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Application of Water Isotopes to Identify the Sources of Groundwater Recharge in a Karstified Landscape of Western Himalaya

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Abstract: The spatial and temporal distribution of $\delta^{18}O$ and δD measurements of precipitation and groundwater were used to identify the recharge areas of groundwater/springs in a mountainous catchment of the western Himalaya. The $\delta^{18}O$ and δD of precipitation showed marked spatial and seasonal variability with $\delta^{18}O$ and δD varied from 0.1‰ to –13.0‰ and 1‰ to –74‰ with an average of –6‰ and –38‰, respectively. $\delta^{18}O$ and δD of precipitation is strongly influenced by the basin relief and meteorology. The mean altitude gradient of –0.15‰ and –1.16‰ per 100 m change in elevation for $\delta^{18}O$ and δD, respectively, was observed based on amount weighted mean precipitation isotopic values. The $\delta^{18}O$ and δD in groundwater showed a narrow spatial and temporal variation in comparison to precipitation and varied from –6.8 ‰ to –10 ‰ and –58‰ to –38‰, with an average of –8‰ and –46‰, respectively. The most depleted (in heavier isotopes) isotopic values were observed in karst springs and most enriched (in heavier isotopes) isotopic values were observed in shallow groundwater samples. The results suggest that the groundwater with most depleted isotopic values have recharge areas at higher altitudes, whereas the groundwater with less depleted isotopic values have recharge areas at lower elevations. The climate change has shown the reduced snowfall and annual discharge in perennial karst springs i.e. 40-70%, besides seasonal springs dried up in recent years.

Keywords: Stable isotopes; Groundwater recharge; Precipitation, Western Himalaya.

Basic Information

In our solar system the earth is a unique planet where water (H_2O) exists in liquid form on its surface (Faure and Mensing, 2009), which is the prerequisite for the existence of life. Hydrogen (H) is the most abundant chemical element in our solar system and oxygen (O) is the most abundant element in the Earth's crust. In nature, there are two stable isotopes of hydrogen: ^{1}H (protium) and ^{2}H (deuterium) and three stable isotopes of oxygen: ^{16}O, ^{17}O and ^{18}O. ^{1}H and ^{16}O are the lightest and most abundant stable isotopes of these elements (Lide and Frederikse, 1995). ^{2}H is 99.8% heavier than ^{1}H and ^{18}O is 12.5% heavier than ^{16}O. The existence of the stable isotopes of hydrogen and oxygen makes nine isotopically different water molecules: $^{1}H_2{}^{16}O$, $^{1}H_2{}^{17}O$, $^{1}H_2{}^{18}O$, $^{2}H_2{}^{16}O$, $^{2}H_2{}^{17}O$, $^{2}H_2{}^{18}O$, $^{1}H^{2}H^{16}O$, $^{1}H^{2}H^{17}O$ and $^{1}H^{2}H^{18}O$. The difference in the masses of these water molecules is the basis for the isotope fractionation to occur. Out of these nine isotopically different water molecules, only three water molecules occur in nature in easily measurable concentrations: $^{1}H_2{}^{16}O$, $^{1}H_2{}^{18}O$ and $^{1}H^{2}H^{16}O$ (Araguás-Araguás et al., 2000). The Vienna Standard Mean Ocean Water

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(V-SMOW) is the internationally accepted standard for the measurement of hydrogen and oxygen isotopes of natural water samples (Coplen et al., 1996). The isotopic ratios of $^{2}H/^{1}H$ and $^{18}O/^{16}O$ of V-SMOW were found to be $155.95\pm0.08\times10^{-6}$ (De Wit et al., 1980) and $2005.20\pm0.45\times10^{-6}$ (Baertschi, 1976). These values are close to the average concentrations of isotopes in ocean water given by Craig and Gordon (1965).

The oceans, covering 71% of the Earth's surface, accounts for 97.25% of the mass of water (Gat, 2010). Most of the freshwaters, whose volume is estimated to be 39×10^{6} km^{3}, are not easily accessible and is locked in glaciers and ice caps (Gat, 2010). Groundwater although constitutes a small percentage (1.7%) of all of Earth's water, it represents a major portion (30.1%) of total freshwater on Earth (Gleick, 1996). Due to growing demand for water, the extraction of groundwater has exceeded the rechargeable water across most regions of the globe resulting into steady lowering of the water table. The continuous lowering of water table would ultimately exhaust the groundwater resources, which would have serious impact on the food security and the local economy. For sustainable groundwater resource management, the quantification of groundwater recharge is significant and prerequisite.

Applications of Stable Isotopes of Hydrogen and Oxygen

Environmental hydrogen and oxygen isotopes are ideal tracers of water, as they are incorporated in the water molecules and their behaviours and variations reflect the origin of, and the hydrological and geochemical processes undergone by, natural water bodies (Gonfiantin, 2001). Being an integral part of water molecule and hydrological cycle, meteoric water in a particular environment exhibits a characteristic isotopic signature, which can be used as a tracer to understand atmospheric circulation patterns (Epstein and Mayeda, 1953; Gedzelman and Lawrence, 1982; Fritz et al., 1987; Jouzel et al., 1991; Edwards et al., 1996; Hoffman et al., 2000; Birks et al., 2002), to identify the recharge areas of groundwater (Gat, 1971; Clark and Fritz, 1997; Lee et al., 1999; Price and Swart, 2006), to quantify infiltration rates (McConville et al., 2001), to identify primary recharge seasons (Abbott et al., 2000; O'Driscoll et al., 2005), to identify the pathways of groundwater (Blasch and Bryson, 2007; Davisson et al., 1999; James et al., 2000; Jeelani et al., 2010; Lee et al., 1999; Mathieu and Bariac, 1996; Price and Swart, 2006; Stimson et al., 1993) and to estimate the mean residence time of water in catchments (e.g. McGuire et al., 2005; Ogrinc et al., 2008; Rodgers et al., 2005). Palaeoclimatic and palaeohydrological studies also rely often on precipitation data, and use speleothem carbonates (Spotl and Mangini, 2002), lake sediment organic material (Wolfe et al., 2001) or trees (McCarroll and Loader, 2004) as tools to reconstruct the events in the past. To understand the processes leading to isotopic variations in precipitation, it is necessary to first determine how the isotopic composition of modern precipitation is linked to today's climatic conditions (Jacob and Sonntag, 1991; Kohn and Welker, 2005).

