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Discussions

Identifying residence times and streamflow generation processes using δ^{18}O and δ^2H in meso-scale catchments in the Abay/Upper Blue Nile, Ethiopia

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Identifying residence times and streamflow generation processes using isotopes

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Abstract

Measurements of the stable isotopes oxygen-18 (^{18}O) and deuterium (^{2}H) were carried out in two meso-scale catchments, Chemoga (358 km^2) and Jedeb (296 km^2) south of Lake Tana, Abay/Upper Blue Nile basin, Ethiopia. The region is of paramount importance for the water resources in the Nile basin. Stable isotope composition in precipitation, spring water and streamflow were analyzed (i) to characterize the spatial and temporal variations of water fluxes; (ii) to estimate the mean residence time of water using a sine wave regression approach; and (iii) to identify runoff components using classical two component hydrograph separations at a seasonal time scale.

The results show that the isotopic composition of precipitation exhibit marked seasonal variations, which suggests different sources of moisture generation for the rainfall in the study area. The Atlantic–Indian ocean, Congo basin, and the Sud swamps are the likely the potential moisture source areas during the main rainy (summer) season. While, the Indian–Arabian, and Mediterranean Sea moisture source areas during little rain (spring), and dry (winter) seasons. The spatial variation of the isotopic composition is affected by the amount effect and to less extent by altitude and temperature effects. A mean altitude effect of $-0.12\,‰\,(100\,\mathrm{m})^{-1}$ for ^{18}O and $-0.58\,‰\,(100\,\mathrm{m})^{-1}$ for ^{2}H were discernable in precipitation isotope composition. The seasonal variations of the isotopic signature of the spring water exhibit a damped response as compared to the river waters, which shows that the spring water has longer residence times than the river water.

Results from the hydrograph separation at a seasonal time scale indicate the dominance of event water with an average of 71 % and 64 % of the total runoff during the wet season in the Chemoga and Jedeb catchment, respectively. The stable isotope compositions of streamflow samples were damped compared to the input function of precipitation for both catchments and this damping was used to estimate mean residence times of stream water of 4.1 and 6.0 months at the Chemoga and Jedeb catchment outlet, respectively. Short mean residence times and high proportions of event

water components suggest catchment management measure aiming at reduction of overland flow/soil erosion and increasing of soil water retention and recharge to enable sustainable development in these agricultural dominated catchments.

1 Introduction

Environmental isotopes as tracers are commonly applied for examination of runoff generation mechanisms at different spatial and temporal scales (e.g. Uhlenbrook et al., 2002; Laudon et al., 2007; Didszun and Uhlenbrook, 2008). Isotope tracer studies are used for hydrograph separations (Sklash and Farvolden, 1979; Buttle, 1994), provide additional information for identifying source areas, flow pathways under different flow conditions and estimating mean residence time of a catchment (Soulsby et al., 2000; Uhlenbrook et al., 2002; McGuire et al., 2005; McGuire and McDonnell, 2006; Soulsby and Tetzlaff, 2008). The application of isotopes in catchment hydrology studies has been carried out in small experimental catchments to meso-scale catchments (e.g. McDonnell et al., 1991; Uhlenbrook et al., 2002; Tetzlaff et al., 2007a) and large scale catchments (Taylor et al., 1989; Liu et al., 2008).

Only few studies have been undertaken to characterize water cycle components using stable isotopes in Ethiopia (e.g. Rozanski et al., 1996; Kebede, 2004; Levin et al., 2009; Kebede and Travi, 2012). The results from these studies indicate that stable isotope composition of precipitation is only little effected by the typical dominant controls: amount, altitude and continental effects. Past studies have shown that Ethiopian meteoric water is uniquely enriched in its isotopic composition as compared to the East African meteoric water (Joseph et al., 1992; Rozanski et al., 1996; Darling and Gizaw, 2002; Levin et al., 2009). Most of these studies hypothesized that the enriched isotopic composition in Addis Ababa is due to advection of recycled moisture from the Congo basin and the Sudd wetland, the increased sea surface temperature at the moisture source, changes in evaporation condition at the source, evaporative enrichment of rain drops, and the Indian ocean moisture results in initial stage of condensation vapor,

which did not undergo major rainout fractionation effect are mentioned as the main contributing factors for the isotopic enrichment.

Nevertheless, the usefulness of stable isotope data for estimation of mean residence times, hydrograph separations and investigation of the dominant runoff components using stable isotopes for catchment hydrological studies are largely unexplored and very little is known about the use of stable isotopes for hydrological studies in the region. Therefore, the utility of stable isotopes will provide enormous use with respect to hydrological process understanding and sustainable planning of water resource management strategies and polices in such data scares areas (Hrachowitz et al., 2011a).

The use of isotope tracer techniques to understand mean residence times (MRTs) and residence time distributions (RTDs) have received much attention in recent time (Uhlenbrook et al., 2002; Rodgers et al., 2005; McGuire and McDonnell, 2006). They are used to gain a better understanding of flow path heterogeneities (Dunn et al., 2007), to get insights into the internal processes of hydrological systems, and are used as a tool for model evaluation (Uhlenbrook and Leibundgut, 2002; Hrachowitz et al., 2011b). Furthermore, they can be used as fundamental catchment descriptors, providing information about the storage, flow pathways and sources of water McGuire and McDonnell (2006), and are used for conceptualizing the differences in hydrological processes by comparing different catchments (McGuire et al., 2005; Soulsby et al., 2006; Tetzlaff et al., 2009).

In the present study, the investigation of meteoric water in the source of the Abay/Upper Blue Nile basin is undertaken as the basis for characterization and better understanding of the dominant runoff components. This is used as a baseline study for future hydrological studies using environmental isotopes in the region. The main objectives of this study are: (a) to characterize the spatio-temporal variations of the isotopic composition in precipitation, spring and stream water; (b) to estimate the mean residence time of stream water; and (c) to separate the hydrograph at a seasonal time scale in the two meso-scale headwater adjacent catchments, Chemoga and Jedeb in the Abay/Upper Blue Nile basin.

2   Study area and data sources

2.1   Study area

The Chemoga and Jedeb rivers are tributaries of the Abay/Upper Blue Nile basin, located south of Lake Tana, and approximately extends between $10°\,10'$ to $10°\,40'$ N latitude and $37°\,30'$ to $37°\,54'$ E longitude. Both rivers originate from Choke Mountain at an elevation of 4000 m a.s.l. (see Fig. 1). The climate in these catchments have a distinct seasonality with three seasons: (i) Summer as the main rainy season from June to September, (ii) winter as the dry season from October to February, and (iii) spring as the short rainy season from March to May (NMSA, 1996).

The long term average annual temperature over the period of 1973–2008 at De-bre Markos weather station is about 16.3 °C. The mean precipitation ranges between 1342–1434 mm a^{-1} (1973–2010) in the lower and upper part of the catchments.

In these two catchments subsistence farming is commonly practiced in which farmers rely on rainfed agriculture for their livelihoods. Barley, Oats and Potato are the main crops grown in the upland area, whereas wheat, tef (*tefferagrostis tef)* and maize are grown in the mid and lower parts of the catchments. According to the study by Bewket and Sterk (2005) and Teferi et al. (2010, 2013) the land use in Chemoga catchment has been subjected to changes since the 1950s. The major change was the increase of cultivated area at the expense of open grazing area and slight increase in plantation forest cover due to eucalyptus plantations. A recent study by Teferi et al. (2013) showed that 46 % of the Jedeb catchment experienced transitions from one land cover to another over the last 52 yr. Nowadays, about 70 % of the land is used for agriculture and 3 % is forest plantations.

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2.2 Data sources

Hydro-meteorological data

Streamflow data sets are based on manual water level measurements (daily at 6 a.m. and 6 p.m.) at Chemoga and Jedeb gauging stations from 1 July 2009 to 31 August 2001. Based on the stage discharge relationships rating curves were developed using regression model.

A network of ten manual rain gauges have been established, since July 2009, for representing the spatial distribution of precipitation and for the purpose of better understanding the rainfall–runoff relationship in the area. Consequently, daily precipitation data were collected from these stations over the same period as the stream flows and daily temperature data at Debere Markos station over the same period were obtained from the Ethiopian National Meteorological Agency. The temperature data at Debre Markos station was used to estimate the temperature at Enerata, Rob Gebeya, Fana Choke, and Yewla stations based on a decrease of $0.6\,^{\circ}\text{C}$ in temperature per 100 m increase in altitude. The catchment average precipitation amount, catchment mean annual temperature, potential evaporation, and isotopic composition of precipitation were computed using the Thiessen polygon method. Due to the limited climatic data availability, the potential evaporation was computed using Hagreaves method (Hargreaves and Samani, 1982). The intra-annual variability of hydro-climatic data within the catchments is shown in Fig. 2. Furthermore, detailed descriptions of the hydro-metrological data and isotope sampling sites are presented in Tables 1 and 2, respectively.

3 Methodology

3.1 Field measurements and sampling

To characterize the spatial and temporal variability of stable isotope composition in precipitation, spring discharge and streamflow, field investigations have been undertaken, from August 2008 until August 2011. In order to represent the spatial variation of the isotope signature in precipitation, samples were taken at five different locations on bi-weekly basis from the rainfall collector locally made of plastic containers.

