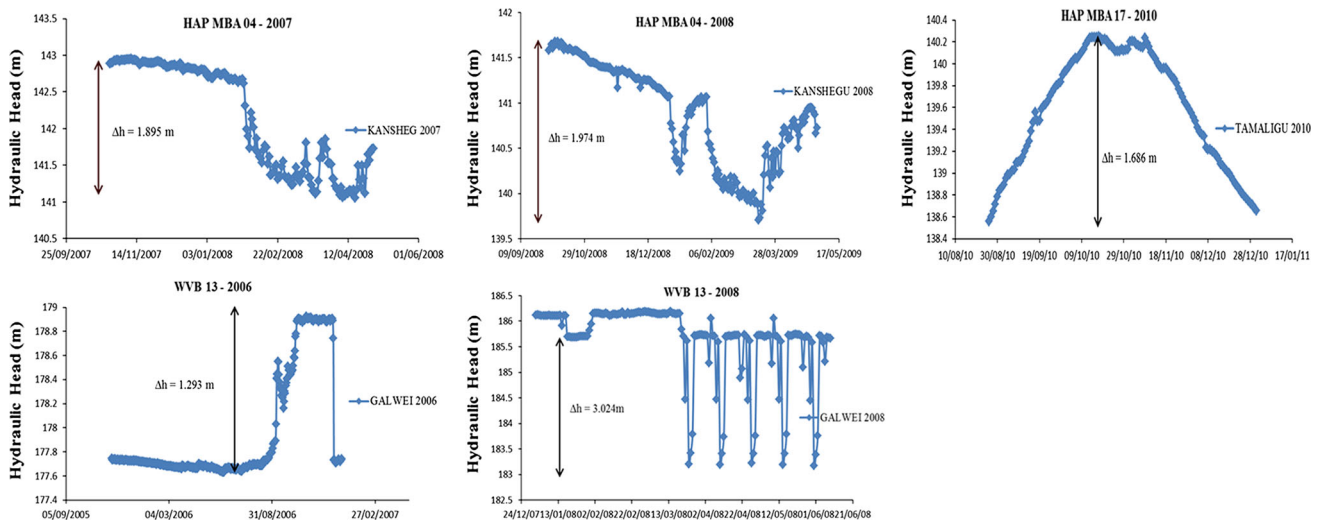
The water table fluctuation method (WTF) is one of the most widely used techniques for estimating groundwater recharge over a wide variety of climatic conditions (Scanlon et al. 2002; Healy and Cook 2002; Hall and Risser 1993). The use of the method requires knowledge of specific yield and changes in groundwater levels over time. Healy and Cook (2002) have suggested that the wide use of this method could be attributed to the abundance of available water level data and the simplicity of estimating recharge rates from temporal fluctuations or spatial patterns of water levels. The WTF method is based on the assumption that rises in water levels in unconfined aquifers are due to recharge water arriving at the water table, and that all other components of the groundwater budget, including lateral flow, are zero during the recharge period (Scanlon et al. 2002; Healy and Cook 2002). The groundwater recharge rate can be estimated as the product of the water level rise and the specific yield of the aquifer material. The recharge can be expressed as (Eq. 13):

$$R = S_y dh/dt = S_y \Delta h / \Delta t \quad (13)$$

where  $R$  is groundwater recharge (L/T);  $S_y$  is specific yield (dimensionless);  $\Delta h$  is peak rise in water level attributed to the recharge period (mm); and  $\Delta t$  is the recharge period. The aquifers in the area are not perfectly confined but semi-confined. As such the lower range of specific yield values for the lithology were used in the estimation.

Groundwater monitoring data from three wells in the area were used to estimate groundwater recharge in the area. The locations of the monitoring wells are indicated in Fig. 3, but Fig. 4 presents the monitored data from the three wells (HAP 04, HAP 17, and WVB 13). As per the statistics on specific yield proposed by Johnson (1967), the range of values for the typical lithology of the study area is 0.02–0.08. An average specific yield of 0.05 was then used to estimate recharge within the study area. Specific yield values depend largely on the grain size, shape, distribution of pores and their interconnectivity and the nature of compaction of the strata in question (Gupta and Gupta 1999). Estimates of groundwater recharge through the WTF method is intended to provide a baseline for comparing the progressive changes in recharge through the porewater data and assist in evaluating the fraction of the infiltrating water that would have been lost between the maximum depth of the soil profile sampling and the saturated zone. The deficit would indicate the significance of evapotranspiration rates





**Fig. 4** Time series fluctuations in the hydraulic head data of three wells used to estimate groundwater recharge in the study area

below 2.5 m depth and would suggest that the extinction depth of evapotranspiration is below 2.5 m.

The water table fluctuations method is particularly suited for unconfined conditions (Healy and Cook 2002). Saghravani et al. (2015) compared estimates of groundwater recharge rates from the WTF method and the CMB methodology. Estimates from the CMB methodology were apparently higher than those of the WTF method. The authors used chloride data of porewater extracted from unsaturated zone material instead of groundwater chloride data, and attributed the difference between the WTF and CMB estimates to anion exchange activities which affected the conservative character of the chloride ion. The estimates, however, provided reasonable values to assist in decision making. It is always impossible to perfectly meet all the conditions and assumptions of any methodology for estimating any of the hydrological parameters needed for basin-wide water resources assessments. Even in the case of the WTF method, significant uncertainties can arise from the specific yield values used in the estimations. This was the point underscored by Jassas and Merkel (2014) when they assessed estimates of groundwater recharge from the WTF methodology against similar estimates from streamflow recession data. Although the account of the hydrogeology did not indicate pure unconfined conditions, the WTF estimates were not statistically different from those of the hydrograph technique.

