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Decoding the hydro-meteorological processes of spring recharge dynamics in the Alakananda River basin through stable and radioisotopes

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ABSTRACT

The Himalayan springs are crucial water sources in mountainous regions, which are drying at an accelerated rate. The current study analyses the spatial and temporal distributions of δD and $\delta^{18}O$ in precipitation, spring, and stream water, along with tritium (^3H), to investigate hydrological processes and understand spring recharge dynamics. Due to contribution from precipitation, spring samples collected from the monsoon period are more isotopically depleted ($\delta^{18}O$: $-8.81 \pm 0.55\text{‰}$; δD : $-60 \pm 4.15\text{‰}$) in comparison to samples taken in the pre-monsoon period ($\delta^{18}O$: $-8.10 \pm 0.42\text{‰}$; δD : $-55.7 \pm 3.07\text{‰}$) and winter ($\delta^{18}O$: $-8.44 \pm 0.43\text{‰}$; δD : $-57.79 \pm 2.43\text{‰}$), which suffer the isotopic fractionation subjected to evaporative loss. A mass balance approach estimates ~82% monsoon contribution to the spring water budget. Tritium analysis reflects contributions from both deep and shallow groundwater sources to the springs. Overall, the hydro-meteorological parameters, including overland flow during the monsoon season and groundwater baseflow conditions, play a crucial role in shaping the spring dynamics.

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1 Introduction

Spring water supports millions of people in the Himalayan region, playing a vital role in the regional water budget, and holds deep cultural significance for mountain communities. These springs serve as the primary water sources for mountainous rivers. Settlements, including villages and their smaller hamlets, are commonly established near springs, while larger towns are often situated along riverbanks. However, a continuous decline in spring and stream discharge, along with an intensified drying of many springs across the Himalayan range (Valdiya and Bartarya 1989, 1991, Kumar *et al.* 1997, Tambe *et al.* 2012, Qazi *et al.* 2017), has been observed in recent years, particularly during the summer months, forcing inhabitants to limit their water usage, which can occasionally give rise to social tensions (Negi and Joshi 1996). This highlights the urgent need for studies that explain the hydro-meteorological processes underlying spring recharge. Despite this, Himalayan spring research has received little attention, and its contribution to river basin hydrology across contrasting catchments remains insufficiently explored. Spring water serves over 40 million people in the Himalayas (Mahamuni and Kulkarni 2012), is crucial for the region's water budget (Bookhagen and Burbank 2010, Andermann *et al.* 2012), supports ecosystem biodiversity, and is deeply rooted in the cultural and spiritual beliefs of mountain communities (Tambe *et al.* 2012). To harness green energy, various hydropower plants were established, which typically rely on spring and stream discharge, especially during non-rainy seasons. Spring-fed, snow-fed, and glacier-fed rivers are vital for

hydropower generation in India and worldwide (Bhadra *et al.* 2015, Rautela *et al.* 2023). The consistent flow from these sources enhances hydropower potential, supports energy demand, and promotes sustainability. Approximately 50% of the Indian Himalayan springs are drying up (NITI 2018), and the reasons for this remain largely unknown. In the Himalayan terrain, high seasonal fluctuations in the groundwater table and surface water discharge, driven by prevailing hydrogeological and hydro-meteorological conditions, make spring water even more important for meeting local water demands and sustaining the area's ecology. Since this part of the Himalayas is an eco-sensitive zone, and the changing climate has impacted the water availability in this region, proper scientific investigations and monitoring are required to characterize the dynamics of springs in the Himalayan region.

The stable isotopes (δD and $\delta^{18}O$) in natural water serve as valuable tracers for understanding hydrological processes (Rozanski *et al.* 1993, Soderberg *et al.* 2013, Li *et al.* 2017, Pradhan *et al.* 2025, Yulianti *et al.* 2025). Recent efforts have focused on disentangling moisture sources using isotopic analysis; tracking recharge zones; improving understanding of water storage, altitude effects, evaporation conditions, and mixing processes; and determining the residence time of water in the watershed (through the tracers like δD , $\delta^{18}O$, ^3H , and deuterium excess, denoted as *d*-excess), which influence glaciers and springs (Shivanna *et al.* 2008, Jeelani *et al.* 2010, 2015, Kumar *et al.* 2021, Shah *et al.* 2023, Lyon *et al.* 2024, Holmes *et al.* 2025, Sharma *et al.* 2025). The isotopic

composition of $\delta^{18}\text{O}$ and δD in rainfall samples collected from various altitudes and latitudes along a hillslope shows distinct signatures, contributing to the development of a local meteoric water line (LMWL; Fontes *et al.* 1967, Gat 1971, Clark and Fritz 1997, Fetter 2001, Price and Swart 2006, Jeelani *et al.* 2010, Kumar *et al.* 2010, Chen *et al.* 2024, Quiroz Londoño *et al.* 2025). However, processes such as evaporation and the mixing of recharge water with groundwater from diverse sources can alter the isotopic signature of springs, causing deviations from the LMWL (O'Driscoll *et al.* 2005, Barry *et al.* 2024, Abudouwaili *et al.* 2025). Significant isotope-based studies on springs have been conducted across the Himalayas (Shivanna *et al.* 2008, Jeelani *et al.* 2010, 2015, Florea *et al.* 2017, Shah *et al.* 2017, 2023, Matheswaran *et al.* 2019, Tarafdar *et al.* 2019, Thakur *et al.* 2020). Tritium, a radioisotope of hydrogen, has been widely used to detect recharge events over the past 50 years (Clark and Fritz 1997). It serves as a tracer in Himalayan springs to determine groundwater residence time, recharge sources, and the dynamic flow behaviour of systems, especially in response to seasonal and altitudinal variations (Shivanna *et al.* 2008, Jeelani *et al.* 2015, Shah *et al.* 2017, Bhat and Jeelani 2018, Thakur *et al.* 2020, Chatterjee *et al.* 2023). The observed trend of declining spring discharge, a critical water resource across the Himalayas coupled with its critical role in river recharge, highlights the urgent need to identify moisture sources, assess the altitudinal gradients of recharge, and address water scarcity issues. This situation has prompted the Himalayan isotopic community to explore these research questions across contrasting valleys of the Himalayas.

In the current climate change context, it is important to have a deep understanding of spring recharge mechanisms and seasonal dynamics. Therefore, in this study, we have combined stable isotope analysis (δD and $\delta^{18}\text{O}$) of seasonal rainwater and spring water (pre-monsoon, monsoon, and winter) with tritium measurements to estimate spring water residence time in the Takoli Gad catchment, Garhwal Himalaya, Uttarakhand. The present study is unique in the Himalayan context, and the general objective of this study is to understand spring recharge processes and residence time using isotopic and hydro-meteorological proxies. The specific objectives of this study are:

- To present the isotopic signatures of precipitation and spring water across different seasons to elucidate the processes involved in spring recharge as well as to explore the interplay of various factors, including geology, slope, topography, soil moisture content, overland flow, and baseflow/groundwater runoff.
- To estimate the relative meteoric contribution during different seasons, which sustains the continuous flow of springs in the study area.
- To assess the utility of remote sensing-derived datasets, i.e. soil moisture, relative humidity, overland flow, and baseflow, on the spring dynamics and investigate their relationship with isotopic interpretations.

As the study involves a high-resolution spring and precipitation isotopic dataset from a relatively small catchment, we expect that these above-mentioned approaches will

provide deep, substantial insights into the hydro-meteorological processes influencing and sustaining spring recharge.

2 Study area and geological setting

The Takoli Gad catchment area is situated within the lesser Himalaya section extending between the geographical coordinates of $30^{\circ}14'N$ to $30^{\circ}23'N$ latitude and $78^{\circ}37'E$ to $78^{\circ}46'E$ longitude, covering an area of $\sim 143\text{ km}^2$ (Fig. 1(a,b)). The watershed is situated in the upper catchment of the Alaknanda River and has an elevation range from 505 m to 2496 m (Fig. 1(b)). The Takoli Gad stream originates from the eastern slope of Chandrabadni Peak (2278 m) and becomes a tributary of the Alaknanda River at Juyalgarh. The region experiences a monsoon climate, with mean annual rainfall ranging from 1829 to 1870 mm (Chhillar and Joshi 2022). The region receives the majority of its precipitation ($\sim 75\%$) during the monsoon season from July to September (see Supplementary material, Fig. S1), totalling an average annual precipitation of ~ 1395 mm. Temperature fluctuations in the watershed range from 4.6°C to 19.6°C in winter and between 32°C and 36.5°C in summer (Bagchi and Singh 2011). The monthly precipitation rate exhibits a pronounced seasonal cycle, with the monsoon season (June–September) experiencing the highest rainfall, particularly peaking in July and August. This is followed by the winter season, while the pre-monsoon period (March–May) shows relatively lower precipitation (see Supplementary material, Fig. S1). The relative humidity (RH%) and potential evaporation data follow an opposing seasonal trend. RH reaches its peak during the monsoon season, while the pre-monsoon period experiences the lowest values (Fig. 2(a)). Potential evaporation is highest in the pre-monsoon months, gradually declining as the monsoon sets in (Fig. 2(b)).