The rainfall sample collectors have a capacity of 10 liters fitted with vertical funnel with a mesh on a top to avoid debris and long plastic tube connected to avoid evaporation out of the collection device. Spring water was sampled at three locations at different altitudes on a weekly basis, and the two weekly samples were mixed and taken for the analysis. Streamflow was sampled at the outlet of the Chemoga and Jedeb rivers on a weekly basis. Details about isotope sample locations and investigation periods are given in Table 2.

During sampling, the water was filled into 2 mL glass bottles and closed immediately to avoid fractionation due to evaporation.

3.2 Laboratory analysis

All water samples were analyzed at UNESCO-IHE (Delft, the Netherlands) using a LGR liquid-water isotope analyzer. The stable isotopic composition of oxygen-18 and deuterium are reported using the δ-notation, defined according to the Vienna Standard Mean Ocean Water (VSMOW) with δ^{18}O and δ^{2}H [‰]; where $\delta = ((R_{\text{sample}}/R_{\text{standard}}) - 1) \cdot 1000$ and R is the ratio of the heavy and light isotopes (e.g., ^{18}O/^{16}O and ^{2}H/^{1}H).

The accuracy of the LGR liquid-water isotope analyzer measurements was 0.2‰ for δ^{18}O and 0.6‰ for δ^{2}H, respectively.

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3.3 Hydrograph separation at seasonal time scale

The classical steady state mass balance equations of water and tracer fluxes in a catchment were used in this study to separate the hydrograph into different components. The assumptions used for the hydrograph separations and the basic concepts are described in detail by e.g., Sklash and Farvolden (1979), Wels et al. (1991) and Buttle (1994).

The mass balance equation used for a time-based two component separation using (^{18}O) as a tracer can be described as:

$$Q_T = Q_E + Q_{Pe} \tag{1}$$

$$C_T Q_T = C_E Q_E + C_{Pe} Q_{Pe} \tag{2}$$

Where Q_T is the total runoff [m^3 s^{-1}], Q_E [m^3 s^{-1}] and Q_{pe} [m^3 s^{-1}] are the runoff event and pre event components, respectively. C_T is the total concentration of tracer observed in total runoff [‰ VSMOW], C_E [‰ VSMOW] and C_p [‰ VSMOW] are the tracer concentrations in event and pre-event water, respectively. Combining Eq. (1) and Eq. (2), the contribution of event water and pre-event water to the total runoff can be estimated as:

$$Q_E = Q_T \left(\frac{C_T - C_{Pe}}{C_E - C_{Pe}} \right) \tag{3}$$

$$Q_P = Q_T \left(\frac{C_T - C_E}{C_{Pe} - C_E} \right) \tag{4}$$

The precipitation isotopic composition was weighted based on the cumulative incremental weighting approach as outlined by (McDonnell et al., 1990):

$$\delta^{18}O = \frac{\sum\limits_{i=1}^{n} p_i \delta_i}{\sum\limits_{i=1}^{n} p_i} \qquad (5)$$

Where p_i and δ_i denotes the rainfall amount and δ value, respectively.

Similarly, the monthly discharge isotopic composition in the rivers were weighted using a step function by dividing the total summation of multiplied volume of flow rate in a day with the corresponding average values of the instantaneous sample isotopic values on the same day to total summation of volume of flow rate in a respective months.

Due to the distinct seasonality, the precipitation during the dry (winter) and little rain (spring) seasons are not contributing significantly to the total streamflow either as surface or subsurface flow. This is due to the fact that the precipitation in these seasons is mostly evaporating without producing direct runoff or recharging the groundwater (Kebede and Travi, 2012). To account for the effects of seasonality on the results of hydrograph separation, the end member signature is not taken as a constant value throughout the whole seasons. Consequently, we set the pre-event water isotopic composition as the monthly isotopic values at each month during the dry (winter) and the spring seasons. As the same time to see the effects of different pre-event end member concentrations on the results of hydrograph separation, the average values for the whole dry season concentration, average value of the concentration in February, which representing the baseflow in the rivers and average concentration of combined dry and spring seasons concentration were evaluated.

The event water $\delta^{18}O$ end member was taken as the weighted mean isotopic composition of precipitation, in each month for the investigated period. The differences in isotopic composition for event water, varies from -6.37 to -4.24 ‰ and pre-event water -0.25 to 0.62 ‰ are adequate for the hydrograph separation in these catchments based on the assumptions of classical hydrograph separation described in Buttle (1994).

Hydrograph separation using isotope technique is prone to error due to the uncertainty in the estimation of end member concentrations (e.g.Genereux, 1994; Uhlenbrook and Hoeg, 2003). In our study the uncertainty in the two component separations during the wet season June to September evaluated based on the Gaussian error propagation technique according to Eq. (6) (e.g. Genereux, 1994; Uhlenbrook and Hoeg, 2003).

$$W_y = \sqrt{\left(\frac{\partial y}{\partial x1} W_{x1}\right)^2 + \left(\frac{\partial y}{\partial x2} W_{x2}\right)^2 + \ldots + \left(\frac{\partial y}{\partial xn} W_{xn}\right)^2} \qquad (6)$$

Where W represents the uncertainty in the variables indicated in the subscript. Assuming that "y" is a function of the variables x_1, $x_2 \ldots x_n$ and the uncertainty in each variable is independent of the uncertainty in the others (Genereux, 1994). The uncertainty in y is related to the uncertainty in each of the subscript variables by using Eq. (6). Application of Eq. (6) in to Eq. (4) gives the propagated total uncertainty related to the different component computed using Eq. (7).

$$W = \left\{ \left[\frac{(C_E - C_T)}{(C_E - C_{Pe})^2} \cdot W_{CPe}\right]^2 + \left[\frac{(C_T - C_{Pe})}{(C_E - C_{Pe})^2} \cdot W_{CE}\right]^2 + \left[\frac{-1}{(C_E - C_{Pe})} \cdot W_{CT}\right]^2 \right\}^{\frac{1}{2}} \qquad (7)$$

Where W is the total uncertainty or error fraction related to each component and W_{CPe}, W_{CE}, W_{CT} are the uncertainty in the pre-event, event and total stream water respectively. The uncertainties related to each component are computed by multiplying the standard deviations by t values from the Student's t distribution at the confidence level of 70 % (Genereux, 1994).

3.4 Estimation of mean residence time

The mean residence time of stream water in a catchment is commonly computed using lumped parameter black box models described in (Maloszewski and Zuber, 1982).

However, the application of this method to short data records and coarse spatial and temporal sampling lead to inaccurate estimates of parameters and tracer mass imbalance if the time scale of transit time distribution is larger than the input data (McGuire and McDonnell, 2006).

Hence, due to the short record length and coarse frequency of spatial and temporal tracer sampling, in this study the mean residence time is estimated based on the sine wave approach fitting the seasonal $\delta^{18}O$ variation in precipitation and streamflow (e.g. McGuire et al., 2002; Rodgers et al., 2005; Tetzlaff et al., 2007b). The method gives indicative first approximation estimates of mean residence times (Soulsby et al., 2000; Rodgers et al., 2005). The predicted $\delta^{18}O$ can be defined as:

$$\delta = C_o + A[\cos(ct - \varphi)] \tag{8}$$

Where δ is the predicted $\delta^{18}O$ [‰] composition, C_o is the weighted mean annual measured $\delta^{18}O$ [‰], A is the annual amplitude of predicted $\delta^{18}O$ [‰], c is the angular frequency constant ($0.017214\,\text{rad}\,\text{d}^{-1}$), t is the time in days after the start of the sampling period and φ is the phase lag of predicted $\delta^{18}O$ in radians. Further Eq. (8) can be evaluated using sine and cosine terms in a periodic regression analysis (Bliss, 1970) as:

$$\delta = C_o + \beta_{\cos}\cos(ct) + \beta_{\sin}\sin(ct) \tag{9}$$

The estimated regression coefficients $\beta_{\cos}$ and $\beta_{\sin}$ are used to compute the amplitude in input, output signal ($A = \sqrt{\beta_{\cos}^2 + \beta_{\sin}^2}$) and consequently the phase lag $\tan\varphi = \left|\frac{\beta_{\sin}}{\beta_{\cos}}\right|$.

The mean residence time from the fitted sine wave in input and output signal was estimated as:

$$T = c^{-1}\left[\left(\frac{A_2}{A_1}\right)^{-2} - 1\right]^{0.5} \tag{10}$$

Where T is the mean residence time [day], A_1 the amplitude of precipitation δ^{18}O [‰], A_2 is the amplitude of streamflow δ^{18}O [‰], and c is defined in Eq. (8).

4 Results and discussion

4.1 Meteoric water

⁵ The relation between δ^{18}O and δ^{2}H isotopic composition for precipitation, spring water, and river water in the study area are shown in Fig. 3. The spatial distribution of δ^{18}O and δ^{2}H composition of precipitation at the highest elevation (Fana Choke) and at the lowest elevation station (Yewla) varies considerably along the elevation gradient. Some deviation from the global meteoric water line might be related to the effect of evapora-
¹⁰ tion of falling rain drops, condensation in the cloud and different moisture sources over different seasons (Dansgard, 1964; Gat, 1996; Levin et al., 2009; Kebede and Travi, 2012).

