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Estimation of soil water evaporative loss after tillage operation using the stable isotope technique

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A b s t r a c t. Application of stable isotopes in soil studies has improved quantitative evaluation of evaporation and other hydrological processes in soil. This study was carried out to determine the effect of tillage on evaporative loss of water from the soil. Zero tillage and conventional tillage were compared. Suction tubes were installed for soil water collection at the depths 0.15, 0.50, and 1.0 m by pumping soil water with a peristaltic pump. Soil water evaporation was estimated using stable isotopes of water. The mean isotopic composition of the soil water at 0.15 m soil depth were -1.15‰ ($\delta^{18}O$) and -0.75‰ (δD) and were highly enriched compared with the isotopic compositions of the site precipitation. Soil water stable isotopes ($\delta^{18}O$ and δD) were more enriched near the surface under zero tillage while they were less negative down the profile under zero tillage. This suggests an occurrence of more evaporation and infiltration under conventional then zero tillage, respectively, because evaporative fractionation contributes to escape of lighter isotopes from liquid into the vapour phase leading to enrichment in heavy isotopes in the liquid phase. The annual evaporation estimated using the vapour diffusion equation ranges from 46-70 and 54-84 mm year^{-1} under zero and conventional tillage, respectively, indicating more evaporation under conventional tillage compared with zero tillage. Therefore, to reduce soil water loss, adoption of conservation tillage practices such as zero tillage is encouraged.

K e y w o r d s: evaporative loss, tillage, isotopic fractionation, isotope technique

INTRODUCTION

Tillage is mechanical manipulation of the soil for the purpose of crop production and it affects significantly soil characteristics such as soil water conservation, soil temperature, infiltration and evapotranspiration processes. As tillage is known to cause soil surface disruption, fractionation of stable isotopes following soil tillage has been documented (Angela *et al*., 2009). Although tilled plots may have very high initial infiltration rate values, these rates rapidly decline with time (Guzha, 2004) probably due to intense alteration of the top centimetres of the soil (Josa *et al*., 2010) and rapid structural deterioration caused by slaking and dispersion. However, due to the effect of increased soil organic matter (Lipiec *et al*., 2006) and soil surface protection, high infiltration under zero tilled (ZT) plots have been documented (Shukla *et al*., 2003). According to Gupta *et al*. (2004) less intense tillage keeps the crop residue at the soil surface, thereby increasing the activity of surface-feeding earthworms, leaving the root channels undisturbed, leading to the presence of numerous surface-connected macro-pores and inter-pedal, hence higher infiltration. It has been reported that lower soil temperature due to higher water content in the topsoil and more plant residues on the soil surface under ZT usually result in reduced evaporation, whereas the opening of topsoil enhances evaporation from the tilled soil layers (Lal and Shukla, 2004).

Due to isotopic fractionation that occurs when soil water is subjected to evaporation, stable isotopes have gained tremendous application in monitoring fluxes of water in soils. The study of water movement in the vadoze zone of the soil is necessary because plant nutrients, applied either in organic or inorganic form, need water to form a solution before it becomes useful to crop plants on the one hand or is

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transported down the profile on the other hand. Information about soil hydrological processes such as infiltration, evaporation, transpiration and percolation can be easily obtained using stable isotope compositions of soil water (Gazis and Feng, 2004). Additionally, Mook and de Vries (2001) stated that quantitative evaluation of soil physical processes such as evaporation as well as distinctive information on soil water movement and its dissolved constituents could be obtained by tracing soil water by means of environmental isotopes.