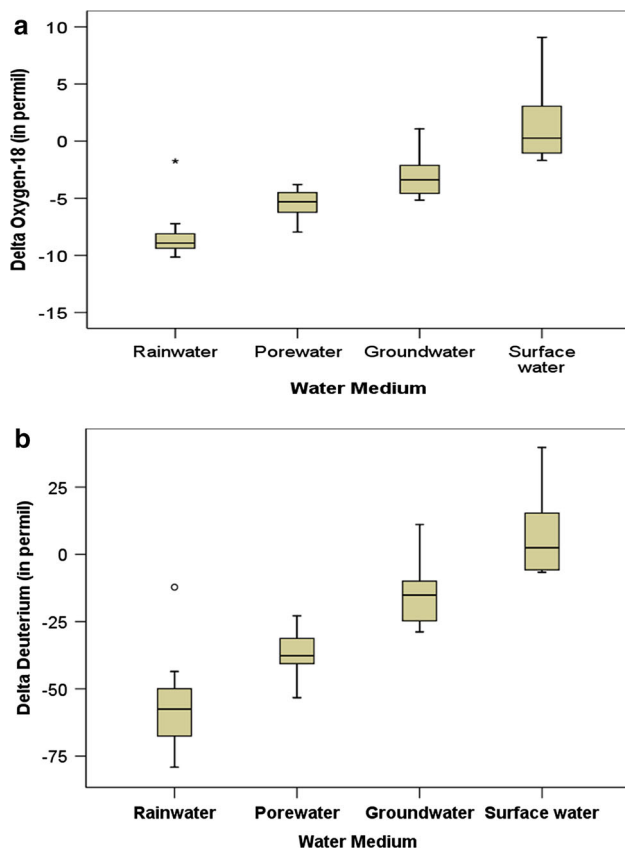
In the current study area, the aquifer is largely unconfined but infiltration rates are spatially variable due to the variabilities in the nature of the material of the vadose zone. In that case, the presumption is that the appreciation in groundwater levels results from direct infiltration of rainwater. The reasonableness of the recharge estimates will be evaluated against estimates using other techniques within the larger White Volta Basin. However, in the

application of the WTF method, unconfined conditions were assumed, with low specific yield values to account for the fact that the material is semi-confined.

## Results and discussions

### General trends in isotope content

The statistical summary of stable isotope content of surface water, porewater, rain water and groundwater in the study area are presented in the forms of box-and-whisker plots (Fig. 5a, b). Obviously the isotope data of surface water are the most enriched in terms of the heavier isotopes of oxygen and hydrogen, whereas those of porewater, rain water and ground water are depleted with rain water exhibiting the most depleted datasets. It is also obvious that rain water and surface water samples display the highest variability in terms of isotope content. The spatial variability in prevailing conditions as well as the amount of precipitation at the sampled locations may have resulted in the observed variability in the rainwater data. The isotopic signature of precipitation is amenable to a wide variety of factors which include the humidity, temperature, latitude, altitude and amount of precipitation (Gonfiantini 1986; Leibundgut et al. 2009; Dogramaci et al. 2012). High temperatures and low relative humidities particularly lead to enriched signatures of the heavier stable isotopes relative to the lighter ones due to fractionation processes that accompany evaporation of raindrops. In lower latitudes, due to high temperatures, rainwater is generally much more enriched in the heavier isotopes, compared to precipitation in higher latitude regions where temperatures are comparatively lower. The extent of dispersion in the isotope data for the rainwater samples is low, probably due to the fact



**Fig. 5** **a** Box-and-Whisker plots showing the statistical summaries of  $\delta^{18}\text{O}$  data of the media sampled for this study. **b** Box-and-Whisker plots showing the statistical summaries of the  $\delta^2\text{H}$  data of the media sampled for this study

that the impacts of the major factors responsible for the isotopic variations are less pronounced over the period of the sampling. The effect of altitude is particularly less likely as there is hardly any significant variation in the surface topography in the area. It is also obvious that the dry season rainwater samples present more enriched signatures and in some cases present signatures similar to some of the surface water data. This is due to the considerably low relative humidities during the dry season and low rainfall amounts. Leibundgut et al. (2009) discuss the amount effect on isotopic composition of precipitation. Generally, the lower the amount, the more enriched the isotopic signatures are, and vice versa for heavier amounts. However, the range of variation in the rainwater data used in this study is largely statistically consistent with data presented by previous workers in other terrains in Ghana (e.g. Akiti 1980; Pelig-Ba 2009).