It is illustrated in Fig. 3 that the local meteoric water line produced from the relation-ship between δ^{18}O and δ^{2}H composition of Addis Ababa precipitation using Global
¹⁵ Network of Isotopes in Precipitation (GNIP) samples are within the same range of iso-topic variation of the present study.

4.2 Spatio-temporal variation of isotope composition in precipitation, spring water and streamflow

4.2.1 Isotope composition of precipitation

²⁰ The results of the measured isotopic composition of precipitation samples exhibit marked spatial and seasonal variations (Fig. 4). The precipitation at Yewla station (low-est altitude) shows the heaviest δ^{18}O and δ^{2}H values in contrast to the lighter isotopic composition at Fana Choke station (highest altitude). This shows the anticipated iso-

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topic composition influenced by the altitude effect (Dansgaard, 1964; Rozanski et al., 1993). Nevertheless, the altitude effect varies temporally over different seasons depending on the moisture source, amount and trajectories of air mass bringing precipitation and local meteorological settings (Aravena et al., 1999). For instance, the seasonal isotopic composition relationship with elevation along the gradient during different seasons shows different values of lighter isotopic composition at higher altitudes (Fig. 5).

The altitude effect entails to decrease the isotopic composition at higher altitudes by -0.12 ‰ and -0.58 ‰ per 100 m increase in altitude for δ^{18}O and δ^{2}H, respectively. This change of isotopic composition with elevation is consistent with earlier finding by Kebede and Travi (2012), who found that a decrease of δ^{18}O by -0.1 per 100 m in the higher elevations of the Blue Nile plateau.

The δ^{18}O and δ^{2}H composition is also affected by the precipitation amount effect (Dansgaard, 1964; Rozanski et al., 1993). Figures 6 and 7 illustrate the δ^{18}O and δ^{2}H composition of precipitation at sampling stations, which show moderate regression coefficients ranging from ($R^2 = 0.36$–0.68) for precipitation and ($R^2 = 0.26$–0.39) for temperature. This suggests that the amount effect at each sampling locations is important for the variation of isotopic composition in the area in addition to other factors. However our results are in contrast to the earlier studies by Kebede (2004), and Kebede and Travi (2012), who reported that weak relationships between rainfall amounts and isotopic compositions in the North-Western Ethiopian plateau.

Moreover, multiple regression models are used to show the effect of monthly precipitation and mean monthly temperature on δ^{18}O and δ^{2}H isotopic composition of precipitation in Chemoga and Jedeb catchments, respectively. The multiple regression models for δ^{2}H composition in Chemoga and Jedeb catchments are described as: δ^{2}H$= -0.096P + 2093T + 0.736$ ($R^2 = 0.74$, $n = 28$, p value $= 0.001$ for precipitation and $p = 0.121$ for temperature) in Chemoga catchment and δ^{2}H $= -0.116P + 2.414T - 1.374$ ($R^2 = 0.76$, $n = 28$, p value $= 0.001$ for precipitation and $p = 0.175$ for temperature evaluated at 5 % significance level) in Jedeb catchment. Where P in the regression equa-

tion is the monthly precipitation ($mm\,month^{-1}$) and T is the mean monthly average temperature (°C). These results from the multiple regression models also support that the amount effect has a more dominant role for the variations in the isotopic composition in the study area than the temperature effect.

Table 3 presented the mean, minimum, maximum and standard deviation of the amount weighted precipitation and volume weighted discharge data and the non weighted composition for spring water. It demonstrates that lighter isotopic compositions are observed during the main rainy season from June to September and the heavier values are observed during winter and spring seasons. This is obviously related to the multiple moisture sources and to the local meteorological processes (Kebede and Travi, 2012).

The seasonal variations in isotopic composition of precipitation are observed among the stations. For instance, the winter seasonal mean weighted δ^{18}O composition of precipitation is with a value of −0.41 ‰ lighter at Fana Choke at a higher altitude and has a heavier value of 3.08 ‰ at the lowest altitude (Yewla station). During spring season the mean weighted δ^{18}O composition is −0.62 ‰ at Fana Choke and 3.3 ‰ at Yewla. Similarly, during summer a lighter isotopic composition of −3.28 ‰ observed at Fana Choke and relatively heavier isotopic composition −1.9 ‰ is observed at Yewla.

In the Chemoga catchment the mean weighted seasonal isotopic composition of δ^{18}O and δ^{2}H in precipitation during winter, spring and summer are 0.72 and 24.85 ‰, 0.86 and 23.71 ‰, and −2.09 and 2.36 ‰, respectively. Obviously the summer seasonal isotopic composition in both δ^{18}O and δ^{2}H are lighter than in the winter and spring seasons owing to the different moisture sources and the local meteorological settings. In comparison to Chemoga catchment, the mean weighted seasonal isotopic composition in Jedeb catchment shows consistently heavier δ^{18}O and δ^{2}H isotopic values of 1.48 and 29.23 ‰, 1.65 and 28.30 ‰, and −1.93 and 2.74 ‰ in winter, spring and summer seasons, respectively. This implies that the heavier isotopic values of precipitation are likely related to different temperatures and altitudes in the Jedeb catchment.

4.2.2 Isotope composition of spring water

The isotopic composition of spring water at the three locations shows distinct variability ranging from −8.5 to 13.5 ‰ and −4.1 to 2.9 ‰ for δ^2H and δ^{18}O, respectively. The spring water of Debre Markos (at elevation of 2339 m a.s.l) and Rob Gebeya (at 2820 m a.s.l) exhibit lighter isotopic composition as compared to the isotopic composition of spring water of Yewla (at elevation of 2255 m a.s.l), which showed heavier isotopic compositions. This is suggesting that the recharge areas for the spring discharges are located at different altitudes. The mean raw isotopic composition of spring water indicates a wide variation at the three locations. The observed mean isotopic variation at the tree locations ranged from −0.6 to 5.5 ‰ for δ^2H and −2.1 to −0.7 ‰ for δ^{18}O. The mean values for δ^{18}O at Debre Markos and Rob Gebeya exhibit similar isotopic composition of −2.1 ‰.

It is interesting that the isotopic composition for the springs at Debre Markos and Rob Gebeya follow similar patterns and exhibit no major distinction in their isotope composition (Fig. 8). This indicates that the spring water isotopic composition for both springs derived from the same altitude range of the recharge area. The mean seasonal isotopic variations at the three spring locations during winter season ranged between −0.2 to 4.7 ‰, and −2.1 to −0.7 ‰ for δ^2H and δ^{18}O, respectively. During spring season the mean seasonal isotopic variations ranged between −2.6 to 6.3 ‰, and −2.4 to −0.5 ‰ for δ^2H and δ^{18}O, respectively. During summer season the mean seasonal isotopic variations ranged between 0.2 to 5.9 ‰, and −2.0 to −0.9 ‰ for δ^2H and δ^{18}O, respectively.

The mean winter and spring seasonal δ^{18}O isotopic composition of precipitation exhibit heavier values greater than 0 ‰ for all five stations (see Sect. 4.2.1), except the more negative values at the highest altitude (Fana Choke station). In contrast to the precipitation signature during these seasons, the spring waters exhibit isotopically lighter composition. This is suggesting that the spring water during winter and spring seasons are merely derived from summer season precipitation in which main recharge area of

the spring water aquifer is in the highlands (see Fig. 9). This in turn reflects the winter and spring season precipitation does not contribute to recharge the groundwater. This finding is in agreement with the previous studies in the region (e.g., Kebede et al., 2003; Kebede and Travi, 2012). They pointed out that during the dry and spring seasons most

5   of the water is evaporated without contributing to the groundwater recharge.

Furthermore, the damped spring water isotopic signature as compared to the river water gives a hint that the spring water could be a mixture of old water components having longer residence times than the river water.

The results of the seasonal variations of isotopic composition of different water sam-

10   ples shown in Fig. 9 are due to different moisture sources. These different moisture sources are confirmed by an investigation of the potential source areas of precipitation for different seasons. Figure 10 presents the potential source areas of precipitation, whereby the starting points of trajectories to the study area were computed using the HYSPLIT model (Hybrid Single Particle Lagrangian Integrated Trajectory Model) devel-

15   oped by NOAA (National Oceanic and Atmospheric Administration) at Air Resources Laboratory (www.arl.noaa.gov/HYSPLIT_info.php). The model computes the trajectories by tracing back an air package for 14 days in different seasons.

It is shown in Fig. 10 that during the three seasons, the source areas for starting points of moisture trajectories into the study area are different. During the main rainy

20   season, i.e. the summer, the Atlantic Ocean, Indian Ocean, the White Nile and the Congo basin are the likely potential source areas of precipitation in the study area. On the other hand, in the spring and winter seasons, the potential source areas of moisture origin that seems responsible for generating the little precipitation in study area is the Arabian Sea–Mediterranean sea and to some extent the Indian Ocean.