Location and Climate of the Study Area

In the present study, stable water isotopes ($\delta^{18}O$ and δD) of precipitation, stream water and spring water were used to identify the sources of groundwater in a mountainous catchment of the western Himalaya (Figure 1). The region experiences a temperate climate

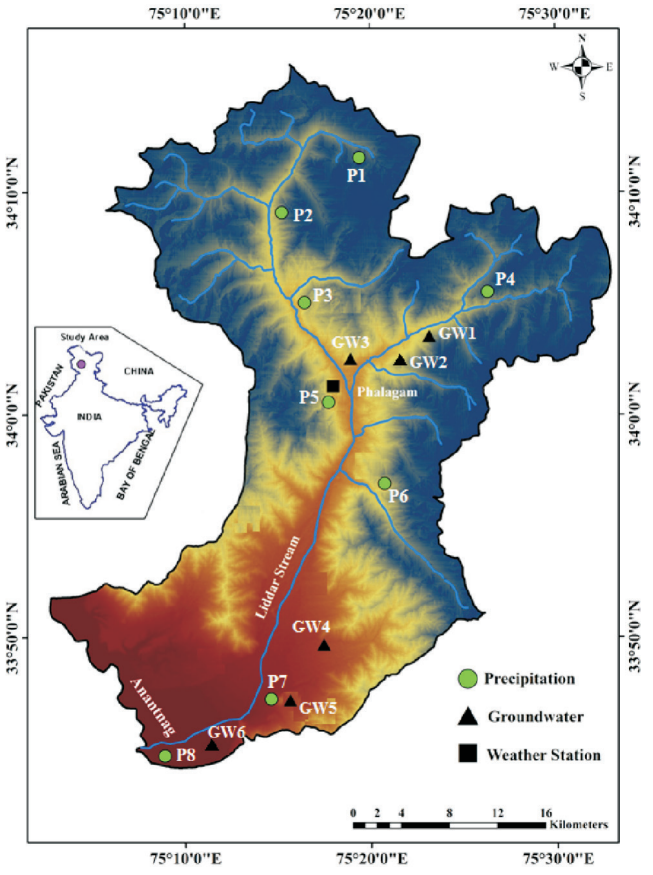

Figure 1: Location map of the study area showing the sampling sites.

characterised by cold winters, warm summers and clear seasonality, with average annual precipitation of about 1200 mm (the average value represents two decades from 1990 to 2009). The temperature and precipitation data used in the present investigation is from the Pahalgam Meteorological Station. There is great variability of temperature, which ranges from 37°C in summer to –15°C in winter. March normally receives the maximum rainfall of the year (183 mm) and November the least (36 mm). The precipitation is dominantly in the form of snow in winter and rain in rest of the seasons.

Geology and Hydrogeology of the Area

The Upper Paleozoic and Triassic rocks dominate the geology of the catchment. Triassic Limestone is surrounded by Palaeozoics and are overlain by Pleistocene (Karewas) and Recent sediments (Middlemiss, 1910, 1911; Wadia, 1975). Palaeozoic rocks include sandstone, silt stone, mudstone and shale (Lower Paleozoic), quartzites (Muth quartzites), Grey limestone, quartzite and shale (Syringothris Limestone), quartzite and shale (Fenestella Shale), pyroclastic slates, conglomerate, quartzite and porphyry granites (Aglomeratic Slate), andesitic and basaltic lavas with amygdaloidal and glomeroporphyritic texture (Panjal Traps), cherts, calcareous and siliceous shales, limestones and quartzites (Gangamopteris Beds) which occur mostly towards the marginal areas. Triassic Limestone, which consists of compact blue limestone, argillaceous limestone and dolomitic limestone intercalated with sand stone and shale, overlies the Paleozoic rocks (Middlemiss, 1910) and occur in the form of dissected ridges. The Karewa deposits of Pleistocene consist mostly of unconsolidated sandstones, beds of loess, conglomerates, etc. Small valleys between Triassic Limestone ridges and Karewa are filled with Recent Alluvium.

In the study area groundwater occurs in alluvium and carbonate aquifers. The Triassic Limestone is well karstified with the development of solution features including caves, caverns, conduits sink holes and sinking streams and is thus an important source of drinking water for local population. Number of karst springs emanate at the contact between the Triassic Limestone and Alluvium (Jeelani, 2008). The discharge of the cold springs is highly variable and fluctuating which ranges from 50 L/s in winter to about 5,000 L/s in summer.

Methodology

Precipitation and groundwater samples were collected across the mountainous watershed of Western Himalaya (Figure 1). The samples were collected between November 2010 and January 2012 on monthly basis. The location of the precipitation and groundwater sites along with their coordinates are presented as Table 1. The precipitation samples were generally collected as rain, except in winter when the samples were collected as snow. Rainwater samples were collected using a standard rain gauge with a long narrow tube attached to the plastic container, fitted with a funnel, to minimize the evaporative loss of stored rainwater. Groundwater samples were collected from springs and wells.

The water samples were analyzed for $\delta^{18}O$ and δD at Physical Research Laboratory (PRL) Ahmedabad using an isotope ratio mass spectrometer equipped with an equilibration unit and following the gas equilibration method (Epstein and Mayda, 1953). Secondary standards used in the batch were pre-calibrated using the primary standards and pre-analysed samples procured from the International Atomic Energy Agency, Vienna (IAEA/WMO, 1999). The $\delta^{18}O$ and δD values, in per mil (‰) are reported relative to Vienna Standard Mean Ocean Water (V-SMOW). In order to check the consistency in measurements, few samples were analyzed in duplicate in each cycle of measurements. The precision of the measurements was within ± 0.1‰ for $\delta^{18}O$ and ±1.0‰ for δD.