Geologically, the North Almora Thrust (NAT) divides the study area roughly into two distinct halves (Fig. 1(a, b)). The study area is mainly composed of the Garhwal and Jaunsar groups of rocks (Fig. 1(a)) (Kumar *et al.* 2026). The geology of the area is diverse, with significant formations including the Berinag Formation, which consists of basic and meta-volcanic rock; the Chandpur Formation, which contains phyllite, quartzite, shale, dolomite, and tuff with dolerite; and the Pithoragarh Formation, made up of limestone, shale, dolomite, and cherty quartzite. Additionally, the Nagnithank Formation is composed of quartzite and slate interbedded with basic meta-volcanic rock, and the Uttarkashi Formation contains quartzite and slate with lenticular deposits of limestone and tuff. A small fraction of the region is composed of the Rautgara Formation, containing eroded gravel, pebbles, silt, and clay (<https://bhukosh.gsi.gov.in>). Furthermore, the watershed comprises two distinct soil types: Dystric Eutrudepts and Typic Udorthents (Chhillar and Joshi 2022, Kumar *et al.* 2026) (National Bureau of Soil Survey and Land Use Planning, ICAR). Dystric Eutrudepts are relatively shallow, well-drained, thermic soils with a fine-loamy texture and slight

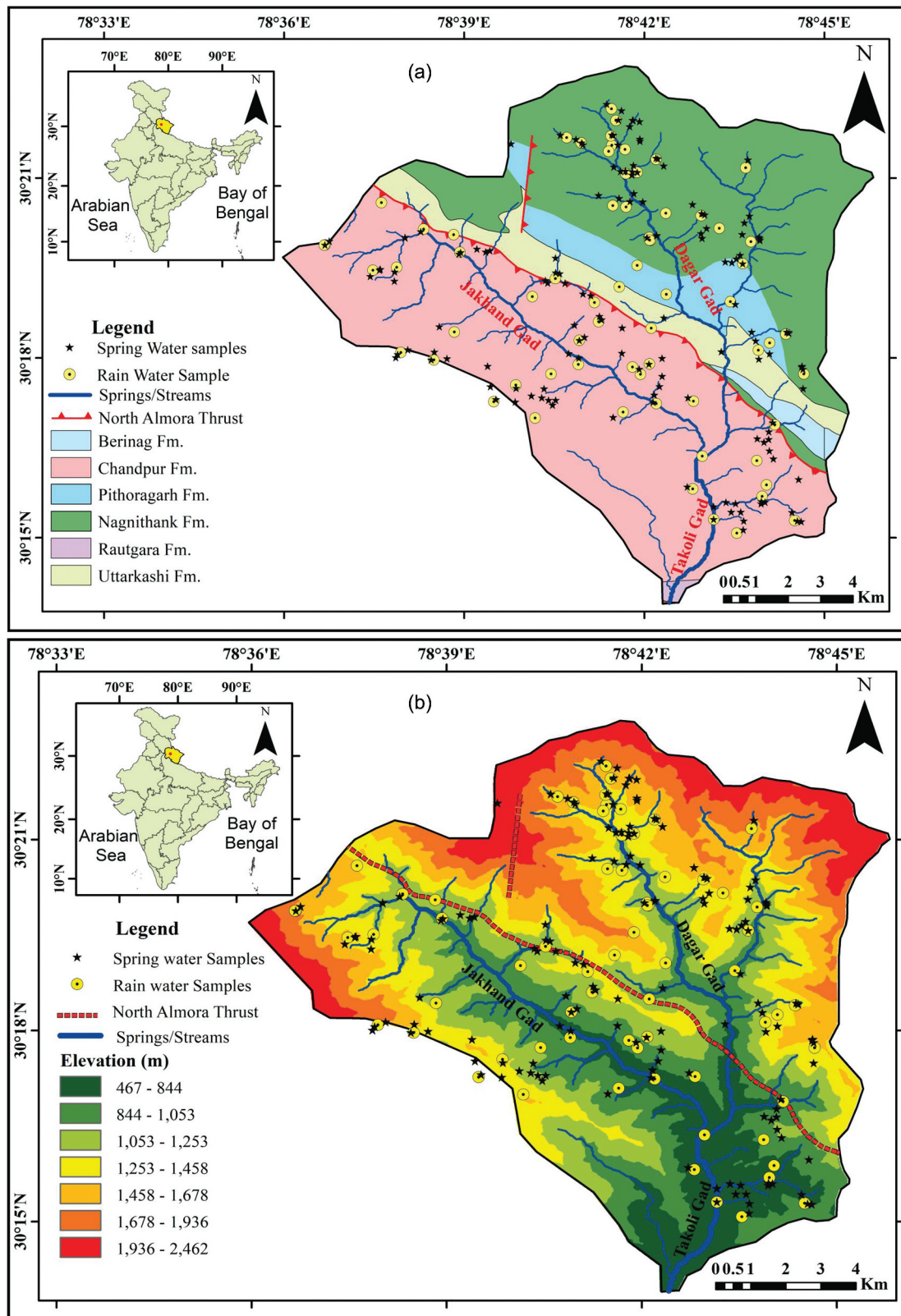


Figure 1. Sampling locations, and (a) geological map and (b) digital elevation map (DEM) of the Takoli Gad catchment. Figure modified after <https://bhukosh.gsi.gov.in>. Elevation data derived from Shuttle Radar Topography Mission (SRTM).

rockiness. In contrast, Typic Udorthents are moderately shallow, excessively drained soils that exhibit moderate erosion, slight stoniness, and a loamy to sandy-skeletal composition.

3 Material and methods

In the Takoli Gad region, 155 spring and stream sampling locations were identified (Fig. 1(a,b)) and repeatedly sampled

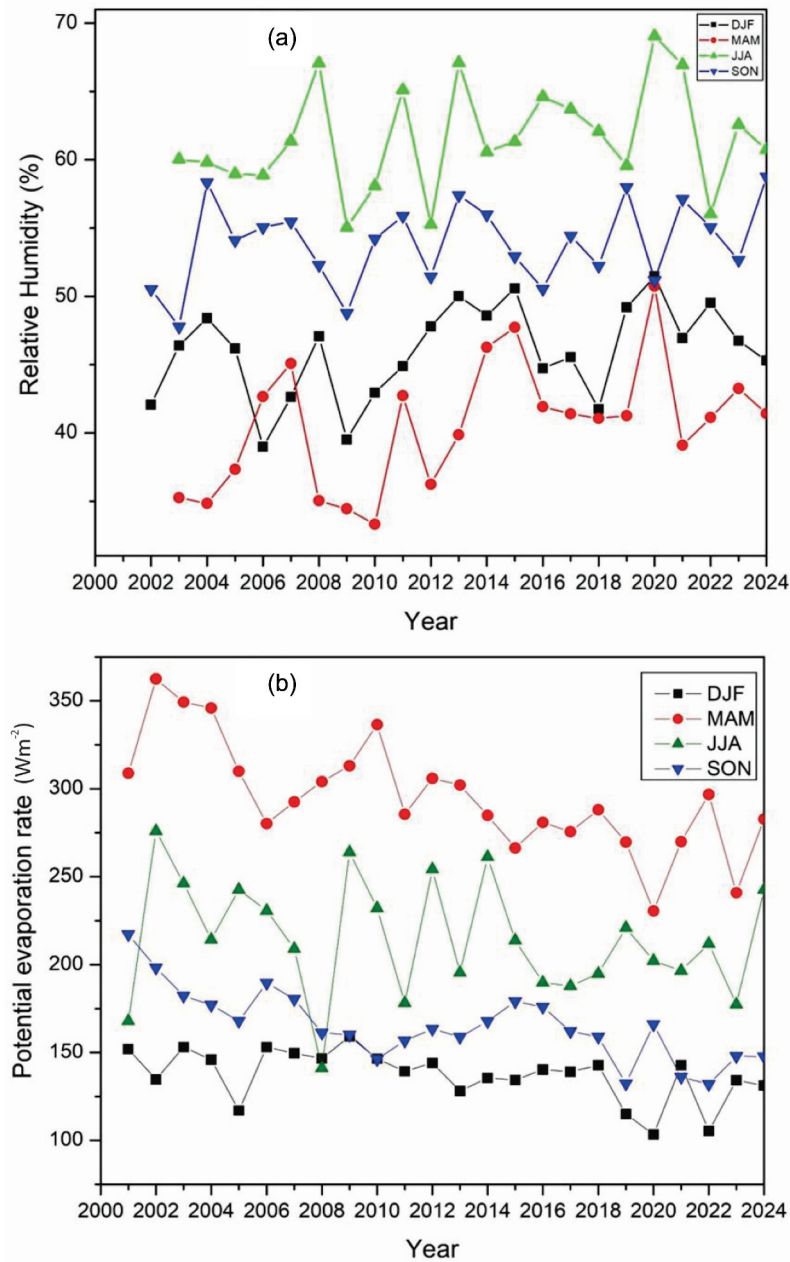


Figure 2. (a) Monthly variations in relative humidity across the study region from 2002 to 2024. (b) Seasonal potential evaporation (DJF: December, January, February; MAM: March, April, May; JJA: June, July, August; SON: September, October, November) in the study region from 2000 to 2024.

during the pre-monsoon, monsoon, and winter seasons of the year 2022. In addition to the spring water samples, rainwater samples were collected from the area during 2020–2021. A total of 434 spring water samples were collected, including 125 samples during the pre-monsoon season, 155 during the monsoon season, and 154 during the winter season, along an elevation transect of 575 to 1800 m. The sample collection was conducted at the same springs throughout all seasons. However, during the pre-monsoon season, the number of collected spring samples is lower than in the monsoon and winter seasons due to the drying of the springs. Similarly, 392 rainwater samples were sampled during the time frame, including 129 samples during the pre-monsoon season, 57 during the monsoon season, and 206 during the winter season, at elevations between 522 and 1745 m. A total of 75 raingauges

were installed at various altitudes across multiple strategically selected locations (Fig. 1(a,b)) to accurately represent the hydrological and isotopic characteristics of the region. We installed the 75 raingauges across 67 villages, covering an area of $\sim 143 \text{ km}^2$. Due to this vast coverage, collecting all samples on the same day was not feasible. Although the majority of samples were collected within 3–7 days of a rain event, some exceptions occurred at locations remote from our base camp. However, to minimize evaporation and water loss, we deployed a specially designed rainwater sampling device (Deshpande *et al.* 2010). Water samples were collected in pre-cleaned 125 mL high-density polyethylene (HDPE) bottles and capped tightly to avoid isotopic fractionation. Electrical conductivity was measured using a Eutech Eco EC Low tester, water temperature with a portable multi-thermometer, and pH

using a Hanna portable pH meter. Details of the sampling locations and the physical parameters are given in Tables S1 and S2 (see the Supplementary material). The collected samples were returned to the laboratory and stored at 4°C until analysis. The hydro-meteorological data for monthly precipitation, overland flow, and baseflow groundwater is used from the GLDAS model (GLDAS_NOAH10_M v. 2.0) at 0.25° spatial resolution. The soil moisture data used in this study were obtained from NASA Giovanni, which provides measurements at fixed, non-overlapping depth intervals (such as 0–10 cm, 10–20 cm, 20–30 cm, and 30–40 cm). In this dataset's convention, the upper boundary value belongs to the subsequent category—for example, 10 cm is included in the 10–20 cm layer, not in the 0–10 cm layer.