The groundwater samples produced generally homogeneous datasets of the heavier isotopes (Fig. 5a, b) although they are relatively much more enriched than the rainwater data. The seemingly low dispersion in the dataset of the

groundwater samples may be due to the fact that the source precipitations are similar in signature and there is little spatial variation in the changes that occur as precipitation transits through the unsaturated zone to the saturated zone. The surface water data on the other hand, exhibits a wide range of variability, reflecting the wide variety of surface water sources sampled and the sources of recharge of these streams. The most enriched signatures are noted amongst the surface water samples due to the level of exposure and the impacts of high dry season temperatures and low relative humidities on isotopic fractionation accompanying evaporation of surface water. These conditions fuel kinetic and equilibrium fractionation processes which impact on the isotope characteristics of the residual water.

### Characteristics and inter-relationship amongst the various water bodies

Bi-plots of  $\delta^{18}\text{O}$  against  $\delta^2\text{H}$  (Fig. 6) were developed to determine the regression models for the local precipitation (Local Meteoric Water Line, LMWL), local groundwater (Local Groundwater Line, LGWL) and local surface water (Local Surface Water Line, LSWL) for the Nabogo area. The Global Meteoric Water Line (GMWL) (Craig 1961) has often been used as the baseline for comparison of recent waters and the characteristics of various reservoirs. The GMWL is defined by Eq. 14.

$$\delta^2\text{H} = 8\delta^{18}\text{O} + 10 \quad (14)$$

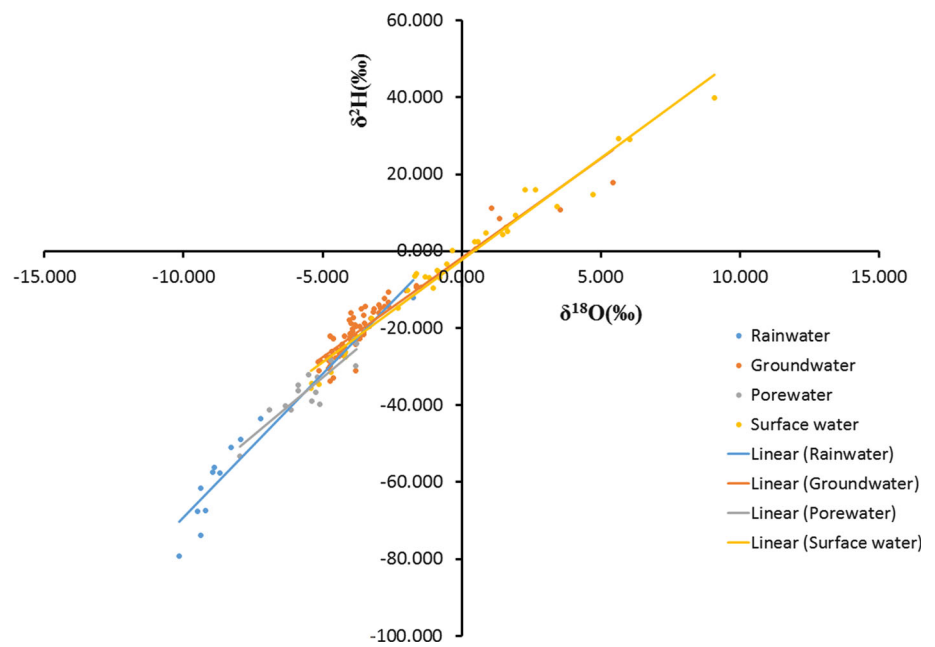
Local variations in temperature, altitude, latitude, the period of sampling, relative humidity amongst others may result in a LMWL that is different from the GMWL (Leibundgut et al. 2009). In such a case, the nature and characteristics of the local precipitation are discussed in relation to the GMWL. A LMWL developed from rainwater data is presented in Eq. 15. With respect to the slope, Eq. 15 bears close similarity to results of previous researches (Acheampong and Hess 2000; Pelig-Ba 2009; Yidana and Koffie 2014; Yidana 2013) within the larger Voltaian basin. The intercept of the current LMWL (Eq. 15) is statistically different from that of Pelig-Ba (2009) (Eq. 16), but similar to that of Yidana (2013) (Eq. 17). The apparent dissimilarity of the intercept in this study compared to that determined in Pelig-Ba (2009) may be due to the effects of the season and spatial distribution of the samples.