Stable Isotopes of Oxygen and Hydrogen in Precipitation

Spatial and temporal variations of stable isotope ratios of hydrogen ($^2H/^1H$) and oxygen ($^{18}O/^{16}O$) in precipitation are caused by equilibrium and kinetic isotopic fractionation mechanisms associated with condensation and evaporation processes during global water vapour circulation. The isotopic signature of precipitation is modified by a multitude of processes, like temperature and continual loss of moisture during the travel of airmass, i.e., Rayleigh distillation (Dansgaard, 1964; Rozanski et al., 1993; Kendall and McDonnell, 1998), mixing of air masses from local vapour sources and storm trajectory (Friedman et al., 1992; Clark and Fritz, 1997), altitude, amount, and continental effect (Dansgaard, 1964; IAEA, 1967; Rozanski et al., 1993, Clark and Fritz, 1997; Kendall and McDonnell, 1998; Arguas-Araguas, 2000; Fleitmann et al., 2004; Mook,

2000; Lachniet, 2009; Pape et al., 2010; Gat, 2010). Dansgaard (1964) first proposed a series of relationships between stable isotope ratios of hydrogen and oxygen in precipitation with several parameters, such as surface air temperature, latitude, altitude, amount of precipitation and distance from the coast. The relationships between the isotopic compositions of hydrogen and oxygen in precipitation with latitude (Lorius and Merlivat, 1977; Rozanski et al., 1982; Fisher, 1990), altitude (Taylor, 1972; Ehhalt, 1974; Rozanski and Sonntag, 1982; Jeelani et al., 2010, 2013), continentality (Salati et al., 1979; Sonntag et al., 1983; Ingraham and Taylor, 1991; Friedman et al., 1992), seasonality (Nativ and Riggio, 1990, Smith et al., 1992), precipitation amount (Matsuo and Friedman, 1967; Yapp, 1982; Lee and Fung, 2008; Aggarwal et al., 2004) and sources of distinct air masses (Gedzelman and Lawrence, 1982; Lawrence et al., 1982) is well documented for last few decades.

In the mountainous catchment of western Himalayas, the $\delta^{18}O$ and δD isotopic values in precipitation samples varied from 0.1‰ to –13.0‰ and 1‰ to –74‰ with an average of –6‰ and –38‰, respectively (Table 1). The lighter isotopic values (depleted in heavier isotopes) in precipitation were found in the samples collected from higher elevations while as the heavier isotopic values (enriched in heavier isotopes) were found in precipitation samples collected at low altitude stations. The depleted isotopic values observed at the higher altitudes and an enriched value towards the plain areas is consistent with the well-known altitude effect. The isotopic values of precipitation showed a good correlation with the altitude of the precipitation sites with $R^2 = 0.88$ and 0.90 for $\delta^{18}O$ and δD, respectively (Figure 2). The altitude gradient of 0.15‰ and 1.16‰ per 100 m changes in elevation amsl was observed for $\delta^{18}O$ and δD, respectively.

A good correlation was observed between temperature and stable isotopes of oxygen and hydrogen in precipitation (Figure 3) similar to that of Dansgaard (1964) with depleted isotopic values during low temperature and enriched values during higher temperatures. January recorded the most depleted value and July recorded the enriched isotopic value during the monitoring period. The isotopic signatures of precipitation showed negative correlation with the relative humidity in the study region. The isotopic values of precipitation are observed to decrease with increasing monthly or annual mean precipitation in many tropical and monsoon climate regions (Dansgaard, 1964; Rozanski et al., 1993) and with

Table 1: Location of the precipitation and groundwater sites along with their coordinates and isotopic values. The location ID is as per Figure 1

Precipitation			*$\delta^{18}O$(‰)*				*$\delta^{2}D$ (‰)*			
Location	*Lat. / Long.*	*Altitude (m, asl)*	*Min*	*Max*	*Mean*	*W.M*	*Min*	*Max*	*Mean*	*W.M*
P1	34° 12′ / 75° 17′	3487	-13	-3.1	-9	-5.4	-74	-18	-57	-32
P2	34° 09′ / 75° 14′	2787	-11.9	-1.4	-8.7	-4.5	-74	-11	-53	-30
P3	34° 05′ / 75° 15′	2400	-10.9	-1.6	-6.2	-3.1	-71	-16	-41	-23
P4	34° 04′ / 75° 22′	2472	-10.5	-1.4	-8.7	-4	-70	-9	-42	-24
P5	34° 01′ / 75° 18′	2200	-11.1	-1.6	-5.9	-3.3	-54	-17	-35	-20
P6	33° 49′ / 75°18′	2019	-9.6	0.1	-4.8	-3	-61	8	-29	-16
P7	33° 45′ / 75° 12′	1680	-8.7	-1	-4	-2	-51	1	-21	-10
P8	33° 43′ / 75° 09′	1616	-12.1	-5.6	-5.8	-3.2	-70	-26	-39	-21
Groundwater			*$\delta^{18}O$(‰)*				*$\delta^{2}D$ (‰)*			
Location	*Lat. / Long.*	*Altitude (m, asl)*	*Min*	*Max*	*Mean*	*WM*	*Min*	*Max*	*Mean*	*WM*
GW1	34° 04′ / 75° 30′	2200	-8.6	-7.8	-8.2	-4.1	-52.3	-42.0	-47.1	-28
GW2	34° 01′ / 75° 30′	1900	-8.7	-6.9	-7.9	5	-51.5	-38	-46.3	-29.6
GW3	34° 01′ / 75° 18′	2118	-9.0	-7.0	-8.5	5.3	-47.7	-41.7	-44.7	-25.4
GW4	34° 02′ / 75° 24′	2252	-8.7	-8.3	-8.4	-4.8	-48.0	-48.0	-48.0	-28.5
GW5	33° 45′ / 75°12′	1680	-10	-7.4	-8.5	-5.4	-58	-43.1	-50	-32.8
GW6	34° 43′ / 75° 10′	1608	-9.3	-6.8	-8	-5.8	-55.7	-38.4	-46	-33.5

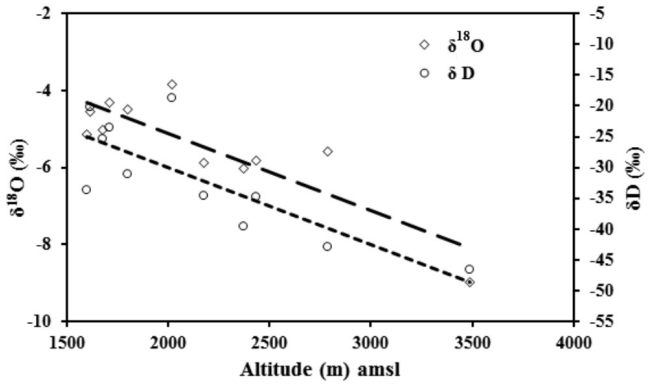

Figure 2: Relationship of stable isotopes with altitude.