3.1 Isotopic analysis

The stable isotope analysis (δD and $\delta^{18}O$) of the collected water samples was conducted at the Stable Isotope Laboratory (SIL) of the Department of Earth Sciences, Indian Institute of Technology Roorkee (IITR). The isotopic analysis was conducted using the standard equilibration method (Maurya *et al.* 2009). For this purpose, ~300 μL of water samples were equilibrated with $He + CO_2$ (for $\delta^{18}O$ measurements) and $He + H_2$ gas (for δD analysis). The resulting equilibrated mixture of gases was then analysed using an Isotope Ratio Mass Spectrometer (IRMS) MAT 253 PLUS in continuous flow mode using Gas Bench II. In addition to the samples, several standards ($n = 20$) and repeated measurements of some samples ($n = 5$) were performed in each batch to ensure the accuracy, precision, and reliability of the measurements. The measurement reproducibility was better than 0.1‰ for $\delta^{18}O$ and 1‰ for δD . The IRMS was calibrated during the analysis using primary standards (VSMOW2, GRESP2, and SLAP2) and a working laboratory standard. Isotopic results were reported in standard delta (δ) notation and presented in per mill (‰) format against the international standard VSMOW2. Tritium analysis was carried out for 29 unfiltered spring water samples collected in 1 L pre-cleaned HDPE sample bottles and securely sealed to prevent exposure to air. The analysis was carried out at the Nuclear Hydrology Laboratory at the National Institute of Hydrology (NIH), Roorkee. The detailed methodology of tritium measurement techniques and spring water age determination from the obtained tritium data is provided separately (see the Supplementary material).

3.2 End-member contribution estimation and spring water recharge

In addition to tracing the role of various hydrological processes in shaping the stable isotopic signature of the catchment, we estimated the relative contribution of seasonal precipitation to spring water. The mixing model assumes that different seasonal precipitation has distinct isotopic compositions, and that spring water discharge reflects the mixture of these inputs, while average $\delta^{18}O$ values of rainwater and groundwater define the extreme ends of this mixing model (Penna *et al.* 2014, Zuecco *et al.* 2019).

Also, any deviation in spring water composition relative to rainwater is attributed to the fractional contribution from groundwater (which represents the isotopic mean of annual precipitation values; Clark and Fritz 1997). The estimation is carried out using the average $\delta^{18}O$, d -excess compositions of the precipitation and that of the spring water and the three-component mixing mass balance equation, similar to Maurya *et al.* (2011) and Sharma *et al.* (2025). Under the assumption that the average $\delta^{18}O$ and d -excess signatures of the spring water are the result of the mixing of precipitation during pre-monsoon, monsoon, and winter precipitation, the mass balance equation can be expressed as:

$$PM + M + W = S \quad (1)$$

where PM, M, W, and S denote the pre-monsoon, monsoon, winter precipitation and spring water, respectively. The fractions of PM, M and W mix to give a unified value of 1, so the Equation (1) becomes

$$C_{PM} + C_M + C_W = 1 \quad (2)$$

where C_{PM} , C_M , and C_W represent the contribution of pre-monsoon, monsoon, and winter precipitation, respectively, to the spring water budget. Using the above-mentioned variables, when expressed along with their respective average $\delta^{18}O$, d -excess values, Equation (1) can be written as

$$C_{PM} \times \delta^{18}O_{PM} + C_M \times \delta^{18}O_M + C_W \times \delta^{18}O_W = \delta^{18}O_S \quad (3)$$

$$C_{PM} \times d_{PM} + C_M \times d_M + C_W \times d_W = d_S \quad (4)$$

Solving Equations (3) and (4) for three variables, i.e. C_{PM} , C_M , and C_W , yields

$$C_W = \frac{[(\delta^{18}O_S - \delta^{18}O_{PM})(d_M - d_{PM})] - [(\delta^{18}O_M - \delta^{18}O_{PM})(d_S - d_{PM})]}{[(\delta^{18}O_W - \delta^{18}O_{PM})(d_M - d_{PM})] - [(\delta^{18}O_M - \delta^{18}O_{PM})(d_W - d_{PM})]} \quad (5)$$

$$C_M = \frac{[(\delta^{18}O_S - \delta^{18}O_{PM})(d_W - d_{PM})] - [(\delta^{18}O_W - \delta^{18}O_{PM})(d_S - d_{PM})]}{[(\delta^{18}O_M - \delta^{18}O_{PM})(d_W - d_{PM})] - [(\delta^{18}O_W - \delta^{18}O_{PM})(d_W - d_{PM})]} \quad (6)$$

After estimation, the fractional C_W , C_M , and C_{PM} were approximated using Equation (2). When converted into percentage form, C_{PM} , C_M , and C_W yield a contribution in all three seasons (pre-monsoon, monsoon, and winter).

3.3 Uncertainty analysis

The uncertainty in the contribution arises from spatiotemporal variability in the end-member tracer signature and analytical uncertainties associated with isotopes and d -excess measurements (Schmieder *et al.* 2018, Popp *et al.* 2025). The three-component mixing model was used, and uncertainty analyses were carried out for the pre-monsoon, monsoon, and winter seasons using $\delta^{18}O$ and d -excess values. Even though analytical uncertainty for d -excess is estimated from both $\delta^{18}O$ and δ^2H measurements, the

uncertainty in the precision of $\delta^2\text{H}$ (± 1) is much higher than the uncertainty in $\delta^{18}\text{O}$ (± 0.1). The Gaussian error propagation based on the first-order Taylor expansion is used to quantify uncertainty in a two- or three-component system (Genereux 1998, Uhlenbrook and Hoeg 2003, Correa *et al.* 2019). The uncertainty (W_y), which is the function of inputs $x_1, x_2 \dots x_n$, was estimated by the error propagation method (Genereux 1998, Lone *et al.* 2022).

$$W_y = \sqrt{\left(\frac{\partial y}{\partial x_1} W_{x_1}\right)^2 + \left(\frac{\partial y}{\partial x_2} W_{x_2}\right)^2 + \dots + \left(\frac{\partial y}{\partial x_n} W_{x_n}\right)^2} \quad (7)$$

Whereas W is the uncertainty in the variable shown in the subscript, X_1, X_2 , and X_n are the end-members, and y is the contribution of a specific component to the spring discharge.

According to Equation (7), uncertainty in the winter contribution (C_W) is calculated as:

$$W_{C_W} = \sqrt{\left(\frac{\partial C_W}{\partial \delta^{18}\text{O}_S} W_{\delta^{18}\text{O}_S}\right)^2 + \left(\frac{\partial C_W}{\partial \delta^{18}\text{O}_{PM}} W_{\delta^{18}\text{O}_{PM}}\right)^2 + \left(\frac{\partial C_W}{\partial \delta^{18}\text{O}_M} W_{\delta^{18}\text{O}_M}\right)^2 + \left(\frac{\partial C_W}{\partial \delta^{18}\text{O}_W} W_{\delta^{18}\text{O}_W}\right)^2 + \left(\frac{\partial C_W}{\partial d_S} W_{d_S}\right)^2 + \left(\frac{\partial C_W}{\partial d_{PM}} W_{d_{PM}}\right)^2 + \left(\frac{\partial C_W}{\partial d_M} W_{d_M}\right)^2 + \left(\frac{\partial C_W}{\partial d_W} W_{d_W}\right)^2} \quad (8)$$

where $W_{\delta^{18}\text{O}}$ and W_d represent the analytical and seasonal uncertainties of spring and end-member isotope compositions. Similar calculations were applied to estimate uncertainties for pre-monsoon (C_{PM}) and monsoon (C_M) contributions.

3.4 Rainout estimation

In addition to the relative contribution of precipitation, we have estimated the amount of moisture removed from the cloud at the same transect over the region using the simple Rayleigh fractionation model (Clark and Fritz 1997, Dalai *et al.* 2002, Das and Rai 2022). The removal of the cloud droplets from the condensing cloud vapour is always under equilibrium, and isotopic composition of precipitation can be represented by Rayleigh condensation equation as

$$R_l = \alpha R_0 f^{\alpha-1} \quad (9)$$

where R_l is the isotopic composition ($\delta^{18}\text{O}$ and δD) in the rainwater (at the highest elevated region), R_0 is the initial ratio of the vapour (at the lowest elevated region), f is the fraction of

vapour remaining in the cloud, and α is the temperature-dependent isotope fractionation factor during the vapour-liquid transition. Equation (9) can be written as

$$f = \exp[\ln(R_l/\alpha R_0)/(\alpha - 1)] \quad (10)$$

as $\alpha R_0 = R_{l_0}$, so (R_{l_0} is the initial liquid ratio)

$$f = \exp[\ln(R_l/\alpha R_{l_0})/(\alpha - 1)] \quad (11)$$

or

$$f = \exp\{[\ln(\delta_l + 1000)/(\delta_0 + 1000)]/(\alpha - 1)\} \quad (12)$$

For $\delta^{18}\text{O}$ enabled rainout estimation, α values vary between 1.0093 and 1.0112, while for δD , the same is considered between 1.079 and 1.105 (Dalai *et al.* 2002).

4 Results

4.1 Physical parameters and isotopic compositional variations in the Takoli Gad catchment

The electrical conductivity (EC) of spring water varies from 70 to 1120 $\mu\text{S}/\text{cm}$ during the pre-monsoon season, from 50 to 730 $\mu\text{S}/\text{cm}$ in the monsoon, and from 70 to 930 $\mu\text{S}/\text{cm}$ in the winter period (see Supplementary material, Table S1). Similarly, the EC values of the rainwater range from 20 to 250 $\mu\text{S}/\text{cm}$, 10 to 90 $\mu\text{S}/\text{cm}$, and 20 to 240 $\mu\text{S}/\text{cm}$ for pre-monsoon, monsoon, and winter seasons, respectively (see Supplementary material, Table S2). The temperature of the collected spring water samples shows a strong seasonal dependence, varying from 10.8 to 38.2°C, 17.6 to 25.5°C, and 7.3 to 20.7°C for the pre-monsoon, monsoon and winter seasons, respectively. Details regarding the physical parameters as well as the isotopic compositions (δD , $\delta^{18}\text{O}$ and d -excess ($d = \delta\text{D} - 8 \times \delta^{18}\text{O}$) of the water samples (both spring and precipitation) are given in Table S1 and S2 (see Supplementary material), respectively, whereas the variation ranges of the same are presented in Table 1.