$$\delta^2\text{H} = 7.5\delta^{18}\text{O} + 5.53 \quad (15)$$

$$\delta^2\text{H} = 6.9\delta^{18}\text{O} - 0.19 \quad (16)$$

$$\delta^2\text{H} = 7.3\delta^{18}\text{O} + 4.77 \quad (17)$$

**Fig. 6** Biplots showing the relationship amongst all the water bodies sampled



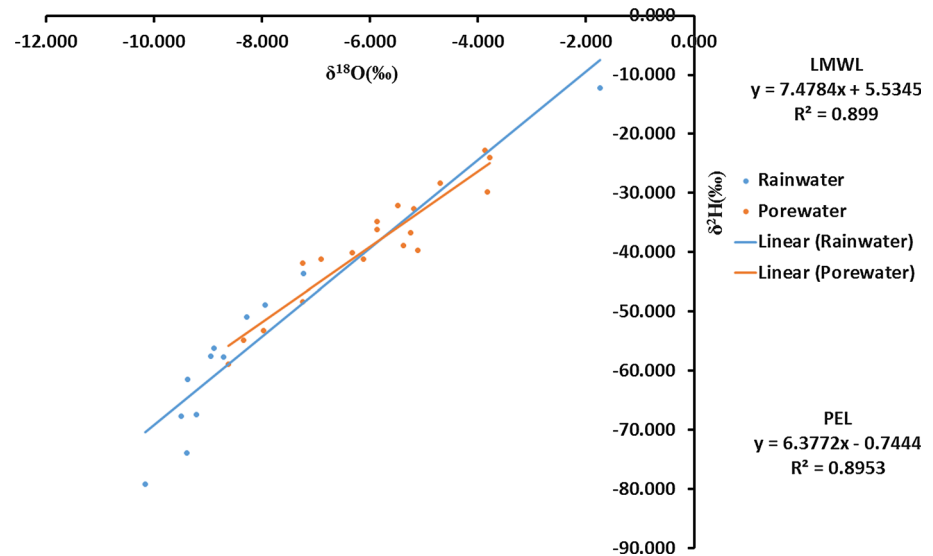
The slope of the current LMWL and those developed by previous researchers are lower than that of the GMWL (Eq. 14). This is quite consistent with the local conditions of high annual temperatures and persistently low relative humidities. Relative humidity data compiled from a monitoring station in Tamale as part of this study for the 1983–2013 period suggests that this parameter hardly reaches 100 % even during rainy events. Gonfiantini (1986) suggests that when the relative humidity is less than 100 % during precipitation, the slope of a  $\delta D$ - $\delta^{18}O$  line developed from such precipitation is  $5 \pm 2$  as obtained in the current research. Thus the local precipitation is slightly more enriched compared to the GMWL.

**Evaluation of water fluxes through the vadose zone**

By comparing porewater isotope composition to that of the rainwater in Fig. 7, the trend of isotope values suggests that the porewater originated from recent precipitation and the fractionation controlled by evaporation during infiltration caused the  $\delta^{18}O$  and  $\delta^2H$  to plot slightly to the right side of LMWL. The regression equation for  $\delta^2H$  against  $\delta^{18}O$  for porewater (Eq. 18) has a slope of 6.4 with an intercept of 0.74. The slope is slightly lower than that observed for rainwater (slope = 7.5) but the intercept is considerably lower. These deviations are indicative of isotopic enrichment attending evaporation of infiltrating rainwater. Cheng et al. (2014) suggest that evaporation is the main factor responsible for isotopic enrichment within the soil zone. The effects of evaporation has, however, been reported to decrease with depth depending the prevailing conditions

and the location. In the study area, due to the low hydraulic conductivities of the largely fine grain material in the shallow subsurface, vertical fluxes of water are restricted, thus exposing infiltrating water to high levels of evaporation especially in the near surface. Barnes and Allison (1984) report that the depth of penetration of evaporation effects down the soil profile depends on the soil type, the time interval, and the rate of evaporation. Thus, an evaluation of the changes in isotopic signature of porewater at different depths will provide indications of infiltration processes and the progressive losses accompanying evaporation of infiltrating water. The dominant mechanism of infiltration is important in the overall assessment of groundwater recharge processes in the basin. The chloride profile data will assist in evaluating the percentage of the original precipitation that would be lost to the combined impacts of evaporation and transpiration through the profile. The water losses estimated with the stable isotope data represent only evaporative losses as transpiration has not been determined to lead to isotopic fractionations. In this respect, it will be possible to estimate the proportion of the evapotranspiration losses that are attributed to transpiration alone. The joint application of the isotope and chloride tracers will assist in evaluating the possible extinction depth of evaporation in the vadose zone. The extinction depth of evapotranspiration is the maximum depth of penetration of the effects of evapotranspiration. In Fig. 7, the LMWL and PEL intersect at a point where  $\delta^2H$  and  $\delta^{18}O$  are, respectively,  $-37.5$  and  $-5.8$  ‰. This is the average signature of the source rainwater (Peng et al. 2011; Dogramaci et al. 2015) that begun transit through the unsaturated zone in the study area, and is consistent with

**Fig. 7** A biplot showing the relationship between porewater isotope content and that of the local rainwater



the signature of rainwater in low latitudes. These values also fall within the range of variation in the rainwater isotope data used in this study. They were used as the starting isotope signatures and the evaporation rates at the various intervals of profile sampling were made in accordance with the approach of Gibson and Reid (2010) and Dogramaci et al. (2012, 2015).