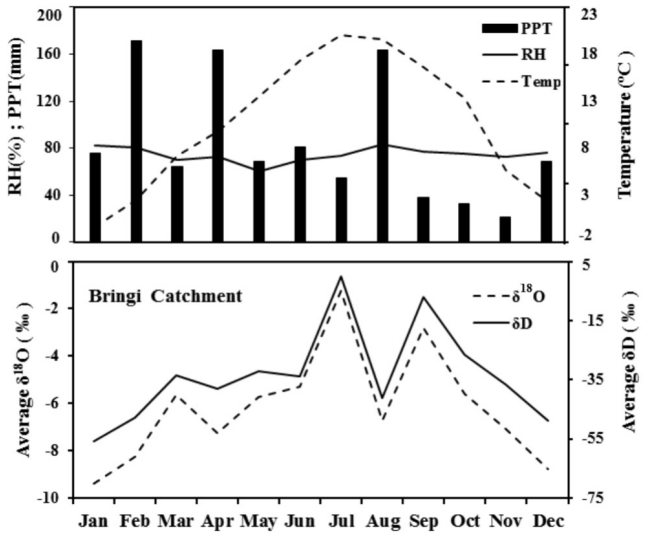

Figure 3: Monthly variation of stable isotopes in precipitation and their relationship with temperature, precipitation amount and relative humidity.

higher precipitation intensity in a single storm (Miyake et al., 1968). This relationship, widely known as amount effect, is also observed in the study area during some months: February, April and August (Figure 3). The temporal variability of isotopic values in precipitation is largely attributed to the prevailing airmass in the area.

Local Meteoric Water Line (LMWL)

The global relationship between δD and $\delta^{18}O$ in natural meteoric waters was recognized by Craig (1961) δD = 8 × $\delta^{18}O$ + 10 widely known as Global Meteoric Water Line (GMWL), serves as a foundational reference to determine regional or local deviations (i.e. local meteoric water line, LMWL) from equilibrium processes. This regression equation was subsequently redefined by Rozanski et al. (1993) based on monthly average values from 379 stations belonging to the Global Network of Isotopes in Precipitation (GNIP) maintained by the International Atomic Energy Agency (IAEA) and the World Meteorological Organization (WMO) as: δD = 8.1 (± 0.07) × $\delta^{18}O$ + 11.27 (± 0.65). The slope of 8.1 for the GMWL depends primarily on the ratio of the equilibrium fractionation factors for $\delta^{18}O$ and δD, while as the intercept of 11.27 is strongly governed by the relative humidity through non-equilibrium evaporation process involved in the formation of water vapour. The line is global in application, and is the average of many regional or local meteoric water lines (LMWL). The regression equations or meteoric water lines vary from place to place depending upon the prevailing weather conditions.

Depending on the region, differences in the amount of precipitation, temperature variations, distinct air mass sources, evaporation and fractionation processes occurring below the cloud base are characteristic at a local scale, which cause the relationship between stable isotopes of water δD and $\delta^{18}O$ to vary from that of the Global Meteoric Water Line (Hoefs, 2004). The LMWL for the study area (Figure 4) based on monthly average values is: δD = 7.1 (± 0.4) × $\delta^{18}O$ + 9.8 (± 1.4); R^2 = 0.95 (Jeelani et al. 2013). The LMWL shows a slightly lower slope and intercept than the GMWL and therefore suggest the effect of evaporation. The effect of evaporation varies with change in weather/seasons. In winter season most of the samples fall above the GMWL, while in summer season most of the samples lie below the regression line reflecting the effect of evaporation of falling raindrops due to increase in ambient temperature. The samples collected in winter season represent the effect of enhanced moisture recycling (Froehlic et al., 2002; Gat, 2010).

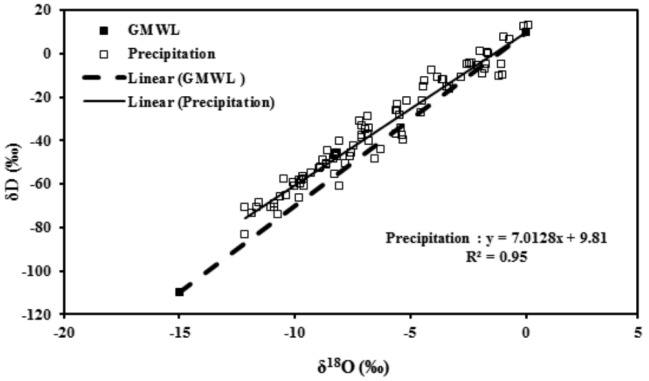

Figure 4: Relationship of $\delta^{18}O$ with δD in precipitation and groundwater.

Variation of Stable Isotopes of Oxygen and Hydrogen in Groundwater

The main causes of variations in the stable-isotope signature of groundwater are natural variations in the isotopic composition of rainfall, mixing with pre-existing waters and evaporation during percolation through soil and/or the unsaturated zone (Kendall and McDonnell, 1998). In comparison to precipitation, stable isotope composition of groundwater is characterised by relatively a narrow range and varied from –6.8 ‰ to –10 ‰ and for $\delta^{18}O$ and δD from 58‰ to –38‰ (Figure 5; Table 1).

The most depleted values were observed in karst springs and most enriched values were observed in shallow groundwater samples. The most depleted isotopic values in karst springs reflect the recharge at higher altitudes and no effect of evaporation, whereas the enriched isotopic values in shallow groundwater is attributed to the local recharge at lower elevation and the effect of evaporation. The groundwater in alluvium showed a narrow range of precipitation compared to the karst springs reflecting the attenuation of isotopic signature in alluvium and the quick flushing nature of the groundwater in karst. Similarly, the concomitant increase and decrease of $\delta^{18}O$ and $\delta^{2}H$ in groundwater with respect to time was also observed (Figure 5). The highest depleted values were observed in late spring and summer seasons whereas the enriched isotopic values were observed in March/July (Figure 5). This implies that groundwater system is not mixed enough to mute out the short-term and seasonal signals (Figure 5).

The abrupt and seasonal variation in groundwater recharge primarily seems to be responsible for the undulating behaviour in stable water isotope of groundwater. The decreasing trend of water isotopes from April to June suggests that the groundwater recharge is dominantly contributed from the melting of snow in spring and summer season. The recharge water released from snowpacks/glaciers is depleted in stable water isotopes. The enriched values in March suggest the contribution from snow melting at lower altitudes whereas the enriched isotopic values in groundwater in July suggest the recharge from rainfall with enriched isotopic values. Throughout the melting period, isotopically enriched rain events generate relatively higher stable water isotope ratios particularly in the summer.