25   These results are in agreement with the earlier findings in the region with regard to the source areas from the Atlantic Ocean, Indian Ocean, and the Congo basin (e.g. Levin et al., 2009; Kebede and Travi, 2012). Moreover, a recent moisture transport study by Viste and Sorteberg (2013) in the Ethiopia highlands reported that the moisture flow from the Gulf of Guinea, the Indian Ocean and from the Mediterranean region across

the Red Sea and the Arabian Peninsula are identified as the main sources for moisture transport in the region.

4.2.3 Isotopic composition of river water

In the Chemoga catchment the mean volume weighted δ^{18}O and δ^2H compositions of river water are −1.4 and 2.7 ‰ ; for winter or dry season 0.1 and 6.2 ‰ ; for the spring little rainy season −0.7 and 11.7 ‰, and for the summer long rainy season −2.3 and −3.3 ‰, respectively.

While in the Jedeb catchment the mean volume weighted δ^{18}O and δ^2H composisitions in river water are −0.6 and 4.9 ‰ ; for dry season (winter) 0.12 and 6.3 ‰ ; for the little rainy season (spring) −0.3 and 8.5 ‰, and for the summer long rainy season −1.3 and 1.7 ‰, respectively. These results show that in all seasons except for the δ^2H composition in spring season, the Jedeb river water exhibit a heavier isotope composition as compared to the Chemoga river. Moreover, the damped response of the isotope signature during the summer season in Jedeb river as compared to Chemoga river might reveal the variation in catchment storage (see Fig. 11). Water balance Study by Tekleab et al. (2011) in these catchments has also shown their hydrological differences.

Figure 11 presents the temporal variation of δ^{18}O and δ^2H for Chemoga and Jedeb catchments. In the figure the isotope composition during the main rainy season reflects damped characteristics as compared to the fluctuations in precipitation as it was observed by (e.g. McDonnell et al.,1990; Buttle, 1994; Soulsby et al., 2000). These investigations indicate that the damping behavior of the isotope signal in streamflow is due to the fact that the pre-event or old water component of the groundwater is a mixture of many past precipitation events and resulted in an isotopic concentration which is higher than the precipitation composition during storm events. The same holds true during summer months, when rainfall generates the highest flows, in the hydrological year 2010 at both catchments, the isotope composition of river water exhibit damped response as compared to the precipitation responses (Fig. 11).

4.3 Hydrograph separation at a seasonal time scale

Figure 12 presents the hydrograph separations based on the two component mixing model at seasonal time scale.

The results of the two component seasonal hydrograph separations reveal the event water proportion is more dominant than the pre-event component in both catchments in particular during the rainy season. The proportion of the summer (rainy season) monthly variation of the event water component varies from 32–99 % with an average of 71 % in Chemoga catchment and 31–96 % with an average of 64 % in the Jedeb catchment over three different wet season of the investigation period. Obviously, pre-event water is almost the sole contribution during the other seasons.

Similarly, the average proportion of the different runoff components due to different end members (i.e. the whole dry season average concentration and average of dry and little rainy season concentration) exhibit higher proportion of event water varies from 62–67 % and 33–38 % for the pre-event water in Chemoga catchment. In the Jedeb catchment the event water component varies from 52–55 % and the pre-event water varies from 45–48 %, respectively.

The proportion of the new water during the three wet seasons in both catchments is more towards the rising limb of the hydrograph. This implying that the new water component generated as a surface hydrological flow pathways in both of the catchments has a greater proportion than the pre-event component. This is in line with visual observations of flow paths and erosion. However, it is noted that the pre-event water dominates more after the main event water peak.

It can be assumed that the highest percentage of event water in both catchments s due to the lower infiltration rate and the compaction of the top soil in the agriculture land. Research in the vicinity of these catchments also suggests that the effect of a plough pan due to long years of ploughing activities reduce the infiltration capacity of the soil (Temesgen et al., 2012). The effect of relief and land degradation could also be another factor for the large proportion of the event water component (Teferi et al.,

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2013). Nonetheless, during winter (dry season) and spring (small rainy season) the river water at both catchments are solely derived from the groundwater recharged during the rainy (summer) season. Similar studies in US from small agricultural dominated catchment showed that event water has large proportion of runoff component due to

5   low infiltration rates of agricultural compacted soils (Shanley et al., 2002).

Event based two component hydrograph separation models have been widely applied in different humid catchments having smaller spatial scales (e.g. Sklash and Farvolden, 1979; Pearce et al., 1986; McDonnell, 1990). The results of these studies indicated that pre-event or old water is the dominant runoff component. Though, the

10 application of the method to semi-arid catchments in Africa is limited (e.g. Mul et al., 2008; Hrachowitz et al., 2011a; Munyaneza et al., 2012). The outcomes from these researches also suggest that pre-event water is the dominant runoff component. However, for the present study, the application in agricultural dominated catchment with high soil erosion affected area, steep slope (e.g. varying between 2 % in the lower part

15 of the catchment and more than 45 % in the upper part) and higher seasonality in climate, the event water proportion is the dominant runoff component at the seasonal time scale.

Uncertainty analysis of the hydrograph separations

The results of uncertainty analysis taking the average value of the streamflow concen-

20 tration in February for the investigation period as a pre-event end member concentration related to the different fraction of streamflow components in wet season at 70 % (approximately one standard deviation) confidence interval is presented in Table 4.

The average uncertainty terms of February concentrations as pre-event end member for the whole wet season is relatively low as compared to the different pre-event

25 end member concentrations. This lower average uncertainty term accounted for ±0.46 for Chemoga and ±0.66 for Jedeb catchment. The higher uncertainty in the Jedeb in June 2009 and September 2010 might be related to the smaller difference between the pre-event and event water composition.

The average uncertainty terms for pre-event, event and the river water for the three wet season periods accounted for 7, 61 and 32 % for Chemoga catchment and 4, 51 and 45 % for the Jedeb catchment, respectively. This is suggesting that most of the uncertainty stem from the event water component. Genereux (1994) pointed out that the greater uncertainty is mainly accounted from the proportions, which gives the highest runoff components.

The error in hydrograph separation originates from different sources (see for details Uhlenbrook and Hoeg, 2003). Based on the uncertainty results, the proportion of the different components using different pre-event end member concentrations gives only a range of values not the exact number. Thus, due to spatial and temporal variation of the end member concentrations, the classical hydrograph separations methods give qualitative description of the runoff components and their variable contributions in time (Uhlenbrook and Hoeg, 2003).

4.4 Estimation of mean residence times

Figure 13 presents precipitation and streamflow seasonal $\delta^{18}O$ patterns for the estimation of the mean residence time of water based on the periodic regression analysis to fit the seasonal sine wave models.

Preliminary estimation of mean residence time was obtained using the model described in Eqs. (8)–(10) and results are provided in Table 5. Based on the seasonal variation of $\delta^{18}O$ both in precipitation and streamflow, the mean residence time is estimated as 4.1 and 6.0 months in Chemoga and Jedeb catchments, respectively. The results of short mean residence times in both catchments are in line with the hydrograph separations, which indicate more surface runoff generation than base flow contribution during the storm events in these steep headwater catchments.

The results of preliminary estimation of mean residence time are plausible and anticipated from steep agriculture dominated catchments with little adoption of soil and water conservation measures, which enhance more surface runoff generation (Temesgen et al., 2012). Consequently, the surface condition or the responsiveness of the soil

due to the plough pan effect could influence the ability of the soil to infiltrate the given rainfall amount to recharge the groundwater system.

Furthermore, studies in the north-western Ethiopian plateau reported that the groundwater in the area is characterized by shallow and rapid circulation, led to a young age of the groundwater system, which may considered to be the cause for drying of groundwater wells after prolonged droughts (Kebede, 2004). However quantification of the age of groundwater is not yet known in the area, and needs further research on age dating.

However, due to heterogeneities in the catchment, climatic setting, soil type, geology and land cover properties, direct comparisons is not possible with similar meso-scale catchment studies in different regions.

5 Conclusions

Characterization of stable isotope composition of precipitation, spring and river water along different altitude gradient were undertaken with the aim of preliminary estimation of mean residence time and hydrograph separation at seasonal time scale. The results show that precipitation, stream and spring waters exhibit noticeable spatial and temporal variations in stable δ^{18}O and δ^2H composition in the study area.

The results further demonstrate that the meteoric water in the study area is influenced by the amount and to less extent by the altitude and temperature effects. The climatic seasonality, which is dominated by different moisture sources along with the local meteorological settings play a significant role for the isotopic composition of rainfall in the area.

The analyses of isotope results reveal the dominance of event water and short mean residence times in both of the catchments. From the point view of managing the water resources and the importance of the available soil water for consumptive use of the crops, catchment management aiming at reducing overland flow/soil erosion and

increasing soil moisture storage and recharge have paramount importance for the farmers residing in these catchments.

It should be noted that the estimated mean residence times and seasonal hydrograph separation represent first approximations and hence there is a need for further research with finer resolution sampling during storm events and long-term isotope tracer data collection at different spatial and finer temporal scales (e.g. daily and hourly) that will improve our understanding of how these catchments are functioning. It is noteworthy that the applied methods were used the first time in the region that has great regional importance regarding the water resources in the Nile. Thus, the results can be used as a baseline for similar hydrological studies for better understanding of the dominant runoff components in the future.