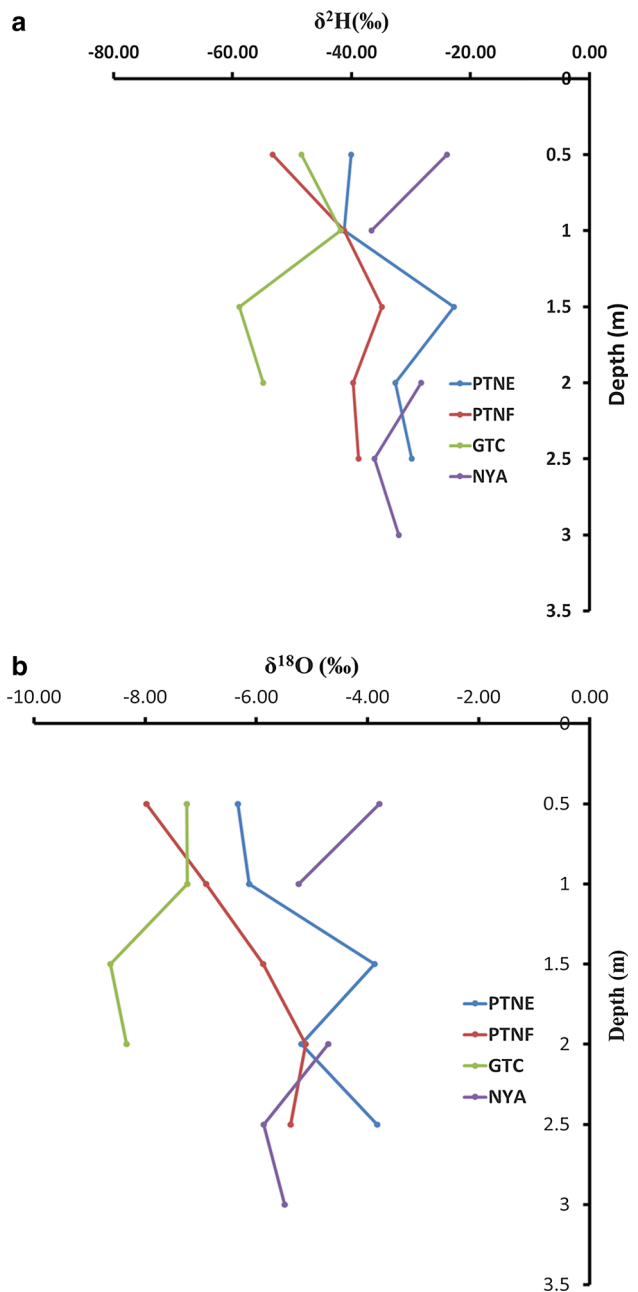
$$\delta^2\text{H} = 6.4\delta^{18}\text{O} - 0.74, \quad R^2 = 0.89 \quad (18)$$

Figure 8a presents the vertical profile of the  $\delta^2\text{H}$  for the four sampled locations. For PTNE and PTNF, there is an obvious enrichment in  $\delta^2\text{H}$  from the very top of the profile to the maximum depth of 2.5 m, without any indication of attenuation. Piston flow (Zimmermann et al. 1966), whereby newer water progressively displaces older water down the soil profile, appears to be the dominant mechanism in these areas. In such a case, the effects of evaporation are expected to be preserved in the stable isotope signature up to the depth whereby it attenuates. In this particular study, there has been no obvious sign of attenuation in the signal of  $\delta^2\text{H}$  up to the maximum depth of the sampling. This suggests that the extinction depth of evaporation in these areas may be deeper than the 2.5 m depth. In the case of GTC, there appears to be an obvious enrichment of  $\delta^2\text{H}$  from 0.5 m to 1.0 m. Thereafter, depletion in  $\delta^2\text{H}$  occurs from 1.0 m to 2.0 m, indicating that preferential flow, rather than piston flow is the main mechanism of vertical transport in the area. Preferential flow occurs when newer precipitation bypasses older ones through ingresses created by roots of plants, insect burrows, vertical fractures and bores, and other features in the unsaturated zone. Such a situation leads to the pattern observed whereby the most depleted signatures are observed in deeper regions whilst older waters remain in

the interstices of soil in shallower regions. The soil in the sampled locations may have been opened up by the roots of the savannah shrubs which have long tap roots and are prevalent in the sampled locations. However, the vadose zone material in much of the study area is clear of thick vegetation and the soils are largely clayey to silty. Although this study did not evaluate grain size distribution of the material sampled, earlier investigations (e.g. Nyarko 2007) in other areas suggest that the material has the propensity to retain infiltrating water over longer periods due to low hydraulic conductivities of the fine material, thus exposing it to higher chances of being affected by evaporative effects especially within shallow zones. It is therefore expected that much of the region will portray the character exhibited by the profiles of PTNE and PTNF. The  $\delta^2\text{H}$  profile for NYA appears to be largely flat, displaying slight depletion down the profile. This may suggest the combined effects of both piston flow and preferential flow whereby older waters in the deeper regions of the profile are mixed with newer infiltrating water to present a largely constant signature. The trend is replicated in the  $\delta^{18}\text{O}$  signatures in Fig. 8b.

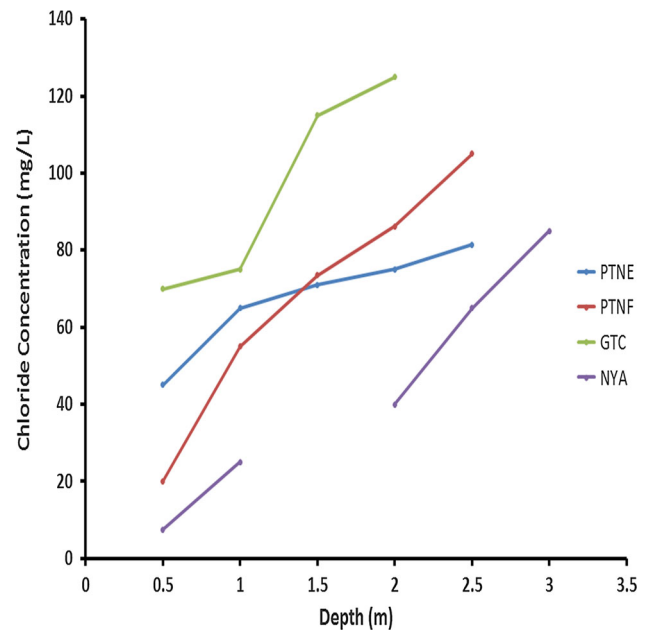
#### Estimation of recharge, evaporation, and transpiration rates