In hydrological studies, the water molecules that are generally of interest are $H_2{}^{18}O$ and $^2H^1H^{16}O$ and their diffusivity ratios are expressed as $D(H_2{}^{18}O)/(D_2{}^{16}O) = 0.9691$ and $D(HD^{16}O)/D(H_2{}^{16}O) = 0.9839$. These molecules contain ^{18}O and deuterium (^2H, or D) isotope, respectively (Mark, 2011). In the determination of ^{18}O and D, the δ notation is expressed as ^{18}O or $D = (R_{sample}/R_{standard} - 1)\ 10^3$, where $R = (^{18}O/^{16}O)$ or (D/H) and $R_{standard}$ is the corresponding ratio in Vienna Standard Mean Ocean Water (V-SMOW). Although either $\delta^{18}O$ or δD could be measured to give pathway of evaporation or infiltration, measurement of both $\delta^{18}O$ and δD has been reported to give a linear relationship expressed as $\delta D = 8\ \delta^{18}O + 10$ (Mark, 2011), which is termed the global meteoric water line (GMWL). This knowledge enables us to obtain the local meteoric water line (LMWL) from the determination of $\delta^{18}O$ and δD of rainfall of a particular location. The LMWL then becomes potentially useful to fingerprint evaporation that has taken place in the location when compared with the GMWL (Welker, 2000).

Kinetic fractionation of hydrogen and oxygen isotopes can be attributed to faster diffusion of molecules containing lighter atoms compared to heavier ones. Consequently, evaporative fractionation (or condensation) causes lighter isotopes to escape from liquid into the vapour phase and the net effect of evaporation is therefore an enrichment in heavy isotopes ^{18}O and D near the soil surface (Gazis and Feng, 2004). Quantitatively, the isotopic changes occurring between the liquid and vapour phases during the various stages of evaporation have been reported by Cappa *et al*. (2003). For example, they reported that at stage one the $\delta^{18}O$ was -7.24 and -22.31‰ for liquid and vapour phases, respectively, while the δD was -69.78 and -160.34‰ for liquid and vapour phases, respectively. This indicates that liquid phases are usually isotopically more enriched (Mark, 2011) than the vapour phases. However, one of the ways by which soil loses water is plant root uptake, water uptake and transport through the root system have been described as passive processes (Mendel *et al*., 2003). Therefore, fractionation of stable isotopes of water does not occur during root uptake as evaporation of soil waters takes place before they have been taken up by plants (Liu *et al*., 2011).

Field observations of $\delta^{18}O$ and δD enrichments of soil water near the surface due to evaporation has been reported by many authors (Gazis and Feng, 2004; Robertson and Gazis, 2006). Therefore, enrichment or depletion of stable isotopes of water near the soil surface, compared to the mean isotopic value of local rain water, can depict soil water evaporation or infiltration, respectively. This is because the composition of stable isotope of soil water depends on precipitation inputs, antecedent conditions, and evaporative losses (Ferretti *et al*., 2003). This study was, therefore, carried out to determine the effect of tillage on fractionation of stable isotopes of soil water so as to estimate the influence of tillage on soil water evaporation or infiltration.

MATERIAL AND METHODS

The study was carried out at the University of Agriculture, Abeokuta, S-W Nigeria in 2009. The study site lies between Latitude 7°14'N and Longitude 3°26'E and is located within a forest-savanna transition zone (Salako e*t al*., 2007) with two distinct seasons – the wet season, which extends from March to October, and the dry season which is usually from November to February. In the study area, the average annual rainfall, based on a ten-year period, is 1058.48 mm. The rainfall is bimodal in distribution – usually from March to July and September to October, with a characteristic August break (Fig. 1). Similar to a decade average (31.6°C), the mean monthly temperature ranged from 23.5 to 34.4°C in 2009 (Fig. 2). The mean relative humidity was 68.3%. The annual evaporation obtained by the class A Pan evaporation method at the study site in 2009 was 915 mm. All agrometeorology data were obtained from the Department of Water Resources Management and Agrometeorology of the University of Agriculture, Abeokuta, with the weather station located about 1 km to the study site. Soil thermometers installed on the field between June and September 2009 revealed that the soil temperature ranged from 28 to 31°C and from 26 to 29°C at 5 and 25 cm soil depths, respectively. The soil of the study site is gravely loamy sand and is