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Table 1. Descriptions of hydro-meteorological characteristics of investigated catchments (2008–2010). P, Q, E_p, E stand for catchment average precipitation, runoff, potential evaporation and actual evaporation, respectively.

Catchment	Area (km^2)	Mean annual values (mm a^{-1})				Mean annual temperature (°C)
		P	Q	E_p	E	T
Chemoga	358	1303	588	1338	715	13.9
Jedeb	296	1306	692	1384	614	15

Table 2. Description of isotope sample locations and total number of samples taken during the investigation period (August 2008–August 2011).

Sample type	Location name	Abbreviation	Elevation (m a.s.l)	Number of samples	Investigation period
Precipitation	Debre Markos	P-DM	2515	58	Aug 2008–Aug 2011
Precipitation	Enerata	P-EN	2517	24	Jul 2009–Aug 2011
Precipitation	Rob Gebeya	P-RG	2962	53	Oct 2008–Aug 2011
Precipitation	Fana Choke	P-FC	3993	41	Jul 2009–Aug 2011
Precipitation	Yewla	P-YW	2219	46	Jul 2009–Aug 2011
Spring	Debre Markos	S-DM	2339	64	Aug 2008–Aug 2011
Spring	Rob Gebeya	S-RG	2820	53	Oct 2008–Aug 2011
Spring	Yewla	S-YW	2255	59	Jul 2009–Aug 2011
Stream	Chemoga	Q-CH	2402	83	Jul 2009–Jul 2011
Stream	Jedeb	Q-JD	2190	98	Jul 2009–Jul 2011

Table 3. Mean, range and standard deviation of δ^2H and δ^{18}O [‰] amount weighted concentration for precipitation and volume weighted for discharge and non-weighted for spring water during different investigation period.

Description	Mean [‰ VSMOW]		Minimum [‰ VSMOW]		Maximum [‰ VSMOW]		Standard deviation [‰ VSMOW]	
	δ^2H	δ^{18}O	δ^2H	δ^{18}O	δ^2H	δ^{18}O	δ^2H	δ^{18}O
Precipitation at Yewla	22.5	1.0	−25.5	−5.7	67.8	7.8	24.7	3.5
Precipitation at Debre Markos	12.3	−0.6	−37.1	−7.0	74.6	4.6	18.4	2.4
Precipitation at Enerata	15.4	−0.8	−7.1	−3.7	29.2	1.3	11.9	1.7
Precipitation at Rob Gebeya	23.0	1.0	−36.5	−6.5	54.8	5.9	21.8	3.1
Precipitation at Fana Choke	8.5	−1.8	−49.3	−9.1	48.3	2.2	25.5	3.0
Chemoga catchment precipitation	15.5	−0.4	−29.8	−6.4	40.2	3.4	16.7	2.2
Jedeb catchment precipitation	18.3	0.13	−28.4	−6.1	45.9	4.4	18.6	2.5
Chemoga discharge	2.7	−1.4	−15.5	−3.9	19.8	0.8	9.2	1.5
Jedeb discharge	4.9	−0.6	−3.3	−3.5	13.1	0.8	4.6	1.1
Yewla spring water	5.7	−0.7	−8.0	−3.9	13.5	2.9	4.0	0.9
Debre Markos spring water	0.1	−2.1	−8.5	−4.1	7.2	−0.6	5.8	0.6
Rob Gebeya spring water	−0.6	−2.1	−9.0	−3.1	5.1	−0.8	3.3	0.5

Table 4. The proportion of event and pre-event components of streamflow with the total uncertainty for the variation of concentrations at 70 % confidence level.

Month	Chemoga catchment streamflow proportion			Jedeb catchment streamflow proportion		
	Event water, %	Pre-event water, %	Total uncertainty, ‰	Event water, %	Pre-event water, %	Total uncertainty,‰
Jul 2009	94.59	5.41	±0.98	96.46	3.54	±0.90
Aug 2009	79.36	20.64	±0.31	68.44	31.56	±0.34
Sep 2009	32.03	67.97	±0.28	69.34	30.66	±1.68
Jun 2010	57.45	42.55	±0.38	89.30	10.70	±1.10
Jul 2010	99.72	0.28	±0.52	58.28	41.72	±0.37
Aug 2010	64.62	35.38	±0.27	34.91	65.09	±0.22
Sep 2010	38.69	61.31	±0.66	31.23	68.77	±0.58
Jun 2011	67.67	32.33	±0.24	67.56	32.44	±0.66
Jul 2011	88.75	11.25	±0.51	62.06	37.94	±0.34
Aug 2011	89.88	10.12	±0.44	64.86	35.14	±0.38

Table 5. Amount weighted mean precipitation and flow δ^{18}O composition, estimated amplitude, phase lag φ and mean residence time in Chemoga and Jedeb catchments over the period July 2009–August 2011.

Description	Mean annual measured δ^{18}O [‰]	Amplitude [‰]	Phase lag φ [radian]	Mean residence time [months]
Chemoga catchment precipitation	−0.58	4.47	1.11	
Chemoga discharge	−1.34	1.89	0.01	4.1
Jedeb catchment precipitation	0.40	4.34	1.12	
Jedeb discharge	−0.70	1.35	0.01	6.0
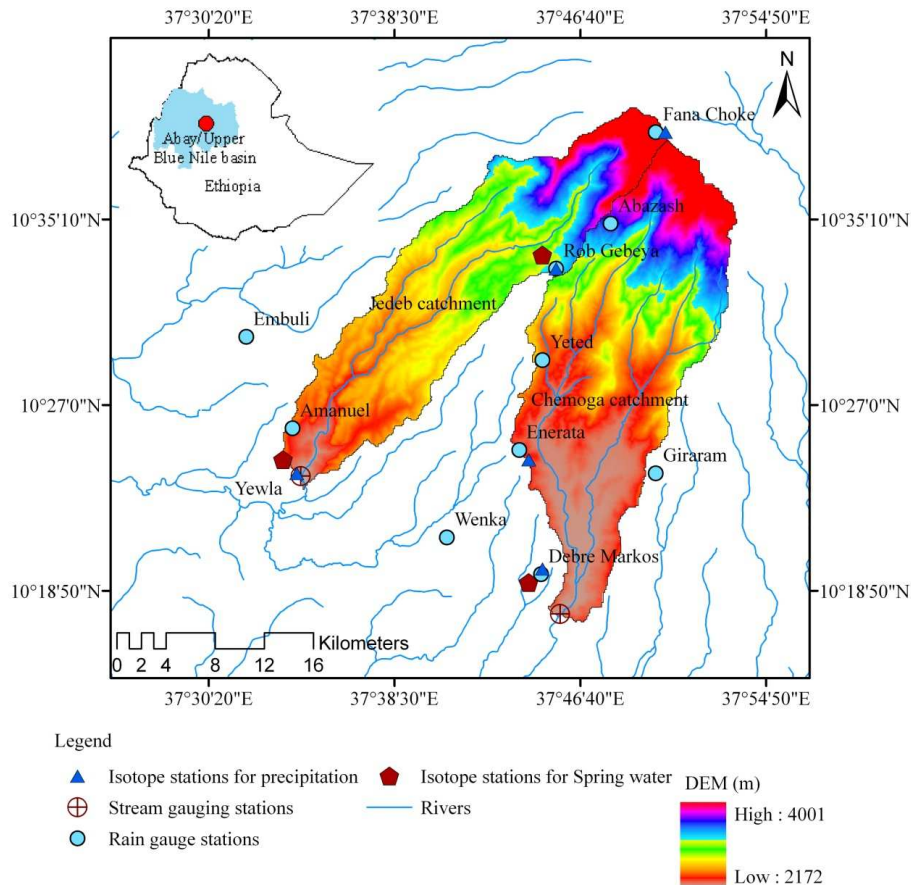
Fig. 1. Location of the study area indicating the network of rain gauges, streamflow gauges and sampling points for stable isotopes of precipitation, surface water and spring water. The red dot within Ethiopian map (inset, top left) indicates the location of Chemoga and Jedeb catchments.