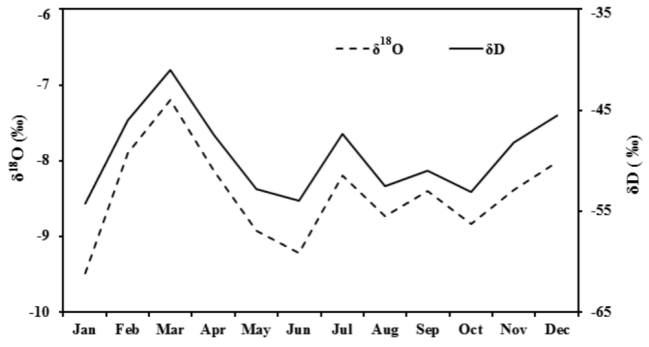

Figure 5: Monthly variation of stable isotopes in groundwater.

Most of the groundwater samples fall above or close to the GMWL indicating the minimal effect of evaporation (Figure 4). However, the clubbing of most groundwater samples in between the GMWL and LMWL indicate the common origin of groundwater system in the region fed by modern recharge source. By identifying the spatial distribution of isotopic signatures in groundwater at different locations in the studied region, we observed that groundwater exhibits mostly negative isotopic values, which infers groundwater recharge areas mostly at higher altitudes.

Recharge Areas

The recharge areas of groundwater also referred as capture zones are critically very important for the protection of groundwater resources from contamination (Abott et al., 2000) especially in karst terrains where the aquifers are susceptible to a greater range of environmental impact problems because of high permeability in the enlarged fissures and conduits and lack of effective attenuation mechanism (Ford and Williams, 1989). There are a number of approaches used to identify the capture zone or the recharge areas of groundwater or springs including hydrogeochemistry, tracer tests, stable isotopes, modelling, etc. However, the use of environmental isotopes in conjunction with hydrogeology has proved to be an effective tool in solving many hydrological problems (Clark and Fritz, 1997; Kendal and McDonell, 1998; Rao, 2006) including its use to identify the capture zones/recharge areas of groundwater/spring water (Shivanna et al., 2008; Yehdegho and Reichi, 2002; Aquilina et al., 2005). The spatio-temporal variation in the $\delta^{18}O$ and δD signature of precipitation is used to provide the variable input functions that are effective in tracing groundwater.

The $\delta^{18}O$ and δD value of groundwater commonly reflects the elevation and temperature of precipitation that infiltrates into the subsurface. With the help of the

local vertical isotopic gradient of precipitation obtained from the weighted mean isotopic data and elevation data of precipitation, the mean elevation of precipitation where the recharge occurs may be estimated (Clark and Fritz, 1997; Longinelli and Selmo, 2003; Yurtsever and Gat, 1981). The mean vertical isotopic gradient in the study area was estimated as –0.15 and –1.16‰ per 100 m increase in elevation, for $\delta^{18}O$ and δD, respectively. Since most of the groundwater samples show mean oxygen and hydrogen isotope ratios ranging from –4‰ to –6‰ and –25‰ to –35‰, respectively, the altitude of the recharge areas ranging from 2800 to 3800 m amsl (based on $\delta^{18}O$) and 3100 to 3800 m amsl (based on δD) was determined (Figure 6). The averages estimated recharge elevation of an individual karst spring, with unique isotopic signature, could be estimated by the following regression equations:

1. Using $\delta^{18}O$ tracer: $\delta^{18}O = -0.001\ (\pm 0.00027) \times$ altitude $- 0.6\ (\pm 0.48)$
2. Using δD tracer: $\delta D = -0.01\ (\pm 0.0018) \times$ altitude $+ 3.9\ (\pm 4.0)$

The inferred recharge altitude estimated by these tracers could also be overestimated or underestimated in areas where the depleted seasonal snowmelt at higher altitudes is brought down by the stream flow to lower altitudes (recharge sites) without much fractionation and in areas where effect of evaporation is prevalent on groundwater.

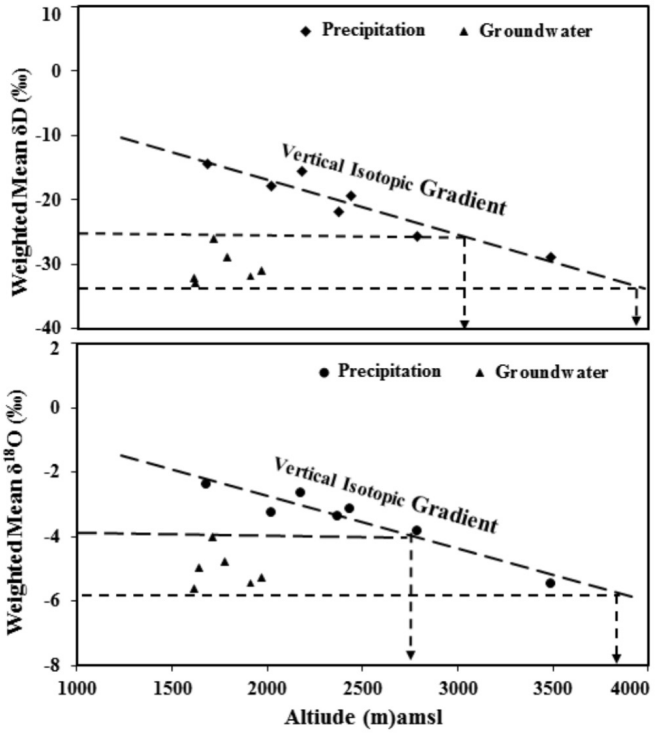

Figure 6: Weighted mean $\delta^{18}O$ (lower panel) and δD (upper panel) vs. altitude of the samples.

Impact of Climate Change on Spring Flow

Temperature and precipitation records available for the western Himalayas indicate an overall increase of temperature and precipitation. The long-term data (1901-2010) from Srinagar weather station indicate a warming of about 0.9º C (Jeelani et al., 2012). An increase in precipitation of about <3% was also observed, with a significant variation in short-term trends over the entire time period. The average seasonal temperatures reflected a little warming in summer (0.001 °C yr^{-1}) and strong warming in winter season (0.02 °C yr^{-1}), with a 0.15 mm yr^{-1} decreasing trend of winter precipitation (Jeelani et al., 2012).

As most of the recharge areas lie at higher altitudes in the western Himalayan catchments, any change in snow and ice cover would substantially affect the groundwater resources.