4.1.1 $\delta^{18}\text{O}$ - δD relationships through space and time

The LMWLs for Takoli Gad catchment were determined based on $\delta^{18}\text{O}$ - δD relationships of rainwater showing distinct slopes and intercepts. Slopes for the winter (8.10 ± 0.19) and monsoon (8.01 ± 0.14) seasons closely align with the global meteoric water line (GMWL), whereas the lower slope in pre-monsoon season (7.50 ± 0.08) indicates a relatively higher degree of sub-cloud evaporation (Fig. 3(a),

Table 1. $\delta^{18}\text{O}$, δD value and deuterium excess (in ‰) values of spring and precipitation samples in Takoli Gad catchment.

Seasons	Values	Samples					
		Spring water			Precipitation		
		Min.	Max.	Avg.	Min.	Max.	Avg.
Pre-monsoon	δD	-65.9	-43.8	-55.7 ± 3.07 (n = 125)	-55.29	64.86	15.29 ± 30.50 (n = 129)
	$\delta^{18}\text{O}$	-9.5	-6.55	-8.1 ± 0.42	-9.0	7.99	-0.67 ± 4.03
	d -excess	4.35	13.54	9.1 ± 1.75	-3.7	22.36	9.9 ± 4.05
Monsoon	δD	-73.1	-51.2	-60.06 ± 4.15 (n = 155)	-110.14	-13.95	-64.45 ± 15.53 (n = 57)
	$\delta^{18}\text{O}$	-10.5	-7.71	-8.81 ± 0.55	-14.81	-3.07	-9.21 ± 1.92
	d -excess	5.71	13.56	10.3 ± 1.47	1.47	13.06	9.3 ± 2.05
Winter	δD	-65.4	-50.6	-57.79 ± 2.43 (n = 154)	-69.59	31.79	-34.74 ± 17.16 (n = 206)
	$\delta^{18}\text{O}$	-9.44	-7.21	-8.44 ± 0.43	-9.88	-3.0	-5.36 ± 2.01
	d -excess	1.1	14.63	9.7 ± 2.32	0.16	34.87	13.7 ± 5.42

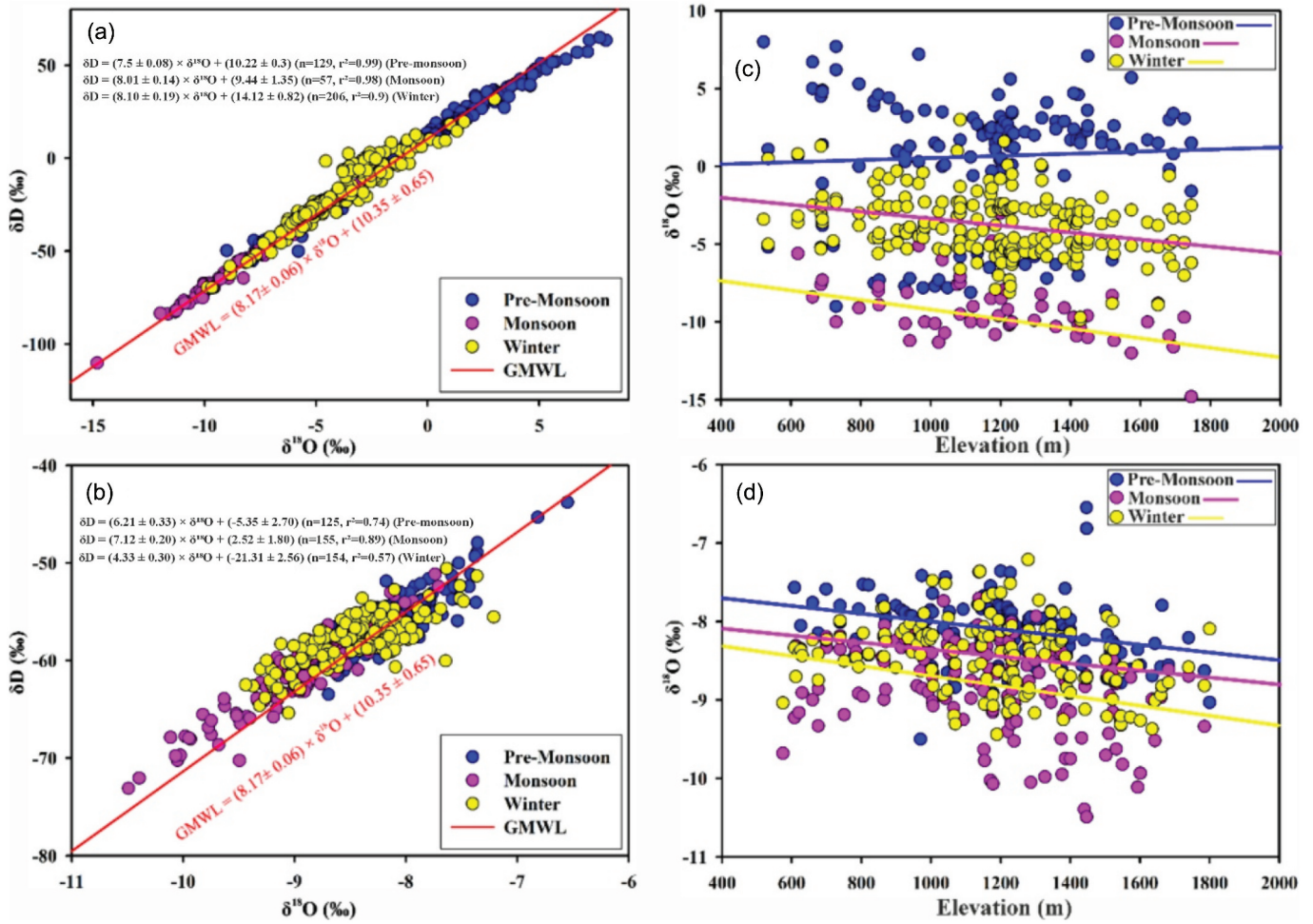


Figure 3. (a) $\delta^{18}O$ and δD values of precipitation water along with the global meteoric water line (GMWL) in the Takoli Gad catchment. (b) $\delta^{18}O$ and δD values of spring water along with the GMWL. The figure infers the preservation potential of the meteoric water signatures by the spring water during monsoon season, whereas the role of evaporation is evident from the $\delta^{18}O$ – δD slope during the pre-monsoon as well as the winter season. (c) $\delta^{18}O$ -enabled altitude effect in rainwater. (d) $\delta^{18}O$ -enabled altitude effect in spring water.

Table 2. $\delta^{18}O$ – δD slope and deuterium excess (in ‰) values of spring and precipitation samples of Takoli Gad catchment across various season and elevations.

Season	Elevation (m)	Samples					
		Spring water			Precipitation		
		$\delta^{18}O$ – δD slope	Intercept	<i>d</i> -excess (‰)	$\delta^{18}O$ – δD slope	Intercept	<i>d</i> -excess (‰)
Pre-monsoon	All combined	6.21 ± 0.33 ($n = 125$, $r^2 = 0.74$)	-5.35 ± 2.70	9.12 ± 1.7	7.50 ± 0.08 ($n = 129$, $r^2 = 0.99$)	10.2 ± 0.31	9.9 ± 4.0
	<1000	7.17 ± 0.70 ($n = 30$, $r^2 = 0.8$)	1.78 ± 5.61	8.4 ± 1.5	7.25 ± 0.09 ($n = 38$, $r^2 = 0.99$)	9.99 ± 0.48	9.3 ± 4.8
	1000–1500	5.92 ± 0.40 ($n = 77$, $r^2 = 0.73$)	-7.49 ± 3.3	9.2 ± 1.7	7.78 ± 0.1 ($n = 75$, $r^2 = 0.98$)	10.05 ± 0.39	9.8 ± 3.5
	>1500	7.86 ± 1.43 ($n = 21$, $r^2 = 0.61$)	8.82 ± 12.13	9.9 ± 1.9	7.27 ± 0.41 ($n = 21$, $r^2 = 0.95$)	12.8 ± 1.17	11.8 ± 4.3
Monsoon	All combined	7.12 ± 0.20 ($n = 155$, $r^2 = 0.89$)	2.52 ± 1.80	10.3 ± 1.5	8.01 ± 0.14 ($n = 57$, $r^2 = 0.98$)	9.4 ± 1.35	9.3 ± 2.0
	<1000	7.97 ± 0.56 ($n = 39$, $r^2 = 0.84$)	9.23 ± 4.9	9.4 ± 1.4	7.99 ± 0.39 ($n = 14$, $r^2 = 0.97$)	8.92 ± 3.36	9.3 ± 2.2
	1000–1500	7.20 ± 0.24 ($n = 92$, $r^2 = 0.9$)	3.36 ± 2.12	10.4 ± 1.4	8.02 ± 0.2 ($n = 33$, $r^2 = 0.98$)	9.02 ± 1.92	9.0 ± 2.1
	>1500	7.41 ± 0.45 ($n = 24$, $r^2 = 0.92$)	5.79 ± 4.15	11.1 ± 1.1	8.35 ± 0.31 ($n = 9$, $r^2 = 0.99$)	13.8 ± 3.46	10.1 ± 2.6
Winter	All combined	4.33 ± 0.30 ($n = 154$, $r^2 = 0.57$)	-21.31 ± 2.56	9.7 ± 2.3	8.10 ± 0.19 ($n = 206$, $r^2 = 0.9$)	14.12 ± 0.82	13.7 ± 5.4
	<1000	2.87 ± 1.0 ($n = 39$, $r^2 = 0.17$)	-33.3 ± 8.66	9.1 ± 2.0	7.34 ± 0.41 ($n = 53$, $r^2 = 0.85$)	10.1 ± 1.4	12.2 ± 5.1
	1000–1500	4.36 ± 0.36 ($n = 91$, $r^2 = 0.61$)	-20.83 ± 3.0	9.8 ± 2.4	8.32 ± 0.24 ($n = 126$, $r^2 = 0.9$)	15.2 ± 1.08	14.0 ± 5.6
	>1500	4.42 ± 0.93 ($n = 24$, $r^2 = 0.84$)	-20.7 ± 8.1	10.6 ± 2.3	9.34 ± 0.41 ($n = 26$, $r^2 = 0.95$)	21.9 ± 2.13	15.6 ± 4.6

Table 2). Similarly, the best-fit lines (BFL) for the δD and $\delta^{18}O$ compositions of spring water exhibit varying seasonal slopes. The slopes for the monsoon (7.12 ± 0.20), pre-monsoon (6.21 ± 0.33), and winter (4.33 ± 0.30) seasons were consistently lower than the corresponding LMWL slopes, indicating that spring water was influenced by kinetic evaporation process before or during recharge (Fig. 3(b) and Table 2).