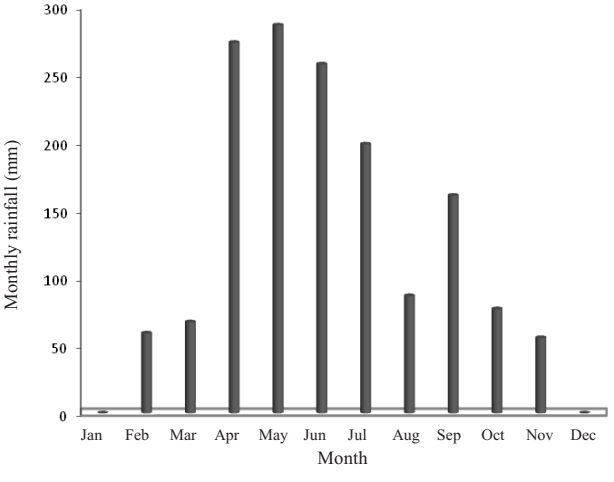

Fig. 1. Monthly rainfall distribution of the study site in 2009 (Data source: Water Resources and Agrometeorology Department, University of Agriculture, Abeokuta, Nigeria).

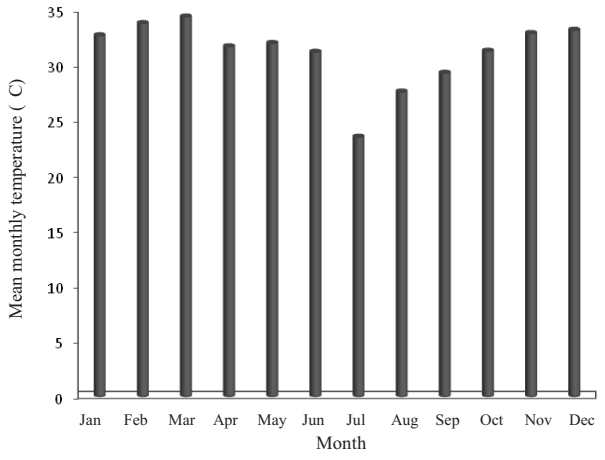

Fig. 2. Mean monthly temperature distribution of the study site in 2009. Explanations as in Fig. 1.

underlain by an undifferentiated basement complex of alluviocolluvial parent material. Taxonomically, the soil was classified as Arenic Plinthic Kandiudalf in the USDA classification system (Soil Survey Staff, 2006) and Arenic Lixisol in the FAO/UNESCO classification system (WRB, 2006).

The treatments imposed were zero tillage (ZT) and conventional tillage (CT) systems. Under ZT, no ploughing was carried out but weeds were destroyed using contact herbicides (paraquat), while CT involved ploughing followed by harrowing. Maize (*Zea mays*) was planted in May 25, 2009 at the study site and was harvested 12 weeks after planting but the maize stovers were left standing on the field as means of reducing soil surface bareness.

Cumulative rainfall in the field was collected between June and September, 2009 by installing three rainfall collectors – two of them on the outer edges and one in the middle of the field (dimension = 50 by 80 m). The rainfall collectors were made of sampling bottles connected to funnels and were protected against flies and large particles with a polyethylene net. Prior to the installation, the rainfall collectors were cleaned with deionized water. The bottles were wrapped with aluminium foil to reduce the direct effect of radiation on the rainwater. The rainfall collectors were placed on a platform, raised 1 m above the soil surface to prevent soil splashes from entering the bottle during rainfall events.

Soil suction tubes were installed for soil water collection under zero and conventional tillage at the depths of 0.15, 0.50, and 1.0 m by pumping the soil water with a peristaltic pump. Soil water was extracted in June, August, and September of 2009. Water extraction by the suction tubes was not possible in July apparently due to a low soil water potential. The pressure pump extracted water from the soil at a tension of 0.78 bar. Water extraction at each sampling point proceeded for an average of 1.5 h, by which more than required soil water (10 cm^3) needed for analysis would have been extracted.