As groundwater plays an important role in the lives of the people across the globe, in the western Himalayas, where regional economy is chiefly dependent on snow and glaciers, the transition from snow to rain in a warming climate are expected to significantly impact the cryospheric processes and hydrology of the region (Jeelani et al., 2012, 2017b). The continuous increase in atmosphere temperature will lead to a continuous shift of snow line toward higher altitude (Jeelani et al., 2012). This means that higher-altitude areas will receive more liquid precipitation and less solid precipitation, which could increase the episodic discharge of the karst springs. The episodic recharge/discharge is not good for the sustainability of the springs. This would not only lead to the degradation of the water quality of the springs but also the significant attenuation of the spring flow. The shorter recharge season due to shorter persistence and significant reduction in snow/ice cover under changing climate will thereby strongly affect the hydrological behaviour in the region by causing uncertain trends in the recharge and discharge of the springs.

The climate change has already shown its presence in the region as lesser amounts of snowfall in preceding winters, the annual discharge of perennial karst springs drastically reduced by 40-70%, besides, seasonal springs dried up in summer in 1999 and 2001 (Jeelani, 2008).

Similarly, in 2011 and 2013 due to the below normal precipitation (24% in 2012 and 36% in 2013), 4-18% reduction in annual discharge of some karst springs in the region was recorded. However, 28% above normal precipitation in 2014 promoted higher flows in the same springs. 1259, 1084 and 1711 mm of precipitation was recorded in 2011, 2013 and 2014, respectively. As hydrological cycle is intimately linked to the climate system, dependence of groundwater flow on snow and glaciers thus suggest that changes in timing and magnitude of solid precipitation would substantially affect the groundwater resources.

Conclusions

Characterization of the stable isotopes of oxygen and hydrogen in precipitation and groundwater has helped to identify the recharge areas of groundwater/springs in a mountainous catchment of western Himalaya. A marked spatial and seasonal variability of stable isotopes of oxygen and hydrogen was observed in precipitation. The lighter isotopic values were observed in precipitation samples at higher altitudes and the heavy isotopic values were observed at lower elevations. The $\delta^{18}O$ and δD in groundwater showed a narrow spatial and temporal variation in comparison to precipitation. Based on the estimated mean altitude gradient of –0.15‰ and –1.16‰ per 100 m change in elevation for $\delta^{18}O$ and δD, respectively, the altitude of the recharge areas of groundwater were identified, which ranges from 2800 to 3800 m amsl (based on $\delta^{18}O$) and 3100 to 3800 m amsl (based on δD). The results suggest that the groundwater with most depleted isotopic values have recharge areas at higher altitudes, whereas the groundwater with less depleted isotopic values has recharge areas at lower elevations. The study revealed that the groundwater discharge/spring flow is expected to be substantially affected by the climate change.

References

Abbott, M.D., Lini, A. and Bierman, P.R., 2000. $\delta^{18}O$, $\delta^{2}H$ and ^{3}H measurements constrain groundwater recharge patterns in an upland fractured bedrock aquifer, Vermont, USA. *Journal of Hydrology*, **228:** 101-112.

Aquilina, L., Ladouche, B. and Dorfligre, N. 2005. Recharge processes in karstic systems investigated through the correlation of chemical and isotopic composition of rain and spring-waters. *Applied Geochemistry*, **20(12):** 2189-2206.

Aggarwal, P., Fröhlich, K., Kulkarni, K. and Gourcy, L., 2004. Stable isotope evidence for moisture sources in the Asian summer monsoon under present and past climate regimes. *Geophys. Res. Lett*., **31:** (L0823).

Araguas-Araguas, L., Froehlich, K. and Rozanski, K., 2000. Deuterium and oxygen-18 isotope composition of precipitation and atmospheric moisture. *Hydrol. Process.*, **14:** 1341-1355.

Baertschi, P., 1976. Absolute 18O content of Standard Mean Ocean Water. *Earth Planet Sc Letter*, **31:** 341-344.

Blasch, K.W. and Bryson, J R., 2007. Distinguishing sources of groundwater recharge by using δD and $\delta^{18}O$. *Ground Water*, **45:** 294-308.

Birks, S.J., Gibson, J.J., Gourcy, L., Aggarwal, P.K. and Edwards, T.W.D., 2002. Maps and animations offer new opportunities for studying the global water cycle. *EOS*, **83:** 406.

Clark, I. and Fritz, P., 1997. Environmental Isotopes in Hydrogeology. Lewis Publishers, New York.

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

Craig, H. and Gordon, L.I., 1965. Deuterium and oxygen-18 variations in the ocean and marine atmosphere. *In:* Stable isotopes in oceanographic studies and paleotemperatures. Laboratory of Geology and Nuclear Science, Pisa, Italy.

Coplen, T.B., 1996. New guidelines for the reporting of stable hydrogen, carbon, and oxygen isotope ratio data. *Geochimica et Cosmochimica Acta*, **60:** 3359.

Dansgaard, W., 1964. Stable isotopes in precipitation. *Tellus*, **16:** 436-468.

De Wit, J.C., Van der Straaten, C.M. and Mook, W.G., 1980. Determination of the absolute hydrogen isotope ratio of V-SMOW and SLAP. *Geostandards Newsletter*, **4:** 33-36.

Davisson, M.L., Smith, D.K., Kenneally, J. and Rose, T.P., 1999. Isotope hydrology of southern Nevada groundwater: Stable isotopes and radiocarbon. *Water Resource Research*, **35:** 279-294.

Edwards, T.W.D., Wolfe, B.B. and MacDonald, G.M., 1996. Influence of changing atmospheric circulation on precipitation $\delta^{18}O$-temperature relations in Canada during the Holocene. *Quaternary Research*, **46:** 211-218.

Ehhalt, D.H., 1974. Vertical profiles of HTO, HDO and H_2O in the troposphere. NCAR Technical Note, NCAR-TN/STR-100, NCAR, Boulder, CO, 133 pp.

Epstein, S. and Mayeda, T., 1953. Variations of ^{18}O content of waters from natural sources. *Geochim. Cosmochim. Acta*, **4:** 213-224.

Froehlich, K., Gibson, J.J. and Aggarwal, P.K., 2002. Deuterium excess in precipitation and its climatological significance. *In:* Study of Environmental Change Using Isotope Techniques. IAEA, Vienna.

Fleitmann, D., Burns, S.J., Neff, U., Mudelsee, M., Mangini, A. and Matter, A., 2004. Palaeoclimatic interpretation of high-resolution oxygen isotope profiles derived from annually laminated speleothems from Southern Oman. *Quaternary Science Review*, **23:** 935-945.

Friedman, I., Smith, G.I., Gleason, J.D., Warden, A. and Harris, J.M., 1992. Stable isotope composition of waters in southeastern California 1. Modern precipitation. *Journal of Geophysical Research*, **97:** 5795-5812.