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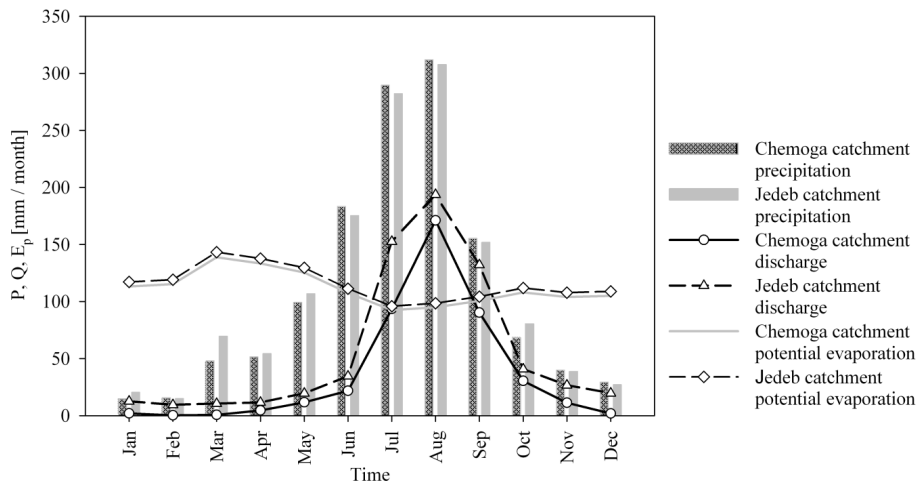
Fig. 2. Intra-annual variability of hydro-climate data for the period 2008–2010 showing similar climate and distinct streamflow response in Chemoga and Jedeb catchments. P, Q and E_p in the y-axis stand for precipitation, discharge and potential evaporation, respectively.
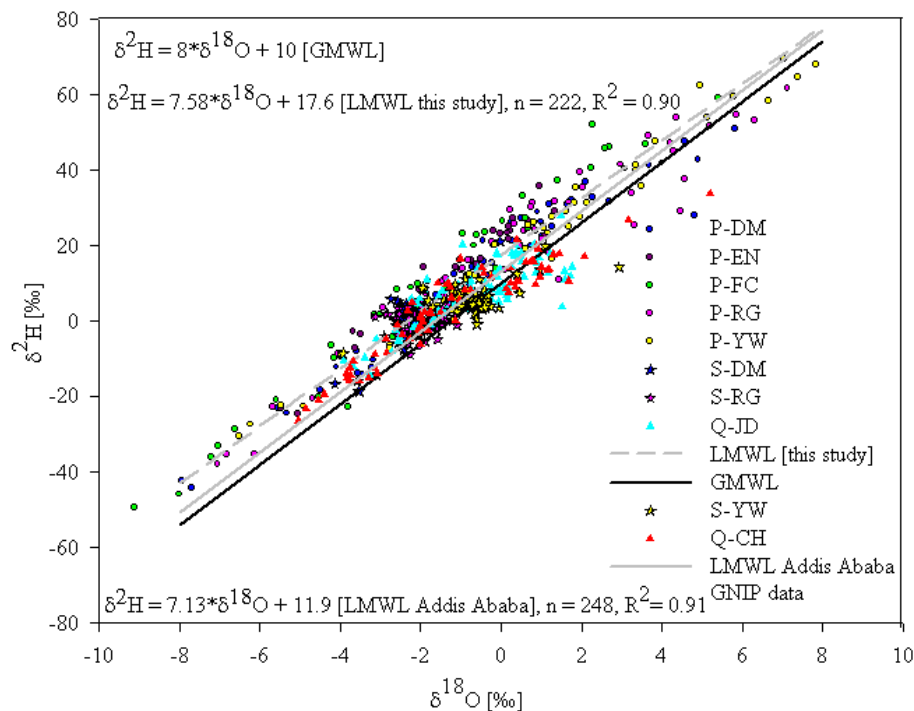
Fig. 3. Relationship between, δ^{18}O and δ^2H for precipitation, stream and spring water in the study area. The abbreviations in the legend are described in Table 2.
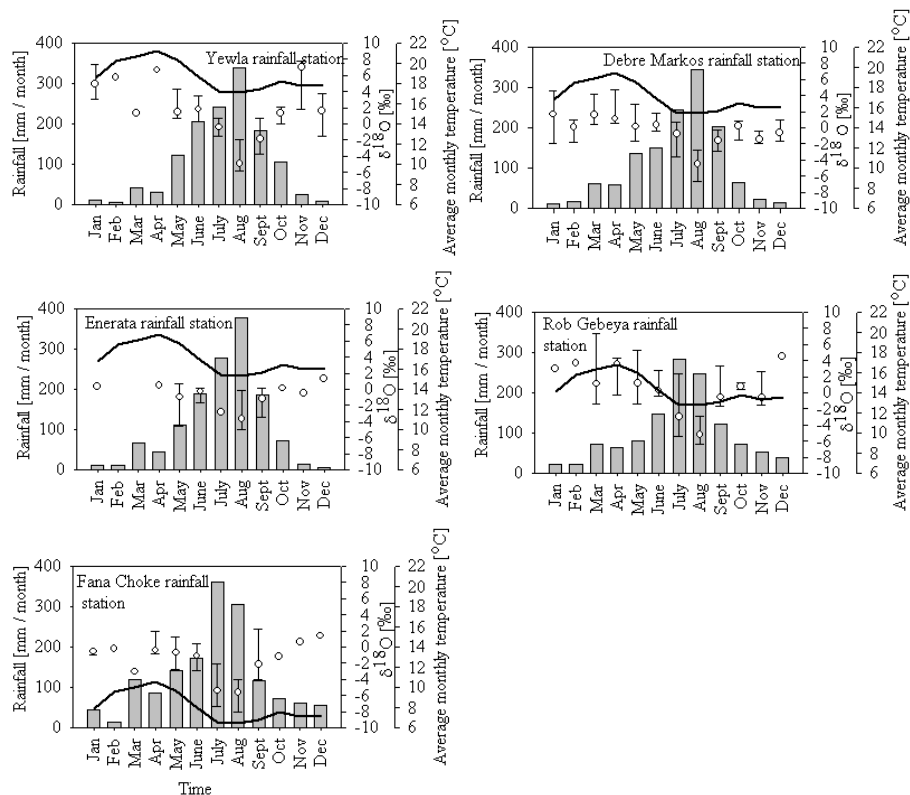
Fig. 4. Spatial and intra-annual variations of average monthly precipitation, temperature and isotopic composition of δ^{18}O in precipitation. The error bar of the isotopic measurements stand for the standard deviation. The gray bar, black solid line and open circle with error bar are the precipitation, temperature and isotopic composition, respectively. For the isotopic composition, the open circle, the lower and upper error bars indicate the median, the 25 and 75 percentiles for the raw (non-weighted) precipitation isotope samples data, respectively.