In addition to the seasonal variations, the associated controlling factors for the spatial variations (with respect to varying altitudes) of the obtained spring and precipitation isotopic compositions were also analysed. In the Takoli Gad region, the elevation of the springs and precipitation samples varies from 575 to 1800 m and 522 to 1745 m, respectively. Therefore, the δD – $\delta^{18}O$ relationship and *d*-excess of the samples (both spring water and

rainfall) were analysed across different elevation transects i.e. <1000 m, 1000–1500 m and >1500 m, as given in Table 2 and Fig. S2 (see Supplementary material). Spring water showed a relatively lower slope at the 1000–1500 m transect during pre-monsoon (5.92 ± 0.40) and monsoon (7.20 ± 0.24) seasons, while the winter season showed a progressive increase in slope from lower to higher altitudes. Furthermore, precipitation samples from the pre-monsoon season showed an almost uniform slope at each elevation transect, while during monsoon season, slopes were nearer to LMWL and consistently higher than pre-monsoon precipitation at all elevations. Winter precipitation slopes exhibited a progressive increase from lower to higher altitude, and higher slope than the LMWL, reflecting the impact of westerlies. Like the δD – $\delta^{18}O$ slopes, d -excess values of springs and precipitation differ along different elevation transects (i.e. <1000 m, 1000–1500 m and >1500 m) and are presented in Table S1 and S2 (see Supplementary material), whereas the variation ranges of the same are presented in Table 2. Twenty-nine selected spring/stream samples are measured for their tritium dating, and their tritium concentration varies between 1.5 and 14.2 TU (Table S3; Fig. 4(a)).

4.1.2 End-member contribution and rainout percentage estimation

Estimation of end-member contribution (pre-monsoon, monsoon, and winter precipitation) to the spring water budget showed a dominance of monsoon precipitation ($82 \pm 21\%$), and a relatively minor contribution from pre-

monsoon ($8 \pm 14\%$) and winter ($10 \pm 18\%$) precipitation (Fig. 5). The altitude– $\delta^{18}O$ relationship for the rain and spring water samples of the Takoli Gad catchment for the pre-monsoon, monsoon and winter seasons are given in Fig. 3(c,d) and Table 3.

For rainout percentage estimation, considering the $\delta^{18}O$ and δD signatures of the rainwater during the monsoon period at the lowest and highest elevation locations (Dangchaura-1; $\delta^{18}O$: -5.7% , δD : -37.0% and Molthachok-1; $\delta^{18}O$: -14.8% , δD : -110.1% , respectively; see Table S2), the $\delta^{18}O$ -enabled rainout percentage was estimated to be 56–63% (average: $59 \pm 5\%$), whereas the estimation incorporating the δD of the samples yields a percentage of 53–63% (average: $58 \pm 7\%$). The rainout estimation involving both compositions ($\delta^{18}O$ and δD) overlap each other and fall within the uncertainty, indicating the reliability of our calculation. So, it can be concluded that $\sim 59 \pm 6\%$ of the moisture was removed from the S–N ascending cloud in the study area. Surprisingly, the estimated rain in the Takoli Gad is found to be significantly higher than in the other Himalayan catchments (Yamuna: 25%, Ganga: 30% and Teesta: 35%; Dalai *et al.* 2002, Das and Rai 2022). This abruptly high value may be the result of torrential rainfall events, which are common in the mountainous region during the monsoon season. Therefore, to cross-validate our rainout estimation, we have re-estimated the same considering the highest elevation sampling location, Silod-2: $\delta^{18}O$: -9.7% , δD : -68.4% (see Supplementary material, Table S2). This gives the moisture removal percentage from the cloud as 31–36% (average: $33 \pm 4\%$; using $\delta^{18}O$) and 27–34% (average: $31 \pm 5\%$; using δD) with an overall average of $32 \pm 4\%$.

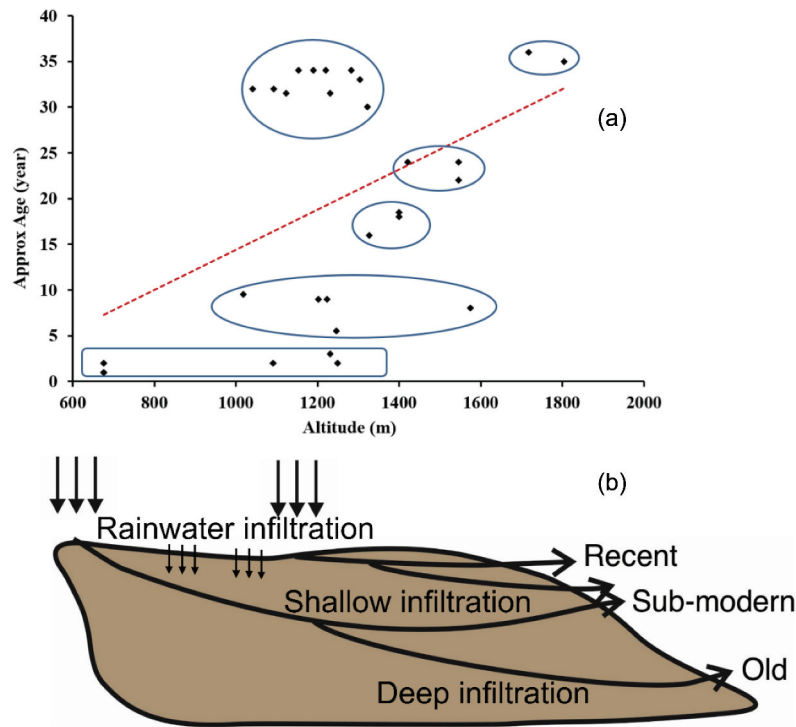


Figure 4. (a) Scatter plot between altitude and approximate age of the water, inferring a random distribution and mixing of water (of various ages) in the watershed. (b) Schematic diagram (not to scale) showing rainwater infiltration (downward arrows) and percolation into shallow and deep infiltration zones. Shallow infiltration has a shorter residence time (recent water: 1–2 years), while deep infiltration has a longer residence time (old water: >30 years). The mixing of these waters at different altitudes results in spring water of varying ages. The arrow endpoints represent spring discharge points, indicating where shallow and deep infiltration waters emerge.

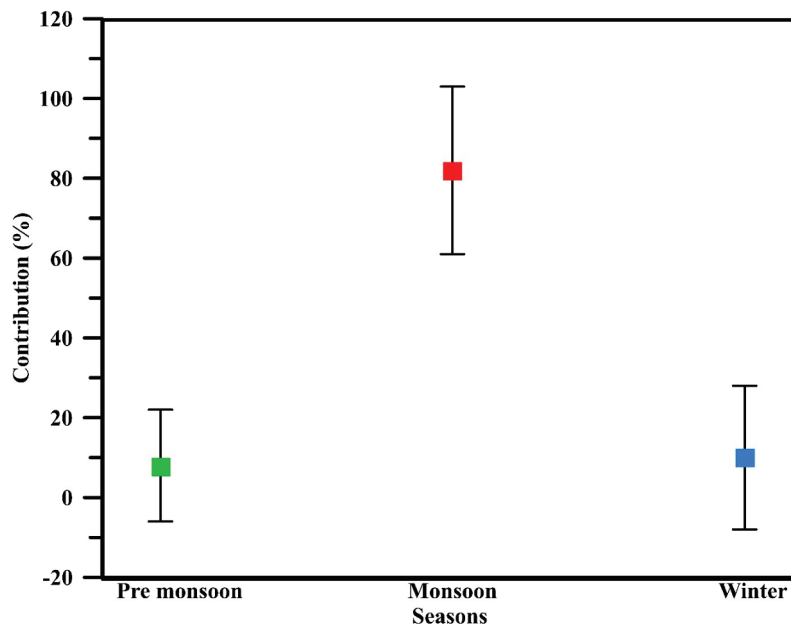


Figure 5. Box plot with error bars showing seasonal rainwater contribution to the spring water.

Table 3. $\delta^{18}\text{O}$ enabled altitude effect (per 100 m) for the spring and precipitation samples in Takoli Gad catchment.

Seasons	Samples			
	Spring water		Precipitation	
	Altitude effect	Intercept	Altitude effect	Intercept
Pre-monsoon	$-0.05 \pm 0.0001\text{‰}$ (n = 128, $P < .0001$)	-7.5 ± 1.4	$0.07 \pm 0.0012\text{‰}$ (n = 128, $P = .9006$)	-0.18 ± 1.44
Monsoon	$-0.06 \pm 0.0002\text{‰}$ (n = 155, $P < .0001$)	-8.05 ± 1.9	$-0.31 \pm 0.0008\text{‰}$ (n = 56, $P < .0001$)	-5.49 ± 0.99
Winter	$-0.04 \pm 0.0001\text{‰}$ (n = 154, $P < .0001$)	-7.91 ± 1.5	$-0.23 \pm 0.0005\text{‰}$ (n = 205, $P = .05$)	-7.91 ± 0.15

4.2 Variation in hydro-meteorological components of the Takoli Gad

In addition to contribution estimation, this study delineates the spring recharge mechanism using hydro-meteorological data from GLDAS and AIRX3STM models. The soil moisture content exhibits both seasonal and depth-wise variability. We analysed soil moisture at different depths (0–10, 10–40, 40–100, and 100–200 cm; see Supplementary material, Figs S3, S4, S5, S6) throughout the sampling period to gain insight into this variability, as shown in Table 4. This pattern reflects the complex interactions between precipitation sources (Indian Summer Monsoon (ISM) and westerlies), evapotranspiration, vegetation, and soil characteristics that shape the local water balance. The observed variability highlights the dynamic and interconnected nature of the regional climate system.