Fisher, D.A., 1990. A zonally-averaged stable-isotope model coupled to a regional variable elevation stable-isotope model. *Ann. Glaciol*., **14:** 65-71.

Ford, D.C. and Williams, P.W., 1989. Karst geomorphology and hydrology (vol. 601). London: Unwin Hyman.

Friedman, I., Smith, G.I., Gleason, J.D., Warden, A. and Harris, J.M., 1992. Stable isotope composition of waters in southeastern California, Part 1. Modern precipitation. *J. Geophysical Research*, **97:** 5795-5812.

Fritz, P., Drimmie, R.J., Frape, S.K. and O'Shea, K., 1987. The isotopic composition of precipitation and groundwater in Canada. *In:* Isotope Techniques in Water Resources Development. International Atomic Energy Agency, Vienna.

Gedzelman, S.D. and Lawrence, J.R., 1982. The isotopic composition of cyclonic precipitation. *J. Appl. Meteorology*, **21:** 1387-1404.

Gat, J.R., 1971. Comments on the stable isotope method in regional groundwater investigations. *Water Resources Research*, **7:** 980-993.

Gat, J.R., 2010. Isotopes Hydrology: A case study of the water cycle. Vol. 6, Imperial College Press.

Gleick, P.H., 1996. Water resources. *In:* Encyclopedia of Climate and Weather. S.H. Schneider (ed.). Oxford University Press, New York, **2:** 817-823.

Gonfiantini, R., Roche, M.A., Olivry, J.C., Fontes, J.C. and Zuppi, G.M., 1998. The altitude effect on the isotopic composition of tropical rains. *Chem. Geol.*, **181:** 147-167.

Hoffman, G., Jouzel, J. and Masson, V., 2000. Stable water isotopes in atmospheric general circulation models. *Hydrological Processes*, **14(8):** 1385-1406.

Hoefs, J., 2004. Stable Isotope Geochemistry. Springer, Germany.

IAEA, 1967. Tritium and other environmental isotopes in the hydrological cycle. Vienna, IAEA-Technical Report, **73:** 1-83.

IAEA/WMO, 1999. Global network for isotopes in precipitation. The GNIP database. Release 3, October, 1999. http://www.naweb.iaea.org/napc/ih/HIS_resources_gnip.html.

Ingraham, N.L. and Taylor, B.E., 1991. Light stable isotope systematics of large-scale hydrologic regimes in California and Nevada. *Water Resource Research*, **27(1):** 77-90.

Jacob, H. and Sonntag, C., 1991. An 8-year record of seasonal variation of ^{2}H and ^{18}O in atmospheric water vapour and precipitation at Heidelberg, Germany. *Tellus*, **B43:** 291-300.

Jouzel, J., Koster, R.D., Suozzo, R.J., Russel, G.L., White, J.W.C. and Broecker, W.S., 1991. Simulations of the HDO and $2H^{18}O$ atmospheric cycles using the NASA GISS general circulation model: Sensitivity experiments for present-day conditions. *J. Geophysical Research*, **96:** 7495-7507.

Jeelani, G., 2008. Aquifer response to regional climate variability in a part of Kashmir Himalayas in India. *Hydrogeology Journal*, **16:** 1625-1633.

Jeelani, G., Bhat, N.A. and Shivanna, K., 2010. Use of ^{18}O tracer to identify stream and spring origins of a mountainous catchment; A case study from Liddar watershed, Western Himalaya, India. *J. Hydrol.*, **393:** 257-264.

Jeelani, G., Feddema, J.J., van der Veen, C.J. and Stearns, L., 2012. Role of snow and glacier melt in controlling river hydrology Liddar watershed in the (western Himalaya) under current and future climate. *Water Resource Research*, **48:** 1-16.

Jeelani, G., Kumar, U.S. and Kumar, B., 2013. Variation of $\delta^{18}O$ and δD in precipitation and stream waters across the Kashmir Himalaya (India) to distinguish and estimate the seasonal sources of stream flow. *Journal of Hydrology*, **481:** 157-165.

Jeelani, G., Shah, R.A., Jacob, N. and Deshpande, R.D., 2017. Estimation of snow and glacier melt contribution to Liddar stream in a mountainous basin, western Himalaya: An isotopic approach. *J. Isotop. Environ. Health Stud.*, **53:** 18-35.

James, E.R., Manga, M., Rose, T.P. and Hudson, G.B., 2000. The use of temperature and the isotopes of O, H, C, and noble gases to determine the pattern and spatial extent of groundwater flow. *J. Hydrol.*, **237:** 100-112.

Kendall, C. and Mc-Donnell, J.J., 1998. Isotope Tracers in Catchment Hydrology. (Eds.) Elsevier Science B.V., Amsterdam.

Kohn, M.J. and Welker, J.M., 2005. On the temperature correlation of $\delta^{18}O$ in modern precipitation. *Earth and Planetary Science Letters*, **231:** 87-96.

Lide, D.R. and Frederikse, H.P.R., 1995. Handbook of chemistry and physics. 76th edition CRC Press, Boca Raton Florida.

Lachniet, M.S. and Patterson, W.P., 2009. Oxygen isotope values of precipitation and surface waters in northern Central America (Belize and Guatemala) are dominated by temperature and amount effects. *Earth Planet. Sc. Lett.*, **284:** 435-446.

Lee, K.S., Wenner, D.B. and Lee, I., 1999. Using H- and O-isotopic data for estimating the relative contributions of rainy and dry season precipitation to groundwater: Example from Cheju Island, Korea. *J. Hydrol.*, **222:** 65-74.

Lee, J.E. and Fung, I., 2008. Amount effect of water isotopes

and quantitative analysis of post-condensation processes. *Hydrol. Process*., **22:** 1-8.

Longinelli, A. and Selmo, E., 2003. Isotopic composition of precipitation in Italy: A first overall map. *Journal of Hydrology*, **270:** 75-88.

Lawrence, J.R., Gedzelman, S.D., White, J.W.C., Smiley, D. and Lazov, P., 1982. Storm trajectories in eastern US: D/H isotopic composition of precipitation. *Nature*, **296:** 638-640.

Lorius, C. and Merlivat, L., 1977. Distribution of mean surface stable isotope values in East Antarctica: Observed changes with depth in the coastal area. In: Isotopes and Impurities in Snow and Ice, IAHS Publication, **118:** 127-137.