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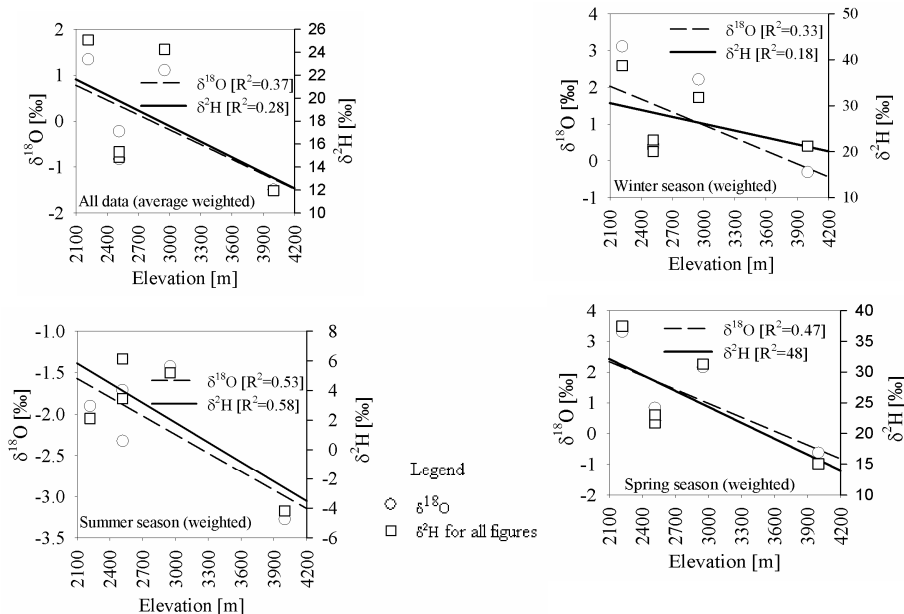
Fig. 5. Relationships between average and seasonal amount weighted isotopic composition of precipitation with elevation at five precipitation sampling stations.
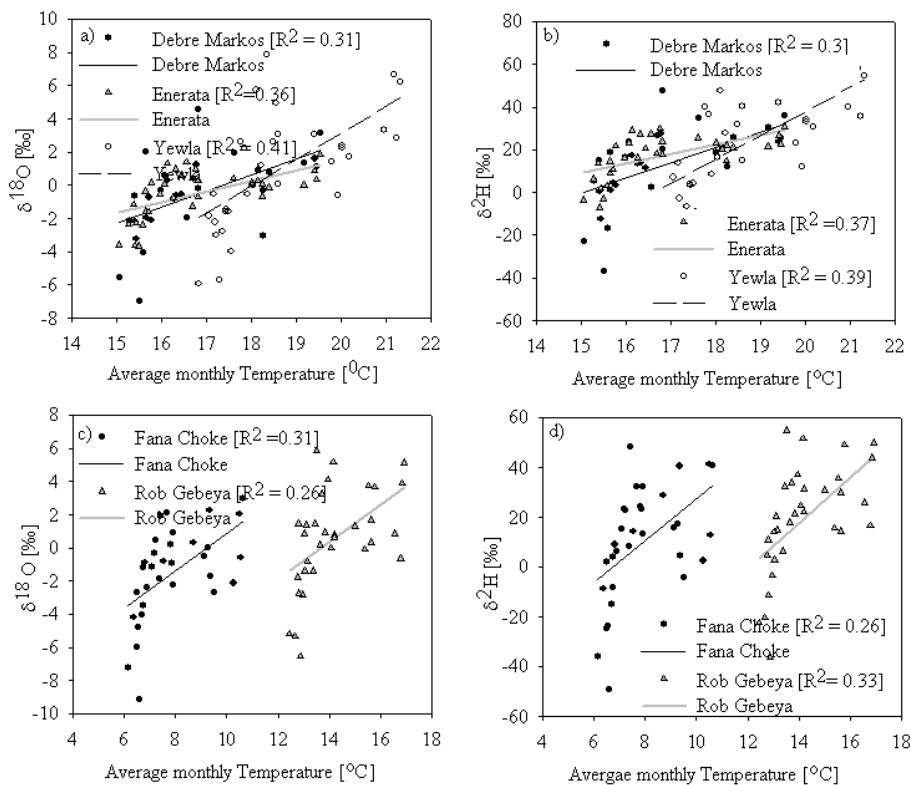
Fig. 6. Relationships between amount weighted isotopic composition of precipitation samples at different stations with monthly average temperature at respective stations during the investigation period.
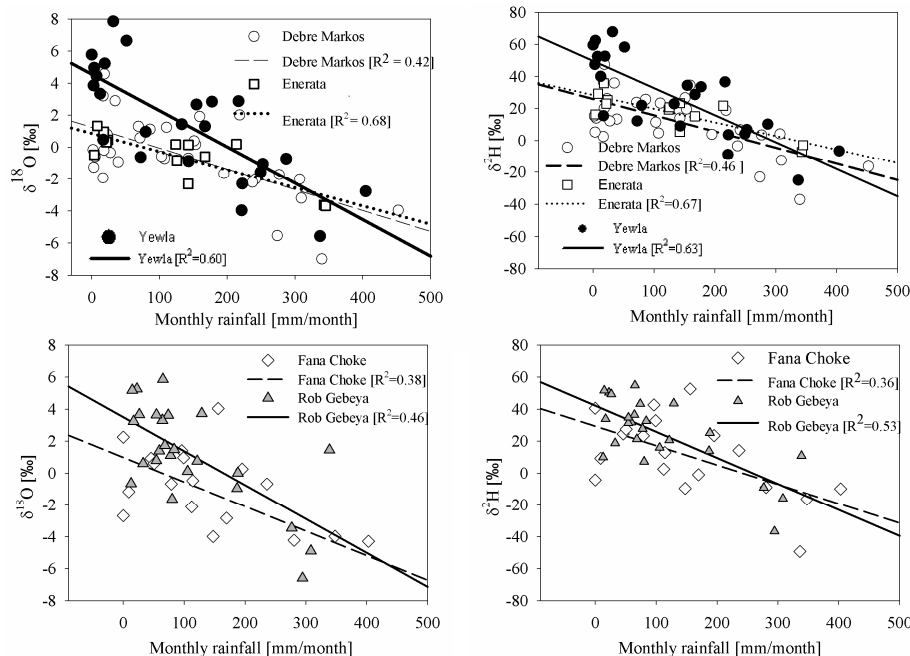
Fig. 7. Relationship between amount weighted isotopic composition of precipitation samples at different stations with monthly precipitation amounts at respective stations during the investigation period.
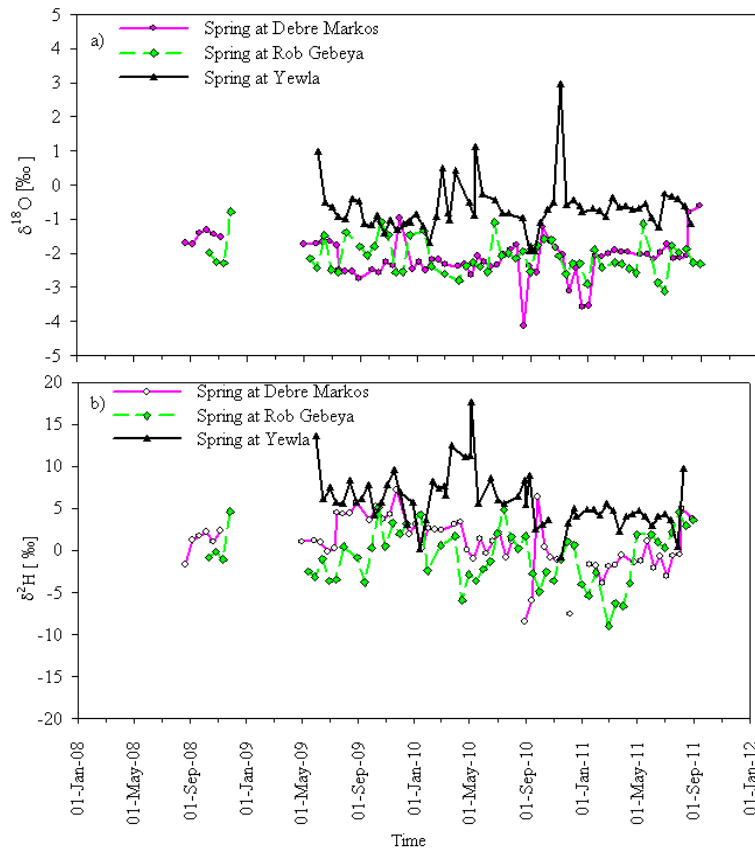
Fig. 8. Temporal isotopic variability of three different non-weighted spring water samples: **(a)** δ^{18}O and **(b)** δ^2H composition over different investigation periods.
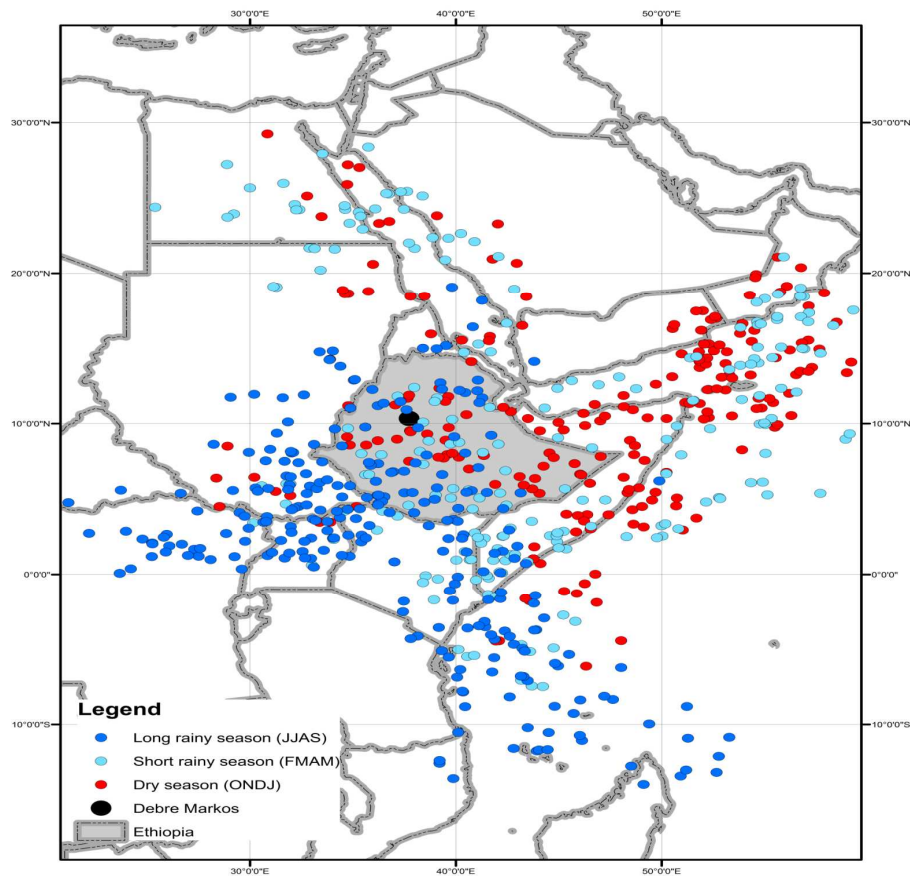
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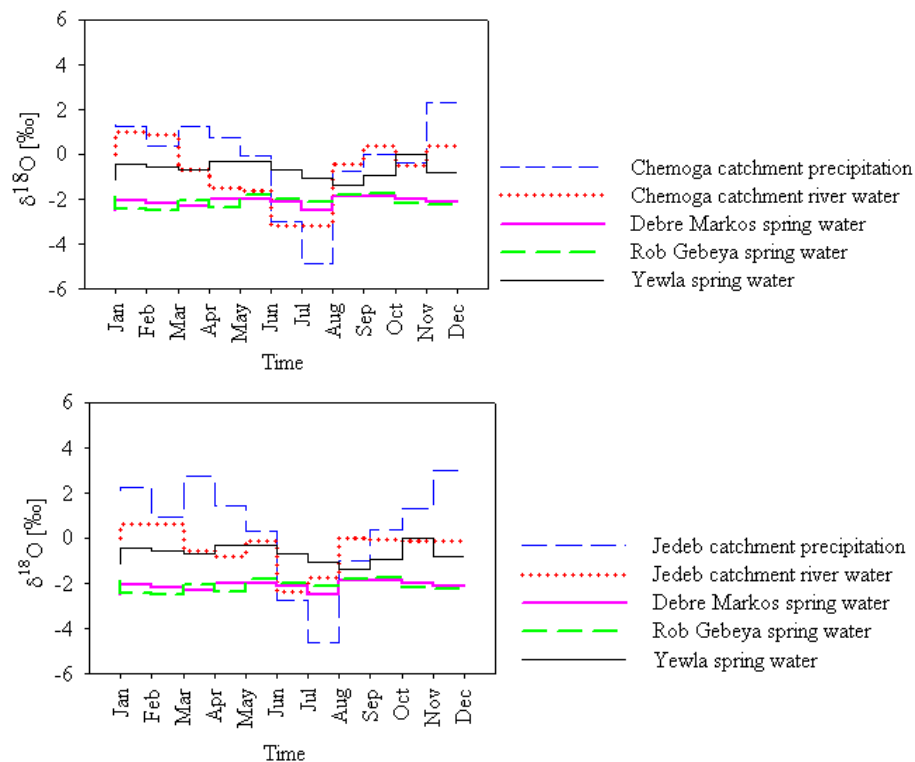
Fig. 9. Monthly δ^{18}O [‰] isotopic variation of amount weighted precipitation, volume weighted discharge and non weighted spring water (top panel) Chemoga and (bottom panel) Jedeb catchment for the period July 2009–August 2011.