The overland runoff, including throughflow ($\text{kg m}^{-2} \text{s}^{-1}$) and baseflow–groundwater runoff ($\text{kg m}^{-2} \text{month}^{-1}$), exhibits distinct seasonal patterns. Overland runoff surges during the monsoon, followed by lower values in the winter and pre-monsoon seasons (Fig. 6(a)). During the monsoon season, groundwater baseflow fluctuates between 0.0254 and 0.517 $\text{kg m}^{-2} \text{month}^{-1}$, with an average of 0.098 $\text{kg m}^{-2} \text{month}^{-1}$ (Fig. 6(b)) recorded from 2000 to 2024. Interestingly, the inter-annual variability in overland runoff is more pronounced, with some years experiencing significantly higher or lower flows. In contrast, the baseflow–groundwater runoff maintains a more consistent contribution, with the highest

values occurring during the monsoon and the lowest during the drier periods.

5 Discussion

5.1 $\delta^{18}\text{O}$ – δD relationship in precipitation and spring water

The recharge mechanism of springs in the Takoli Gad catchment is influenced by various factors, including geology, slope, topography, soil moisture content, overland flow, groundwater baseflow, and seasonal precipitation sources. The temporal and spatial changes in the isotopic signature of samples depend on the seasons, moisture sources, elevation, relative humidity, and evaporation over the regions, which are discussed in detail.

The isotopic analysis of water from springs and precipitation, combined with hydrogeological studies and remote sensing hydro-meteorological data from the Takoli Gad catchment, aids in understanding the role of various hydrological processes in isotopic alterations, fractional contribution during different seasons, amount of moisture removal from the precipitation in the region and mechanisms of spring recharge. The composition of water is used to get insight into the origin of moisture sources and transport pathways taken by moisture-laden air masses, as indicated by the $\delta^{18}\text{O}$ – δD relationship with the meteoric water lines (Hren *et al.* 2009, Kumar *et al.* 2021, Chen *et al.* 2024, Abdelfadel *et al.* 2025,

Table 4. Variation of soil moisture content at depths of 0–10, 10–40, 40–100, and 100–200 cm across different seasons using the GLDAS model (GLDASNOAH025_Mv2.1) in kg/m² for the period 2002–2024.

Time	January	February	March	April	May	June	July	August	September	October	November	December
Soil moisture content at depths of 0–10 cm (kg/m ²)												
Mean	18.42	21.11	22.02	22.77	23.22	22.84	28.49	30.71	29.54	24.65	20.05	17.84
Max.	28.51	27.28	28.63	29.69	29.52	28.83	31.75	33.54	33.37	30.42	26.75	25.24
Min.	15.05	16.72	16.58	16.55	17.05	14.27	21.38	26.83	23.72	20.04	15.52	14.50
SD	3.07	2.81	3.18	2.90	2.96	3.82	2.62	1.40	2.36	2.51	2.51	2.44
Soil moisture content at depths of 10–40 cm (kg/m ²)												
Mean	69.60	73.61	75.50	74.97	73.54	72.60	85.57	93.23	90.42	79.69	72.23	69.30
Max.	84.98	84.73	88.71	88.95	91.94	87.33	96.20	101.97	100.37	91.58	81.67	79.77
Min.	62.60	64.26	64.34	63.04	61.43	58.39	67.33	81.61	79.56	70.26	64.29	61.28
SD	5.48	5.84	6.75	6.54	6.50	8.00	7.46	4.47	5.58	5.23	4.65	4.58
Soil moisture content at depths of 40–100 cm (kg/m ²)												
Mean	140.89	142.02	144.62	143.31	140.64	139.24	158.68	183.57	182.37	165.02	150.36	143.29
Max.	169.93	167.19	168.83	170.55	178.32	167.92	190.98	203.31	202.65	188.18	169.51	159.20
Min.	125.66	128.81	125.88	122.43	122.29	118.61	123.27	144.69	167.41	146.77	134.94	128.67
SD	10.27	9.91	11.54	12.10	11.91	12.91	18.72	12.34	10.55	10.52	9.40	8.62
Soil moisture content at depths of 100–200 cm (kg/m ²)												
Mean	266.16	263.40	261.74	259.05	255.77	252.98	264.53	300.17	311.33	296.81	280.77	271.18
Max.	286.79	287.81	283.23	286.03	292.29	287.99	322.85	342.94	345.11	327.56	304.67	292.18
Min.	244.62	242.60	241.21	237.65	236.99	232.97	235.23	242.64	270.62	271.83	258.51	249.79
SD	12.37	12.47	12.69	12.64	12.61	12.90	22.97	25.51	19.61	16.99	13.89	12.43

Quiroz Londoño *et al.* 2025). The LMWL for the Takoli Gad watershed, based on $\delta^{18}\text{O}$ and $\delta^2\text{H}$ values of rainwater during the monsoon season, has a slope of 8.01 ± 0.14 (Table 2, Fig. 3(a)). This slope is closely aligned with the GMWL (8.17, Rozanski *et al.* 1993) and those of other key regions like the Himalayan and Tibetan Plateau meteoric water line (HTPMWL: 8.52, Hren *et al.* 2009), the Khumbu Valley and adjacent valley of Nepal Himalaya (~ 8.12 – 8.18 , Balestrini *et al.* 2014, Matheswaran *et al.* 2019, Hassenruck-Gudipati *et al.* 2023), and the western and northwestern Himalaya (7.95 and 7.59, Kumar *et al.* 2010, Jeelani *et al.* 2013). These alignments suggest that the rainwater isotopic compositions are well preserved in the Takoli Gad watershed.

The relationship between δD and $\delta^{18}\text{O}$ in precipitation samples across different seasons, namely pre-monsoon, monsoon, and winter, reveals significant findings (Fig. 3(a)). The results show that the steepest slope (8.10 ± 0.19) occurs during winter, while notably low slopes are recorded during both monsoon (8.01 ± 0.14) and pre-monsoon (7.5 ± 0.08) seasons (Table 2). In contrast, spring water samples in monsoon are characterized by the highest slope (7.12 ± 0.2), followed by the pre-monsoon (6.21 ± 0.33) (Table 2, Fig. 3(b)) and winter season (4.33 ± 0.3) (Table 2), which deserve detailed scientific study to unravel the complexities of water movement and sourcing within the Takoli Gad catchment area. These changes in slope are associated with evaporation rates and meteorological factors in the Himalayas and other regions (Hren *et al.* 2009, Kumar *et al.* 2010, Florea *et al.* 2017, Chen *et al.* 2024, Lyon *et al.* 2024). The time series graph over the ISM-dominated study region shows that the rate of evaporation is high and relative humidity is low during the pre-monsoon (MAM) (Fig. 2(a,b)), which modifies the isotopic signature of precipitation and infiltrating soil moisture, leading to a lower slope in pre-monsoon precipitation (7.5 ± 0.08) and spring water (6.21 ± 0.33). During the active monsoon months (June to August), deviations from

the GMWL can occur due to variations in atmospheric temperature and lower humidity at the onset of the monsoon in June. As the monsoon progresses, the overall high relative humidity limits evaporation, resulting in a slope that remains closer to the GMWL (Fig. 3(a,b)).

In contrast, during winter season, when westerly winds bring precipitation, the slope is 8.1 ± 0.19 , similar to that of the GMWL, which may be attributed to lower temperatures and a reduced potential evaporation rate during this particular period of the year (Fig. 2(b)). The varying slopes of spring isotopic signatures across different seasons necessitate a deeper understanding of the hydro-meteorological processes influencing these changes in the Takoli Gad catchment. Interestingly, the δD – $\delta^{18}\text{O}$ slope of spring water during the winter (4.33 ± 0.30) (Table 2, Fig. 3(b)) is significantly lower compared to those observed during the pre-monsoon and monsoon seasons (for both spring and precipitation samples). As winter is the coldest season of the year, and precipitation is minimal during this time, this cold and arid climate seems to decrease the relative humidity and initiate evaporation, which is also evident from the other regions of the Himalayan Mountains as well as the Tibetan plateau (Tian *et al.* 2007, Bershaw *et al.* 2012, 2016).

5.2 Altitudinal variations of hydrological processes in the Takoli Gad catchment

In addition to seasonal variations, the controlling factors associated with the spatial variations (with respect to varying altitude) in the obtained spring and precipitation isotopic compositions have been discussed. From Tables S1 and S2 (see Supplementary material), it is evident that in the Takoli Gad region, elevation of the spring water and precipitation samples ranges from 575 to 1800 m and from 522 to 1745 m, respectively. Therefore, we have discussed the δD – $\delta^{18}\text{O}$ relationship of the samples (both spring water and

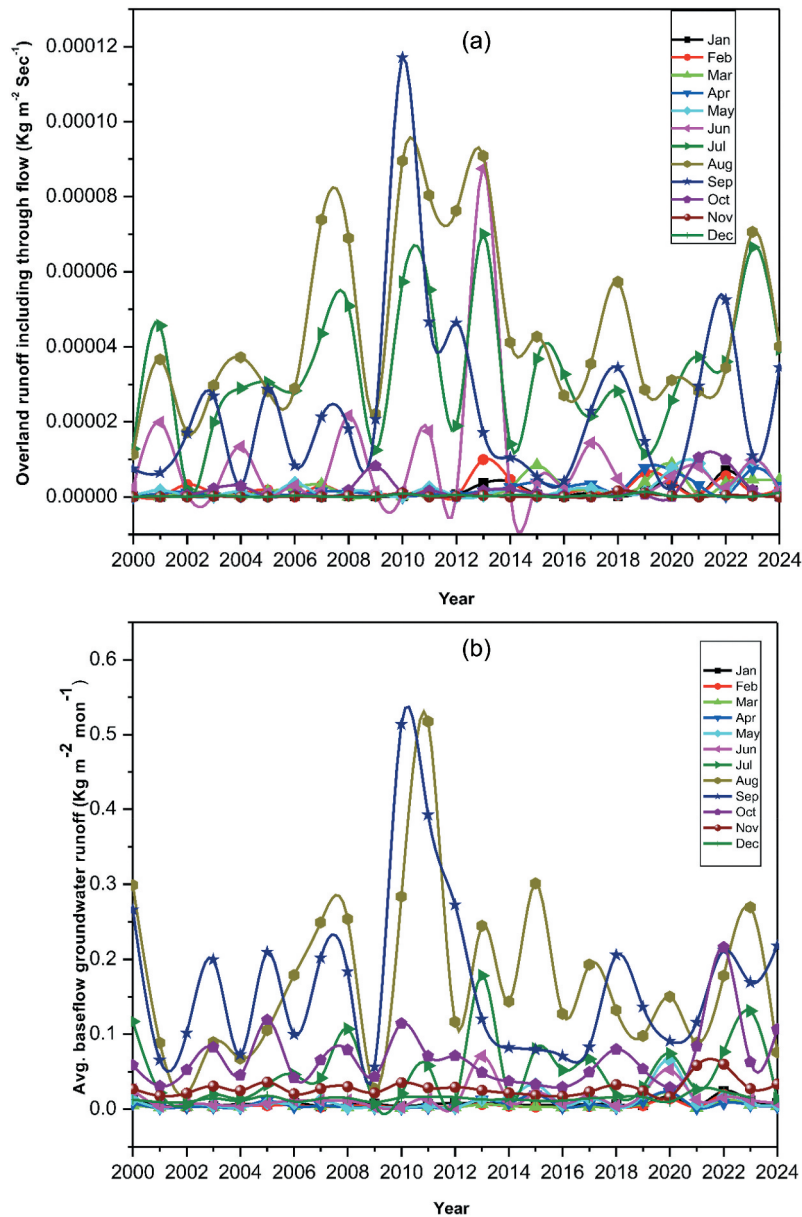


Figure 6. Inter-annual variability of overland flow and baseflow groundwater runoff based on the GLDAS model (GLDAS_NOAH025_M v. 2.0). (a) Inter-annual overland runoff ($\text{kg m}^{-2} \text{s}^{-1}$) including throughflow in the region from 1980 to 2024. (b) Baseflow groundwater runoff from 1948 to 2024 ($\text{kg m}^{-2} \text{month}^{-1}$).

rainfall) in three elevation transects, i.e. <1000 m, 1000–1500 m and >1500 m, in the study area.