The $\delta^{18}O$ and δD of the precipitation and soil water were analyzed using a PICARO (L1102-i Isotopic Liquid Water analyzer) Cavity Ring Down Spectrometer (Picarro Inc. Sunnyvale, CA, www.picarro.com) at the Department of Geosciences, University of Trieste, Italy. The precision of the measurement was 0.1 and 0.5‰ for $\delta^{18}O$ and δD, respectively. All the results were expressed in the standard notation as relative differences (δ value) from Vienna Standard Mean Ocean Water (V-SMOW) in parts per mil (‰). All the samples were run three times and the reported values were means of the readings. All the measurements were carried out against the laboratory standards that were periodically calibrated against the IAEA international isotope water standard (V-SMOW).

At the quasi steady-state, during the second stage of evaporation from the soil surface, the evaporation rate (E) from the evaporation front was estimated using Fick law (Liu *et al*., 1995):

$$E = D_{v*}\frac{\partial(hN_s)}{\partial Z} = D_v\tau(n-\theta)N_s\left(\frac{h_{ef}-h_a}{\rho Z_{ef}}\right),$$

where: E (mm $year^{-1}$), D_{v*} and D_v (cm^2 s^{-1}) are the diffusivities of water vapour in the porous medium and air, respectively.

$D_v = 0.229\ 10^{-4}\left(\frac{1+T}{273.16}\right)^{1.75}$ where T is in °C (Ebrahimi *et al*., 2004), τ (dimensionless) is tortuosity, n (%) is soil porosity, θ (v/v) is the volumetric soil water content, h (%) is relative humidity, N_s (g cm^{-3}) is saturated water vapour density as a function of temperature obtained from List Smithsonian Meteorological Tables (Kostinski and Cantrell, 2008), ρ (g cm^{-3}) is the density of water, Z (cm) is depth, and subscripts ef and a represent evaporation front and free air, respectively.

The volumetric soil water content was measured using a time domain reflectometer or TDR (Ferré and Topp, 2002) with 7.5 cm (short) prongs (Field scout TDR 100/200, Spectrum Technologies, Inc. http://www.spectmeters.com).

RESULTS AND DISCUSSION

The mean isotopic compositions of the site precipitation were -2.96‰ ($\delta^{18}O$) and -13.76‰ (δD) and were more depleted relative to -1.15‰ ($\delta^{18}O$) and -0.75‰ (δD) for the soil water at 0.15 m soil depth (Table 1). It has been reported that soil water is normally more enriched because, as rain water enters the soil, it mixes with antecedent soil water that has been modified by evaporation (Ferretti *et al*., 2003). Isotopic fractionation is markedly influenced by temperature (Mark, 2011), thus the lower temperature in August (27.6°C) compared with the other months could be implicated for more isotopic depleted precipitation recorded in this month (Table 1).

T a b l e 1. Isotopic composition of rainfall and soil surface (0.15 m) water

Sampling date (2009)	Rainfall water		Average soil surface water stable isotope*	
	δ^{18}O‰ V-SMOW	δD‰ V-SMOW	δ^{18}O‰ V-SMOW	δD‰ V-SMOW
June	-2.69	-13.21	-2.67	-9.00
August	-3.42	-15.83	-0.25	0.85
September	-2.77	-12.23	-0.54	5.90
Mean	-2.96	-13.76	-1.15	-0.75

*The stable isotope reported in each month is an average of values obtained at the soil surface (0.15 m) under both conventional and zero tillage. V-SMOW – Vienna Standard Mean Ocean Water.