Matsuo, S. and Friedman, I., 1967. Deuterium content in fractionally collected rain. *J. Geophys. Res.*, **72:** 6374-6376.

McCarroll, D. and Loader, N.J., 2004. Stable isotopes in tree rings. *Quaternary Science Reviews*, **23:** 771-801.

McGuire, K.J., Mc Donnell, J.J., Weile, M., Kendall, C., Mc Glynn, B.L., Welker, J.M. and Seibert, J., 2005. The role of topography on catchment-scale water residence time. *Water Resource Research*, **41:** 1-14.

McConville, C., Kalin, R.M., Johnston, H. and McNeil, G.W., 2001. Evaluation of recharge in a small temperate catchment using natural and applied $\delta^{18}O$ profiles in the unsaturated zone. *Ground Water*, **39(4):** 616-624.

Middlemiss, C.S., 1910. Revision of Silurian-Trias sequence of Kashmir. *Record Geological Survey of India*, **40:** 206-260.

Middlemiss, C.S., 1911. Sections in the Pir Panjal range and Sindh Valley, Kashmir. *Record Geological Survey of India*, **1:** 115-144.

Miyake, Y., Matsubaya, O. and Nishihara, C., 1968. An isotopic study on meteoric precipitation. Papers in Meteorology and Geophysics, **19:** 243-266.

Mook, W.G., 2000. Environmental isotopes in the hydrological cycle. *In:* Vol. I Introduction. Mook, W.G. (ed). International Atomic Energy Agency, Groningen.

Nativ, R. and Riggio, R., 1990. Precipitation in the southern high plains: Meteorological and isotopic features. *J. Geophys. Res.*, **95(13):** 22559-22564.

O'Driscoll, M.A., DeWalle, D.R., McGuire, K.J. and Gburek, W.J., 2005. Seasonal $\delta^{18}O$ variations and groundwater recharge for the three landscapes types in the Central Pennsylvania, USA. *Journal of Hydrology*, **303:** 108-124.

Ogrinc, N., Kanduc, T., Stichler, W. and Vrec, P., 2008. Spatial and seasonal variations in $\delta^{18}O$ and δD values in the River Sava in Slovenia. *J. Hydrol.*, **359:** 303-312.

Pape, J., Banner, J., Mack, L., Musgrove, M. and Guilfoyle, A., 2010. Controls on oxygen isotope variability in precipitation and cave drip waters, central Texas, USA. *J. Hydrol.*, **385:** 203-215.

Price, R.M. and Swart, P.K., 2006. Geochemical indicators of groundwater recharge in the Surficial Aquifer System, Everglades National Park, Florida, USA. *In:* Harmon, R.S. and Wicks, C. (Eds), Perspectives on Karst Geomorphology, Hydrology, and Geochemistry. A Tribute Volume to Derek C. Ford and William B. White. Geological Society of America.

Rozanski, K. and Sonntag, C., 1982. Vertical distribution of deuterium in atmospheric water vapor. *Tellus*, **34:** 135-141.

Rozanski, K., Aruguas-Aruguas, L. and Ganfiantini, R., 1993. Isotopic patterns in modern global precipitation. *Geophys. Monogr.*, **78:** 1-36.

Rozanski, K., Sonntag, C. and Munnich, K.O., 1982. Factors controlling stable isotope composition of European precipitation. *Tellus*, **34:** 142-150.

Rao, S.M., 2006. Practical Isotope Hydrology. New India Publishing Agency, New Delhi.

Shivanna, K., Tirumalesh, K., Noble, J., Joseph, T.B., Singh, G., Joshi, A.P. and Khati, V.S., 2008. Isotope techniques to identify recharge areas of springs for rainwater harvesting in the mountainous region of Gaucher area, Chamoli District, Uttarakhand. *Current Science*, **94:** 1003-1011.

Stimson, J., Rudolph, D., Farvolden, F., Frape, S. and Drimmie, R., 1993. Causes of groundwater salinization in a low lying area of Cochabamba Valley, Bolivia. *In:* Isotope Techniques in the Study of Past and Current Environmental Changes in the Hydrosphere and the Atmosphere. IAEA-SM-**329(26):** 185-198.

Spotl, C. and Mangini, A., 2002. Stalagmite from the Austrian Alps reveals Dansgaard-Oeschger events during isotope stage 3: Implications for the absolute chronology of Greenland ice cores. *Earth and Planetary Science Letters*, **203:** 507-518.

Salati, E., Dall'Olio, A., Matsui, E. and Gat, J.R., 1979. Recycling of water in the Amazon basin: An isotopic study. *Water Resource Research*, **515(5):** 1250-1258.

Smith, G.I., Friedman, I., Gleason, J.D. and Warden, A., 1992. Stable isotope composition of waters in southeastern California: Ground waters and their relation to modern precipitation. *J. Geophs. Res.*, **97(5):** 5813-5823.

Sonntag, C., Rozanski, K., Munnich, K.O. and Jacob, H., 1983. Variations of deuterium and oxygen-18 in continental precipitation and groundwater and their causes. *In:* Variations in the Global Water Budget, (eds) Reidel Dordrecht.

Taylor, C.B., 1972. The vertical variations of the isotopic concentrations of tropospheric water vapor over continental Europe and their relationship to tropospheric structure. *NZ Inst. Nucl Sci Rep*, INS-R-**107:** 44.

Yapp, C.J., 1982. A model for the relationships between precipitation D/H ratios and precipitation intensity. *J. Geoph Res*, **87:** 9614-9620.

Yurtsever, Y. and Gat, J.R., 1981. Atmospheric waters. *In:*

Stable Isotope Hydrology: Deuterium and Oxygen-18 in the Water Cycle. Gat, J.R. and Gonfiantini, R. (eds). 210 IAEA: Vienna; 103-142.

Yehdegho, B. and Reichl, P., 2002. Recharge areas and hydrochemistry of carbonate springs issuing from the Semmering massif, Austria, based on long-term oxygen-18 and hydrochemical data evidences. *Hydrogeol Journal*, **10:** 628-642.

Wolfe, B.B., Edwards, T.W.D., Elgood, R.J. and Beuning, K.R.M., 2001. Carbon and oxygen isotope analysis of lake sediment cellulose: Methods and applications. *In:* Last, W.M. and Smol, J.P. (eds), Tracking Environmental Change Using Lake Sediments: Physical and Geochemical Methods. Kluwer Academic Publishers, Dordrecht. The Netherlands, **2:** 373-400.

Wadia, D.N., 1975. Geology of India. Tata McGraw Hill. New Delhi.