During the pre-monsoon season, the stable isotopic compositions of spring water reveal distinct relationships based on these elevation ranges, specifically for the sampling elevations below 1000 m, between 1000 and 1500 m, and above 1500 m (Table 2; see Supplementary material, Fig. S2). The above equations indicate that the spring water in the elevation ranges of 1000–1500 m is experiencing a relatively higher magnitude of evaporation (slope: 5.92) during the pre-monsoon summer spells, owing to a higher temperature during the course of the year. Surprisingly, the samples collected from the lower elevation (where the temperature is relatively higher) and at higher elevations have a $\delta\text{D}-\delta^{18}\text{O}$ slope of 7.17 and 7.86, respectively (Table 2), implying the lower evaporative enrichment of the spring water. Besides the temperature-influenced

$\delta\text{D}-\delta^{18}\text{O}$ variations, the obtained slopes can be attributed to the contribution of groundwater, which tends to enhance the $\delta\text{D}-\delta^{18}\text{O}$ slope values (Dalai *et al.* 2002). In contrast, the precipitation samples collected during the pre-monsoon season (see Supplementary material, Fig. S2) are represented by relatively similar $\delta\text{D}-\delta^{18}\text{O}$ slopes irrespective of the elevation, as presented in Table 2. The above $\delta\text{D}-\delta^{18}\text{O}$ relationships for the rainwater of the region point towards the presence of evaporation from falling raindrops at all of the elevation transects.

Similarly, $\delta\text{D}-\delta^{18}\text{O}$ slopes for spring water during the monsoon period (Table 2; see Supplementary material, Fig. S2) are higher at all elevation regimes than those for pre-monsoon spring water. The higher slope indicates that relatively depleted water mixes with pre-monsoon spring water. The maximum variation is observed at the mid-altitude level, which is due to

recharge by isotopically depleted monsoonal rain. The lower slope compared to the GMWL (8.17) indicates evaporative water loss, possibly owing to the high atmospheric temperature. Furthermore, as in the pre-monsoon season, the spring water samples belonging to the elevation range of 1000–1500 m during the monsoon period were also characterized by slightly lower δD – $\delta^{18}O$ slopes (Table 2), implying the role of evaporation or groundwater contribution in spring water isotopic alteration. The δD – $\delta^{18}O$ relationship of the monsoon precipitation of the catchment (Table 2; see Supplementary material, Fig. S2) for different elevations is similar to the slope of the GMWL (8.17), indicating the preservation of original isotopic signatures by the falling raindrops.

Furthermore, the spring water samples collected during the winter season (Table 2; see Supplementary material, Fig. S2) attain significantly lower δD – $\delta^{18}O$ slopes from the entire catchment than are seen in the other two sampling periods. These lower slopes during the winter indicate the presence of a high magnitude of evaporation under dry and arid climatic conditions and/or significant contribution of recycled moisture, which is common in mountainous terrain (Dalai *et al.* 2002, Bershaw *et al.* 2012). The δD – $\delta^{18}O$ slopes of the winter precipitation (Table 2; see Supplementary material, Fig. S2) are much higher than those of the corresponding spring water samples, which indicates no or very little contribution of winter rain in spring water, and the samples collected from the elevation range of 1000–1500 m and >1500 m are characterized by δD – $\delta^{18}O$ slopes greater than that of the GMWL (8.17). In the Himalayan terrain, the significantly low temperature of the precipitation also causes an increase in δD – $\delta^{18}O$ slope values (Pande *et al.* 2000). The observed high slope values of precipitation can be explained by lower winter temperatures in the study region and/or the influence of western disturbances. This is supported by the high intercept values of the corresponding regression lines (Table 2).

5.3 Deuterium excess (*d*-excess) and hydrological processes

The obtained *d*-excess (Rozanski *et al.* 1993) values for the monsoon precipitation ($9.3\text{‰} \pm 2\text{‰}$) and spring water samples ($10.3\text{‰} \pm 1.5\text{‰}$) (Table 1) are similar to the earlier reported values for precipitation in the NW Himalaya ($\sim 8\text{‰} - 15\text{‰}$; Kumar *et al.* 2010, Maurya *et al.* 2011, Sharma *et al.* 2017), confirming the isotopic preservation potential of precipitation during the monsoon period. However, from the data, it is clear that the *d*-excess values for the pre-monsoon ($9.9\text{‰} \pm 4\text{‰}$) and monsoon precipitation ($9.3\text{‰} \pm 2\text{‰}$) (Tables 1 and 2), as well as spring water ($9.1\text{‰} \pm 1.7\text{‰}$ and $10.3\text{‰} \pm 1.5\text{‰}$, respectively), are within overlapping ranges and not distinctly distinguishable. These similar *d*-excess values of precipitation and spring water during both monsoon and pre-monsoon seasons imply the presence of regionally originated recycled moisture later, which has the potential to enrich the *d*-excess values (Dalai *et al.* 2002, Das and Rai 2022). In addition to the precipitation, the contribution from the groundwater of the region, having a *d*-excess value of $\sim 13\text{‰}$ ($15\text{‰} \pm 5\text{‰}$; Maurya *et al.* 2011), to the spring water can result in relatively higher *d*-excess.

The rainwater samples collected during the winter season have significantly higher *d*-excess values ($13.7\text{‰} \pm 5.4\text{‰}$) (Tables 1 and 2) in comparison to the values during the pre-monsoon ($9.9\text{‰} \pm 4\text{‰}$) (Tables 1 and 2) and monsoon periods ($9.3\text{‰} \pm 2\text{‰}$) (Tables 1 and 2). Additionally, as is evident from the lower δD – $\delta^{18}O$ slope during the winter, a significant magnitude of evaporation is altering the isotopic compositions of the spring water. However, evaporation normally reduces the *d*-excess of surface water. Instead, it is characterized by enriched *d*-excess ($9.7\text{‰} \pm 2.3\text{‰}$) (Tables 1 and 2), which is like that of the monsoon period. These higher *d*-excess values of the winter samples may possibly result from (i) westerlies, having *d*-excess values of $15\text{‰} - \sim 25\text{‰}$ (Rozanski *et al.* 1993, Pande *et al.* 2000, Maurya *et al.* 2011), (ii) the addition of groundwater (*d*-excess values of $\sim 13\text{‰}$ (Maurya *et al.* 2011) and (iii) the contribution of recycled moistures.

Similar to the δD – $\delta^{18}O$ slopes, the *d*-excess values of the spring water and precipitation also differ at various elevation transects. For the spring water samples collected during the pre-monsoon season, the *d*-excess values increase with increasing altitude, i.e. from $8.4\text{‰} \pm 1.5\text{‰}$ for <1000 m to $9.2\text{‰} \pm 1.7\text{‰}$ for 1000–1500 m to $9.9\text{‰} \pm 1.9\text{‰}$ for >1500 m sampling elevation (Table 2). This gradual increase of *d*-excess values can be attributed to (i) enhanced magnitude of westerly contribution, (ii) a higher input of recycled moisture through evaporation and/or evapotranspiration to the spring water budget in higher elevation regions, and (iii) a lesser degree of evaporative loss of water from springs. Previously, from the δD – $\delta^{18}O$ slopes (Table 2), it was evident that the magnitude of spring water evaporation is highest within the elevation transect of 1000–1500 m. Therefore, it must have resulted in the lowest *d*-excess values. In contrast, the samples from the elevated regions are characterized by relatively higher *d*-excess values in comparison to the lower elevation transect, indicating the contribution of continentally derived recycled moisture. This differential behaviour of δD – $\delta^{18}O$ slopes and *d*-excess values can be explained by the combined effect of evaporation and recycled moisture contribution which is common in mountainous terrain (Das and Rai 2022). However, from the current dataset it is difficult to isolate the extent of isotopic fractionation through both processes. Similar to the *d*-excess values of spring waters, the *d*-excess values of the precipitation also increase with increasing elevation (Table 2). The highest *d*-excess values of the precipitation at the highest elevation region too imply the presence of evaporation at lower reaches and/or enhanced contribution of recycled moisture at higher elevation regions.

The *d*-excess values of the precipitation and spring water collected during the monsoon season at the basin scale are similar to each other and within the error ranges of the reported *d*-excess values of the ISM precipitation, implying the preservation of meteorological isotopic signatures by the springs. However, gradually increasing *d*-excess values with elevation (Table 2) gesture towards the varied magnitude of evaporation and/or recycled moisture contribution to the spring water. A close examination of Table 2 indicates that the monsoon precipitation samples have slightly lower *d*-excess values compared to corresponding spring water,

owing to the mixing of meteorologically fresh water and previously existing spring water or groundwater in the latter.