The chloride profile (Fig. 9) appears to suggest progressive increase in the concentration of the chloride ion throughout the depth of sampling for all the areas sampled. There is no indication of attenuation of chloride concentration, suggesting continuous impacts of evapotranspiration even below the maximum depth of the profile. The chloride profiles for PTNE and PTNF are largely consistent with those of the stable isotopes and



**Fig. 8** **a** A diagram showing variations in porewater  $\delta^2\text{H}$  content down the soil profile in the study area. **b** A diagram showing variations in porewater  $\delta^{18}\text{O}$  content down the soil profile in the study area

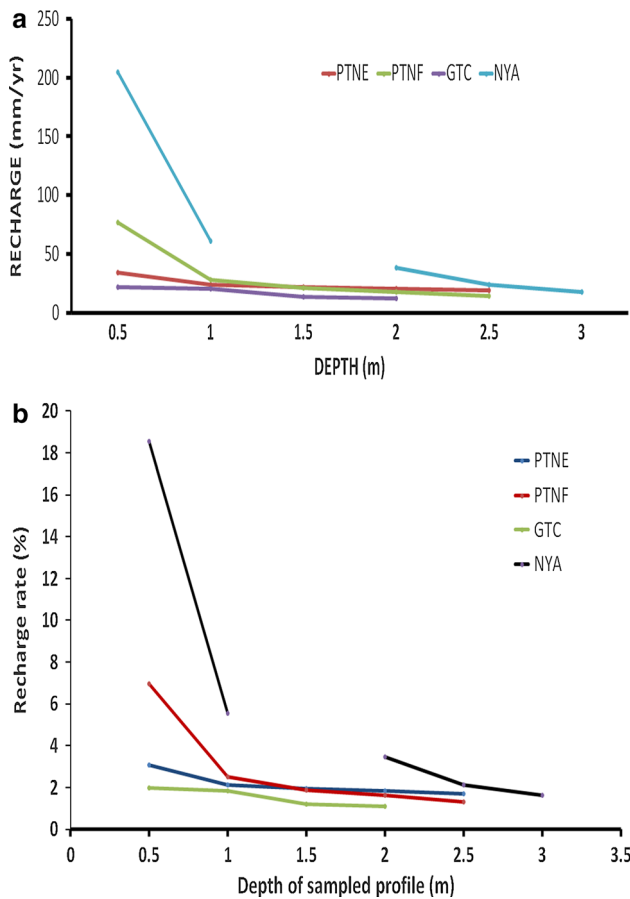
suggest progressive changes in infiltrating water down the profile. In the case of GTC, there's a progressive increase in the chloride content down the profile, although the isotope signatures do not appear to follow in a similar pattern. This may arise from the fact that the chloride ion is much more responsive to the impacts of evapotranspiration than do the stable isotopes. Stable isotope tracers are particularly chemically and biologically stable and do not show fractionation during water uptake by roots of



**Fig. 9** Variations in the porewater chloride content down the profile in the unsaturated zone

plants (Cheng et al. 2014). Therefore, the direct measurement of  $\delta^2\text{H}$  and  $\delta^{18}\text{O}$  in porewater does not provide indication of the effect of transpiration. It is therefore probably on account of this that the isotopic profiles in both GTC and NYA appear to be different from their respective chloride profiles.

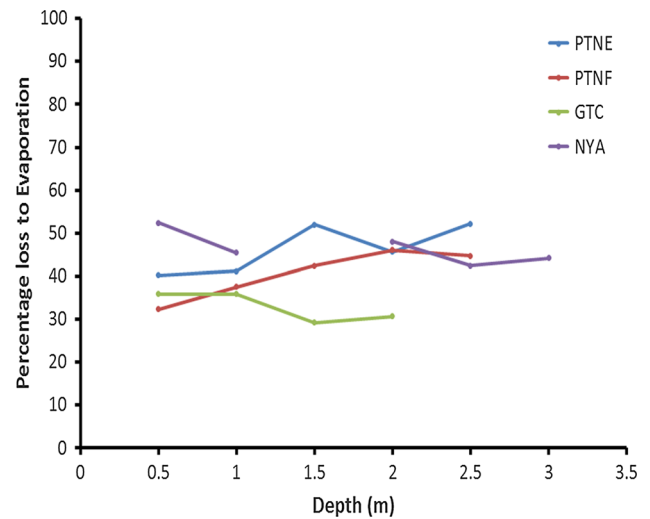
Estimates of the fraction of infiltrating precipitation remaining at every interval of sampling, after the combined effects of transpiration and evaporation, using the chloride mass balance technique are presented in Fig. 10a, b. These estimates have been referred to as diffuse recharge (Ting et al. 1998) as the fraction of the initial precipitation that will eventually reach the saturated zone may be lower. There is a progressive decrease in unsaturated zone recharge down each profile. From the very surface, the four profiles present very different values. However, with increasing depth, the estimates progressively become similar and appear to converge around 2.5 m depth. At this depth, the estimates of recharge suggest approximately an average of 35 mm/year which translates into about 3.2 % of the annual precipitation in the area. The estimated unsaturated zone recharge ranges between 12.2 and 204 mm/year representing 1.1 and 18.5 %, respectively, of the annual precipitation in the area. As the profiles suggest a decreasing trend in the recharge rates, the final fraction of the initial precipitation that will eventually reach the saturated zone as direct recharge could be considerably lower. Although the profiles appear to converge, they all appear to indicate a decreasing trend without any clear indication of