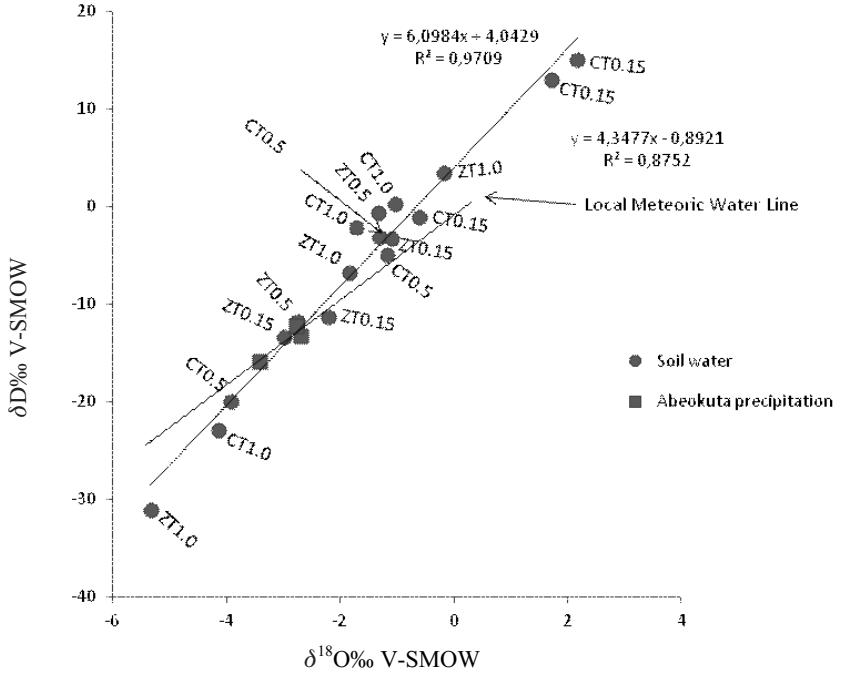

Fig. 3. Relationship between δ^{18}O and δD of precipitation and soil water. CT0.15, CT0.5, CT1.0 – soil water at 0.15, 0.5, 1.0 m depth, respectively under CT; ZT0.15, ZT0.5, ZT1.0 – soil water at 0.15, 0.5, 1.0 m depth, respectively under ZT. Other explanation as in Table 1.

Characteristically, there was a strong linear relationship (R^2 = 0.97) between δ^{18}O and δD of rain and soil water (Fig. 3) but the slope of the relationship was lower than that of Craig value of 8 for the global meteoric water line (Jia *et al*., 2008). When the slope is low, the implication is that soil water is subjected to evaporation through the dry surface layer (Jia *et al*., 2008). The two points located farther from the points of local precipitation and at the upper end of the regression line are samples from the surface horizon (0.15 m) under CT in August and September 2009 (Fig. 3). The location of two soil surface water sample points at the upper end of the regression line means that a higher amount of evaporation occurred under CT from the soil surface because the net effect of evaporation is enrichment in heavy isotopes near the soil surface (Gazis and Feng, 2004).

Soil water stable isotopes (δ^{18}O and δD) were more enriched near the surface under CT in all the three months of soil water sampling, while enrichment near the surface was observed only in June under ZT (Fig. 4). More isotopic enrichment near the surface under CT in all the sampling periods is a clear indication of more water evaporation, while upward isotopic depletion taking place under ZT (with the exception of June) indicates a possibility of soil water infiltration. However, it should be noted that transpiration may reduce the amount of water available for infiltration because transpiration does not induce isotope fractionation (Mook and de Vries, 2001); interestingly, the net water after the evapotranspiration process will result in infiltration. The soil water was isotopically heavier down the profile under ZT in August and September. The downward