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Fig. 10. Potential source areas of precipitation to the study area in different seasons. Points indicate the starting points of 14 days backward calculated trajectories. The black dot indicates the location of the study area.

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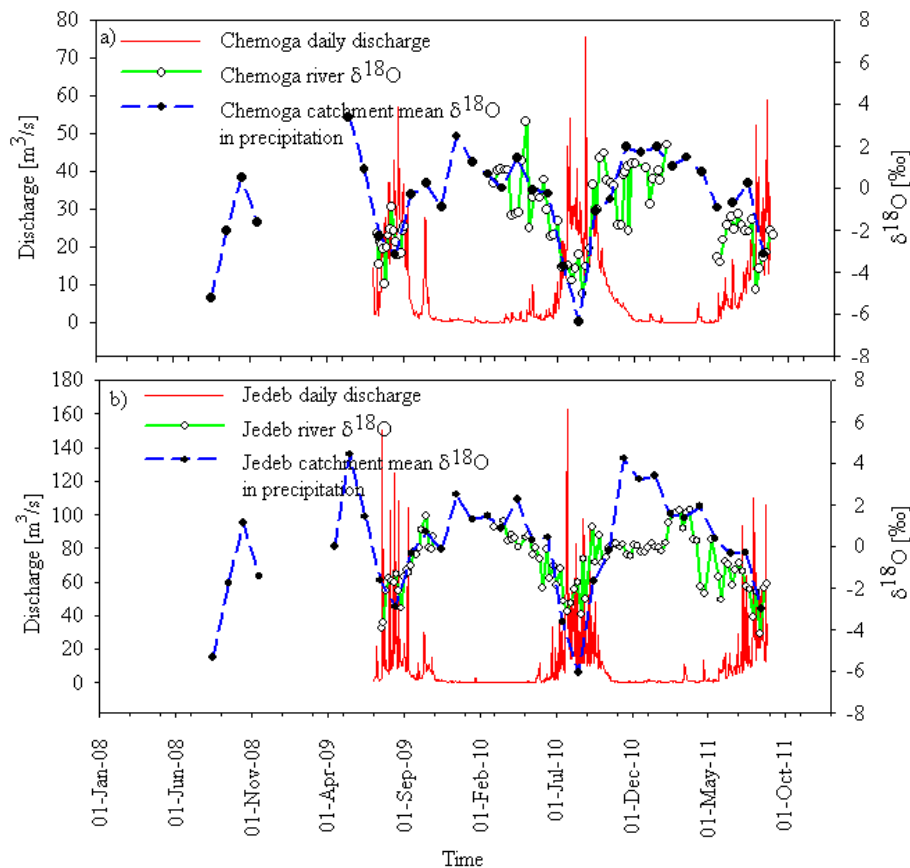
Fig. 11. Temporal variations of δ^{18}O and δ^2H composition in precipitation and river discharge along with the daily flow rate for **(a)** Chemoga catchment and **(b)** Jedeb catchment.

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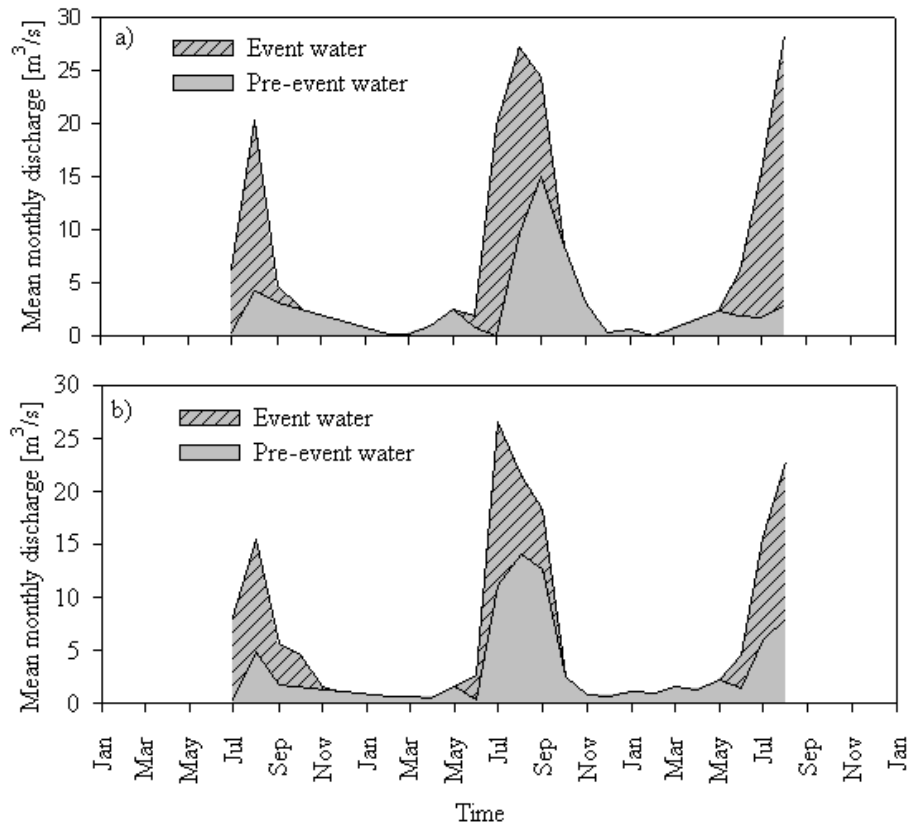
Fig. 12. Two-component hydrograph separations using δ^{18}O as a tracer in **(a)** Chemoga and **(b)** Jedeb meso-scale catchments at a seasonal time scale over the period July 2009–August 2011.
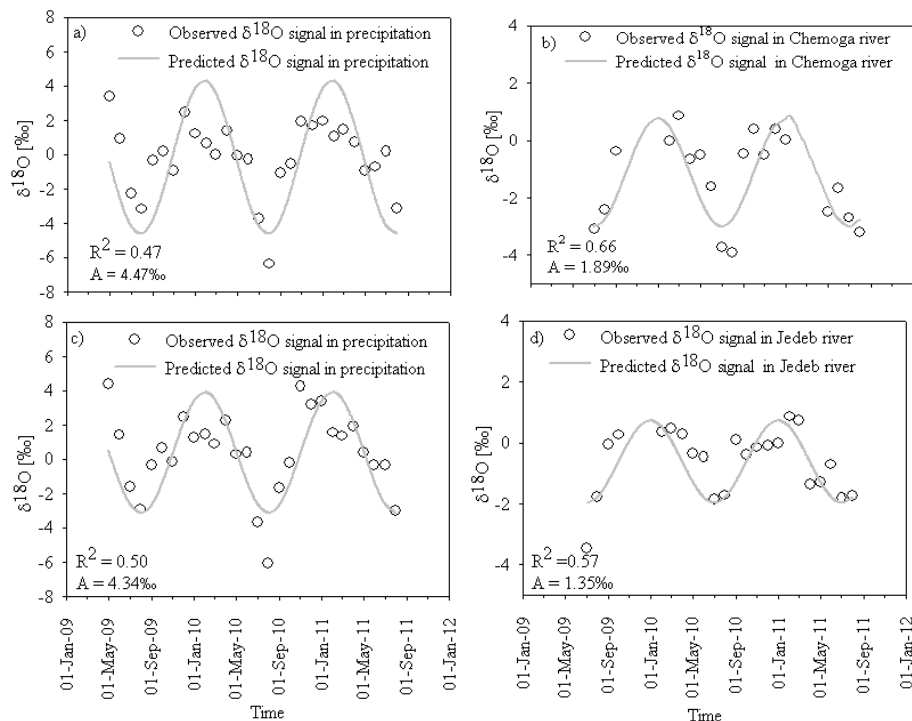
Fig. 13. Fitted sine wave regression models to $\delta^{18}O$ values for precipitation and river water **(a)** and **(b)** in the Chemoga, and **(c)** and **(d)** in the Jedeb river.