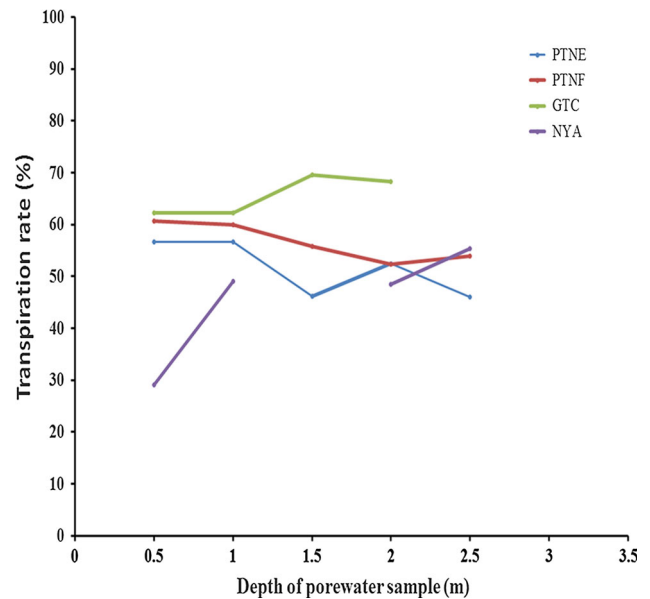
**Fig. 10** Variations in recharge and evaporation rates down the soil profiles in the study area

attenuation. This implies that evapotranspiration losses are still active below the maximum depth of the sampling in this study, and unsaturated zone estimates of recharge may not realistically represent groundwater recharge in the area. It will appear therefore that if the range of recharge estimates at these depths of sampling represent 13.63–0 % of the precipitation in the area, 80–86.3 % of the initial precipitation would have been lost to the combined effects of evaporation and transpiration in the soil zone.

Estimates of evaporation rates (Fig. 11) suggest progressive increase down the profile. Estimated evaporation rates of infiltrating water through the profile, ranges between 29.3 and 52.4 %, with an average of 42 % of the annual average precipitation in the area. At 2.5 m depth, PTNE and PTNF suggest evaporation rates of 52.4 and 45 % respectively. This implies that 45–52.4 % of the precipitation that begins transit through the unsaturated zone would have been lost to evaporation alone at this depth. This is quite significant and has serious implications for groundwater recharge in the area. Water losses arising



**Fig. 11** Estimated evaporation rates and their variations down the soil profile



**Fig. 12** Estimated transpiration rates and their variations down the soil profile

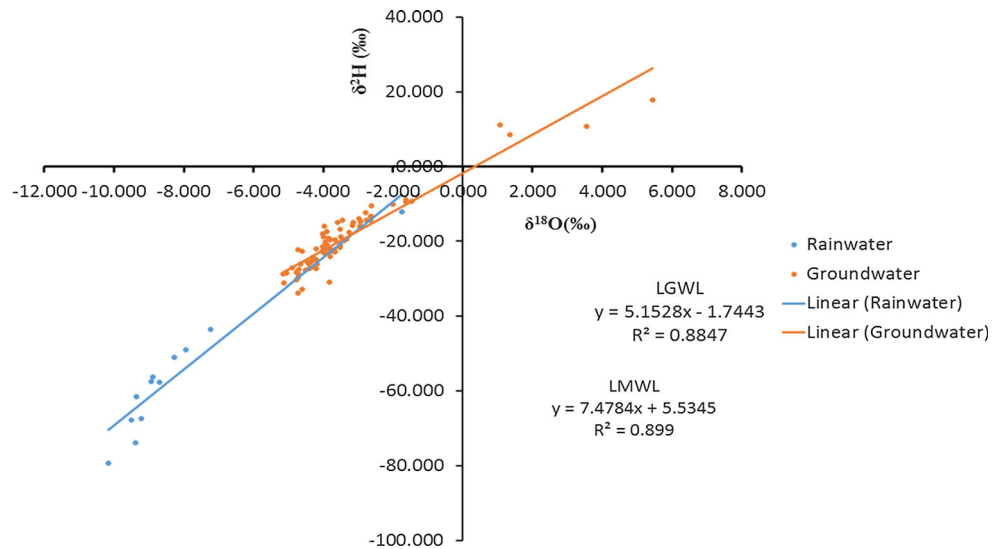
from the impacts of transpiration were estimated using the approach suggested in Dogramaci et al. (2015). If  $t$  represents the fraction of the initial precipitation lost to transpiration, then

$$t = 1 - r_{cmb} - e \tag{19}$$

where  $r_{cmb}$  and  $e$  respectively, refer to the recharge estimates at the profile intervals and evaporation rates, respectively.