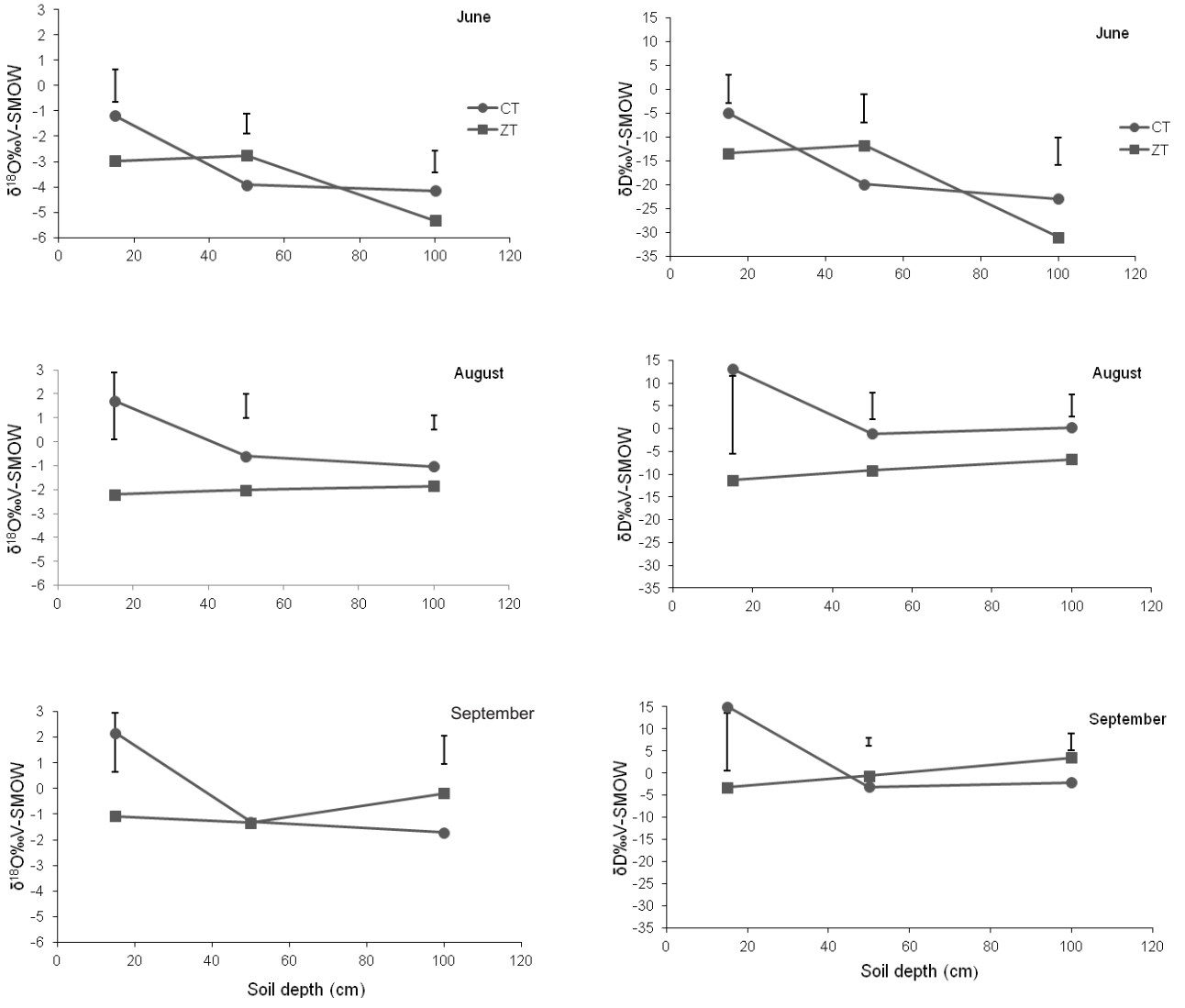

Fig. 4. Comparative effect of conventional tillage (CT) and zero tillage (ZT) on fractionation of: a – $\delta^{18}O$, b – δD of soil water. Vertical lines on data points are standard deviation bars. Other explanation as in Table 1.

isotope enrichment under ZT suggests that while some soil water remains stationary, the mobile water successfully displaces pre-existing mobile soil water pushing it downward (Gazis and Feng, 2004). Field observations of δD and $\delta^{18}O$ enrichments of soil water near the surface have been reported by many studies and have been linked with evaporation (Gazis and Feng, 2004; Robertson and Gazis, 2006).