Figure 12 shows the profile estimates of transpiration rates corresponding to the estimates of recharge and

**Fig. 13** A biplot of the stable isotope data of groundwater in relation to the LMWL



evaporation for the profile intervals. The estimated transpiration rates range between 1.03 and 57.93 % with an average of 32.0 % of the precipitation in the area. The effect of transpiration appears to be increasing from the surface up to the maximum depth of the profile sampling (Fig. 13) except for PTNE which appears to be flat. There appears to be an increasing trend of the transpiration rates down the profile, especially for PTNF and GTC. It implies that transpiration is an active process below the maximum depth of profile sampling in this study, and may be continuous through the entire thickness of the vadose zone. This may be attributed to the impacts of the drought resistant savannah vegetation whose long taproot systems can reach several meters down the soil profile and into the saturated zone. These trees, although scattered in the terrain, are significant in the analysis of water loss through evapotranspiration in the vadose zone.

**Analysis of groundwater recharge and effective water loss**

Estimates of recharge using the porewater chloride data suggest groundwater recharge of about 150–220 mm/year between a depth of 2.5 and 3.0 m representing 13.63–20 % of the average annual precipitation in the area. The trend suggests that these rates may further reduce and the effective groundwater recharge may be considerably lower. It has been noted that due to high clay content of the unsaturated zone material in some places, vertical hydraulic conductivities, and hence infiltration rates are quite low and the resultant groundwater recharge rates may be lower in the long run.

Groundwater recharge rates estimated from the WTF method range between 64.65 and 151.2 mm/year, translating into 5.9 and 13.7 % of the annual depth of

precipitation in the area, with an average of 9.3 % (102.5 mm/year). This compares favorably with the recharge estimates from previous researchers in other parts of the larger Voltaian Basin, using other methods (e.g. Attandoh et al. 2014; Yidana and Koffie 2014). The obvious difference between the estimates from the porewater data and WTF methodology may have arisen from the fact that the porewater data represent water still transitioning through the unsaturated zone material. The data suggests that there may be no well-defined extinction depth for evapotranspiration in the area, and that the process of evapotranspiration may be active throughout the thickness of the vadose zone.

Using the groundwater isotope data, the evaporation rates of precipitation which recharged the aquifers in the area were estimated. The LGWL is defined by Eq. 20. As indicated in Fig. 13, this line intersects the LMWL at a point where the  $\delta^{18}\text{O}$  and  $\delta^2\text{H}$  values are, respectively,  $-3.1$  and  $-20.0$  ‰. This represents the isotopic signature of the source rain which recharged the aquifers after going through transformative processes through the vadose zone. These isotopic signatures were used as the initial signatures for estimating the possible evaporative loss that occurred prior to and in the process of infiltration and percolation into the saturated zone. The estimated evaporation rates range between 31.8 and 54.0 %, with an average of 40.0 %. These estimates are statistically in the range of evaporation rates computed from the porewater data, and suggests that much of the evaporative losses occurred within the depth range of 0–3 m within the vadose zone and that the reduced recharge rate may be attributed largely to water losses due to transpiration. The long tap root systems of the typical savannah plants, coupled with high ambient temperatures and low relative humidities enhance evapotranspiration. Rising temperatures consistent with the impacts of changing

climate (Kunstmann and Jung 2005) will further increase plant water requirements and enhance evapotranspiration rates in the area. Estimated transpiration losses prior to and in the process of infiltration and percolation of precipitation into the saturated zone range between 37 and 59.2 %, with an average of 51.0 %, which is higher than the range estimated from porewater in the vadose zone. This means that unlike evaporative losses which appear to be much more intense near the surface, transpiration is persistent through the entire vadose zone and accounts for a huge percentage of the water loss prior to recharge.

$$\delta^2\text{H} = 5.2^{18}\text{O} - 1.74 \quad (20)$$

## Conclusions

Vertical infiltration of precipitation in the vadose zone of the Nabogo catchment of the White Volta Basin is dominated by piston flows. Evaporation rates computed from stable isotope data suggest water losses to the tune of 29.3–52.4 % of the annual precipitation within the shallow subsurface. The impact of transpiration accounts for 1.03–57.93 %, with an average of 32.0 % of the annual precipitation within the shallow (0–3 m) reaches of the vadose zone in the area. Vadose zone recharge estimated from porewater data suggests 150–220 mm/year between a depth of 2.5 m and 3.0 m representing 13.63–20 % of the average annual precipitation in the area. The impacts of evapotranspiration rates appear to continue below the maximum depth of sampling in this study (3 m) and suggest that extinction depth of evapotranspiration in the area may not be defined in the vadose zone. This appears to be corroborated by groundwater recharge rates estimated with the water table fluctuations method in the domain. Direct groundwater recharge estimates from the Water Table Fluctuation Method, presents recharge in the range of 5.9–13.7 % (64.65–151.2 mm/year) of the annual precipitation in the region, suggesting that there are still water losses below 3.0 m in the vadose zone. This suggests that extinction depth of evapotranspiration in the study area proceed below the maximum depth of sampling of the vadose zone material in the study area as it appears that much of the water in transition is lost between the depth of 3.0 m and the saturated zone.

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