The use of graphs of deuterium excess (Fig. 5) further clarified the relationship between stable isotopes of soil water and soil tillage. Due to the relationship existing between $\delta^{18}O$ and δD, deuterium excess (dex) which is defined as $dex = \delta D - 8\delta^{18}O$ (Lambert and Aharon, 2010), is used to indicate the vapour movement. In this study, d_{ex} demonstrates that more isotopic fractionations occurred due to CT compared with ZT across the depths in all the months under observation. The dex ranged from 10.45-11.61‰ (June); 6.40-8.02‰ (August) and 4.93-10.01‰ (September) under ZT as against the ranges of 4.44-11.39‰ (June); -0.66-8.49‰ (August) and -2.32-11.66‰ (September) under CT (Fig. 5). The d_{ex} values were consistently lower under CT than ZT. Low d_{ex} has been attributed to a high rate of the evaporation process in the atmosphere (Vallet-Coulomb *et al*., 2008), while the higher dex values under ZT signify a reduced degree of evaporation (Lambert and Aharon, 2010).

Except for the rainy season, the soil water cannot meet the evaporative demand of the soil at the study site (Fig. 6), and thus the evaporation processes is likely to proceed in the quasi-steady state stage characterized by vapour transport through a dry soil layer into the atmosphere. Using the vapour diffusion equation, the evaporation in the quasi-steady

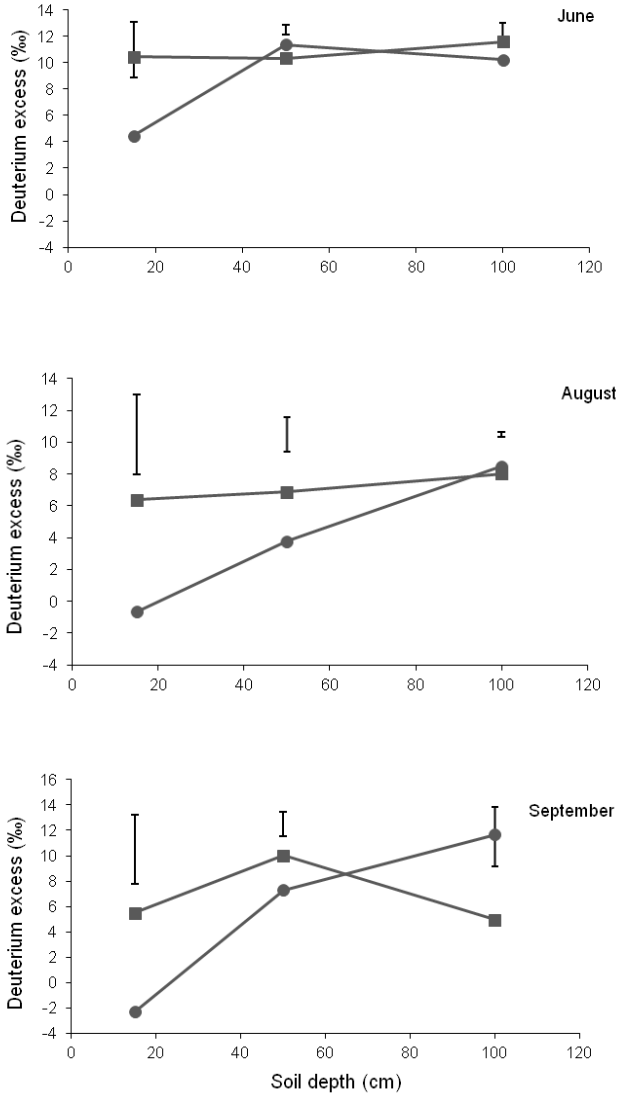

Fig. 5. Comparative effect of CT and ZT on soil water deuterium excess. Vertical lines on data points are standard deviation bars. Other explanation as in Fig. 4.

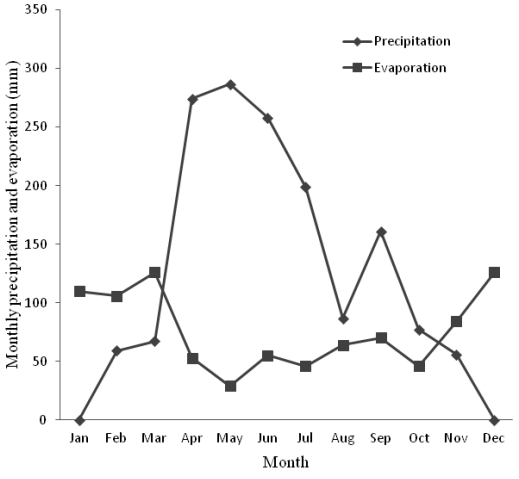

Fig. 6. Monthly precipitation and evaporation of the study site in 2009.

state stage was estimated. The parameters used in the calculation are 0.67 for tortuosity (τ), 0.52 for porosity (n) obtained as the average n of the field; 1 for relative humidity at the evaporation front (h_{ef}) (Liu *et al*., 1995), 1 g cm^{-3} for water density, and 0.15 m for the depth of evaporation front (Z_{ef}) determined from the depth of the δD maximum bulge between CT and ZT in the stable isotope profile. The temperature (T) and relative humidity of air (ha) were obtained from the meteorological station of the University of Agriculture Abeokuta. θ is a weighted mean for the volumetric soil water content, saturated water vapour density (N_s) was obtained from List Smithsonian Meteorological Tables (Kostinski and Cantrell, 2008), and the vapour diffusivity (Dv) was calculated from the empirical function obtained from Ebrahimi *et al*. (2004). The annual evaporation estimated using the equation ranged from 46-70 mm year^{-1} under ZT and from 54-85 mm year^{-1} under CT (Table 2). Furthermore, the estimated evaporation was the lowest in August compared with the other two months under observation (Table 2). Similar to the observation in the stable isotope profiles, estimation of evaporation using the vapour

T a b l e 2. Parameter values for the computation of equation used for estimation of evaporation rate under ZT and CT

Month	Tillage	T (°C)	N_s (g cm^{-3}) (10^{-6})	D_v [a] (cm^2 s^{-1})	θ (v v^{-1})	h_a	E (mm year^{-1})
June	CT	31	32.07	0.2764	0.22	0.84	60
	ZT	31	32.07	0.2764	0.24	0.84	56
August	CT	28	27.24	0.2716	0.09	0.88	54
	ZT	28	27.24	0.2716	0.15	0.88	46
September	CT	29	28.78	0.2732	0.07	0.83	85
	ZT	29	28.78	0.2732	0.15	0.83	70

D_v [a] – diffusivities of water vapour in the free air, T – temperature (°C), h_a – relative humidity of free air.

diffusion equation revealed that more evaporation took place under CT compared with ZT in the three months under consideration. Although Dunnett t-test (SAS, 2001) used to compare the means did not reveal a significant difference ($p < 0.05$) in the estimated evaporation between ZT and CT in June, CT witnessed a significantly higher evaporation than ZT in August and September. Dunnett t-test at a 5% level of probability revealed that the minimum significant differences were 9.07 in August and 13.60 in September. It was noted that the major controlling factor for the estimated evaporation rate in this study was the volumetric water content of the soil in the sampling periods. The lowest temperature (27.6°C) in August (Fig. 2) compared with June and September may be responsible for the lowest estimated evaporation obtainable in this month. The average evaporation rate at all the periods under observation and under the two tillage systems is approximately 62 mm year^{-1}. This value is much lower than the site evaporation rate of 915 mm year^{-1}. Although uncertainty in the position of evaporation used in the calculation may account for this lower value, normally the high evaporative demand is satisfied in the first stage of evaporation and thus a higher evaporation rate is expected to have taken place in the first stage. Consequently, the first stage of evaporation probably accounts for the differences between the estimated second stage of evaporation and the expected annual evaporation (Liu *et al*., 1995).

CONCLUSIONS

1. The study revealed that $\delta^{18}O$ and δD fingerprint the pathway of water in the soil.

2. Soil water enrichment in $\delta^{18}O$ and δD, indicating that evaporation, was more pronounced under conventional tillage due to soil pulverization that usually accompanies this type of tillage. On the other hand, zero-tilled soil had more depleted $\delta^{18}O$ and δD, indicating less evaporation and more water available for infiltration, possibly due to covering of the soil surface by residue and a more organic surface.

3. To minimize soil water loss and increase availability of water to crops, adoption of conservation tillage practices such as zero tillage is necessary.

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