Interestingly, the d -excess compositions of the spring water and precipitation samples during the winter season show the greatest discrepancies (Table 2). The observed incongruity between the d -excess values (also the δD - $\delta^{18}O$ slopes; Table 2) can be explained through the ongoing hydrological processes in the spring water as well as the moisture source of the precipitation. As discussed earlier, the NW sector of the Himalayas generally receives the wintertime precipitation from a westerly source, having enriched d -excess values (~15–25‰, Rozanski *et al.* 1993, Pande *et al.* 2000, Maurya *et al.* 2011). Meanwhile, spring water during the same season normally suffers from evaporation under cold and arid conditions and receives soil moisture from greater depths through groundwater flow, which tends to deplete the d -excess values. The gradual increase in d -excess with elevation for both springs and precipitation indicates lower evaporation rates at higher altitudes and higher rates at lower elevations.

5.4 Isotopic variability with altitude

In the Himalayan region, the orographically induced rainfall results in the gradual depletion of heavier isotope ($\delta^{18}O$) in precipitation with increasing elevation of the sampling locations (Yurtsever and Gat 1981, Clark and Fritz 1997, Dalai *et al.* 2002), a phenomenon referred to as the “altitude effect.” This altitude effect in different Himalayan streams varies from 0.09‰/100 m to 0.29‰/100 m (Ramesh and Sarin 1992, Bartarya *et al.* 1995, Garzione *et al.* 2000a, 2000b, Pande *et al.* 2000, Dalai *et al.* 2002), with a global average of 0.21‰/100 m (Chamberlain and Poage 2000). The altitude- $\delta^{18}O$ relationship for the rain and spring water samples of the Takoli Gad catchment (Fig. 3(c,d)) for the monsoon period are estimated at 0.3‰/100 m for precipitation, and 0.06‰/100 m for spring (Fig. 3(c,d); Table 3). This obtained altitude effect from the spring water is analogous to the effect for the Yamuna basin (0.09‰, Dalai *et al.* 2002) and Indus basin (0.09‰, Pande *et al.* 2000) and lower than the altitude effect reported for the Ganga basin (0.19‰, Ramesh and Sarin 1992), Gaula catchment (0.14‰; Ramesh and Sarin 1992) and Seti as well as Kali-Gandaki basins (0.29‰, Garzione *et al.* 2000a, 2000b). Another noteworthy feature of our altitude effect estimation is the ~5 times higher values in precipitation compared to the spring water. This observed discrepancy can be attributed to the higher contribution of water to the springs from various elevation ranges of soil moisture content and moisture sources (groundwater as well as other lower-order streamlets) (Dalai *et al.* 2002; see Supplementary material, Figs S3–S6) at lower altitudes, whereas the precipitation at a location is believed to represent the original isotopic signatures. Additionally, an earlier study on the Yamuna basin (Dalai *et al.* 2002) found an insignificant impact of evaporation on the altitude effect. To validate this in our catchment, the obtained altitude- $\delta^{18}O$ relationship for rain and spring water samples during pre-monsoon as well as winter seasons (when the spring and precipitation samples are supposed to suffer the evaporation process) are estimated as 0.07 ± 0.0012 for rain and -0.05 ± 0.0001 for spring water during the pre-monsoon

period, and -0.23 ± 0.0005 for rain and -0.04 ± 0.0001 for spring water during the winter season (Fig. 3(c,d); Table 3), reflecting the varying altitude effect owing to evaporation. Interestingly, the altitude effect associated with the precipitation shows an enrichment of $\delta^{18}O$ with increasing elevation, which is highly unusual and can be explained in terms of evaporative enrichment of precipitation in high-altitude isotopes under progressive dry and arid atmospheric conditions. However, a detailed study is needed for a better understanding of this process. In contrast, our study (Table 3) suggests that evaporation has an insignificant impact on the altitude effect as derived from the spring water samples, similar to the conclusion of Dalai *et al.* (2002). As indicated earlier, the average altitude effect in different Himalayan catchments varies from 0.09‰/100 m to 0.29‰/100 m without change in the moisture source(s) to the precipitation. Hence, under the assumption that the observed altitude effect is the manifestation of the rainout process, we have estimated the rainout percentage in the Takoli Gad catchment, which places the moisture removal percentage from the cloud as $32 \pm 4\%$; this is within the range of other Himalayan River basins.

Furthermore, the relative contribution of precipitation during different seasons to the water budget of the Takoli Gad catchment has been estimated as $8 \pm 14\%$, $82 \pm 21\%$, and $10 \pm 18\%$ for the pre-monsoon, monsoon, and winter seasons, respectively (Fig. 5). The estimation indicates that the monsoon precipitation contributes a significantly dominant amount (~82%) to the spring water budget of the catchment. This dominance of the monsoon contribution can be well understood from the fact that the catchment experiences the direct influence of the ISM. Similarly, the estimation indicates that ~90% of the stream discharge during winter and summer is contributed from the groundwater or rainfall of the preceding monsoon, possibly due to interflow. However, the above estimation may be the upper limit, as it assumes a minimal role of other hydrological processes like evaporation, evapotranspiration, etc. in the catchment.

5.5 Spring recharge mechanism and pattern

In addition to contribution estimation, this study has tried to delineate the spring recharge mechanism using available hydro-meteorological data (soil moisture, overland flow, base-flow and relative humidity and potential evaporation data) derived from remote sensing dataset (GLDAS and AIRX3STM models). Soil moisture data are critical for understanding the seasonal variability of infiltration processes within a watershed. When soils are initially dry, their infiltration capacity is high; however, as soil moisture content increases, infiltration capacity gradually declines. When infiltration capacity exceeds rainfall intensity, overland flow or surface runoff is minimized. Soil moisture thus represents an integral component of the catchment water cycle. In this catchment, during the monsoon months of July to September, heavy precipitation increases soil moisture across all depths (Table 4; see Supplementary material, Figs S3b, S4b, S5b, S6b), initiating significant infiltration that recharges underground aquifers. Although intense rainfall events generate overland flow, this diminishes quickly as the monsoon

retreats from the region after September (Fig. 6(a)). Our isotopic end-member contribution calculation indicates that nearly 82% of spring discharge originates from monsoon infiltration, aided by fractures and conduits in the geological matrix. The effect of this recharge persists until October, as shown by high baseflow (Fig. 6(b)) despite the monsoon retreat, reflecting sustained aquifer replenishment. In winter, westerly-derived precipitation contributes to soil moisture throughout the depth (Table 4; see Supplementary material, Figs S3c, S4c, S5c, S6c), strengthening groundwater storage and maintaining spring discharge even during drier conditions. Isotopic signatures, characterized by high d -excess values and lower slopes (Table 2), confirm contributions from both groundwater and winter precipitation. Our isotope-based end-member calculation shows that 10% of spring discharge originates from winter infiltration, despite lower moisture content in winter than in monsoon months at all depths. Similarly, the pre-monsoon season is marked by relatively low soil moisture across depths compared to the monsoon (see Supplementary material; Figs S3a, S4a, S5a, S6a), yet springs continue to flow due to the persistence of baseflow sustained by earlier recharge. Only 8% of spring discharge is contributed by pre-monsoon precipitation infiltration, as indicated by isotopic end-member calculation. Lower moisture content and less recharge from the pre-monsoon season further led to lower discharge and seasonal drying in some springs in this catchment. This sequence highlights a continuous hydrological process in which monsoon infiltration provides the primary recharge, winter precipitation supplements and enhances groundwater contributions, and pre-monsoon discharge is maintained by stored subsurface water, together ensuring perennial spring flow across seasons. Together, these observations provide valuable insights into catchment-scale hydrological processes and strengthen our isotopic interpretation.

5.6 Spring water residence time

The tritium concentration, an excellent tracer of water residence time and depth of water circulation (Clark and Fritz 1997), of the Takoli Gad region distributes the groundwater into six distinct age brackets (Fig. 4): (i) recent, (ii) 3–5 years, (iii) 8–10 years, (iv) 16–18 years, (v) 20–25 years, (vi) above 30 years. Using these categories, 41% of the springs have groundwater older than 30 years, making them less vulnerable to annual rainfall variability and thus reliable for year-round use. In contrast, 17% of the springs are recharged by the recent monsoon with a groundwater age of 1–2 years, reflecting their dependence on seasonal precipitation and higher susceptibility to water scarcity during dry periods. The intermediate age categories likely result from the mixing of these two major groundwater systems (Fig. 4(b)).

Interestingly, the relationship between groundwater age distribution and sampling elevation in the Takoli Gad catchment shows no significant correlation (Fig. 4(a)), suggesting that groundwater preservation potential is independent of altitude. This is further evidenced by the uneven age distribution across the catchment, particularly in mid-altitude regions, where multiple interconnected aquifers contribute to complex recharge dynamics. Understanding these patterns can provide

spatiotemporal insights into spring behaviour and water-age variability.

The schematic diagram in Fig. 4(b) illustrates that springs with older tritium ages are primarily recharged through deep infiltration pathways, aided by extensive forest root networks and soil moisture retention. These conditions allow water to slowly percolate into the aquifers over time. Conversely, springs with younger tritium ages are predominantly influenced by rapid shallow infiltration and overland flow, leading to quicker discharge. Therefore, the depth and rate of water infiltration, governed by the eco-hydrological characteristics of the landscape, are the key factors shaping the tritium age patterns observed across the altitudinal gradient. These findings align with previous studies conducted in the Himalayas (Shivanna *et al.* 2008, Jeelani *et al.* 2015, Shah *et al.* 2017, Bhat and Jeelani 2018, Thakur *et al.* 2020, Chatterjee *et al.* 2023, Kumar *et al.* 2024).

6 Conclusions

Recently, the alarming issue of water scarcity in the Himalayan region has prompted researchers and policymakers to investigate the hydrological processes to meet the water demands of local populations while considering future sustainability. This study investigates the hydro-meteorological processes of the Takoli Gad watershed using stable water isotopes (δD and $\delta^{18}O$) and radioisotopes (^3H) collected from spring and rain-water across different seasons (pre-monsoon, monsoon, and winter). The δD and $\delta^{18}O$ compositions imply a critical role of elevation in the presence and impact of various hydrological processes on the spring water. The source contribution estimation delineates the important influence of ISM on the water budget of the watershed, which is important to properly understand under the current global climate change scenario. The study has further combined the measured isotopic compositions along with the hydro-meteorological data (derived from remote sensing dataset) to comprehend the hydrological behaviour of the region in a simpler way. The groundwater-based flow plays a crucial role in spring recharge across different seasons in this watershed. During the southwest monsoon, particularly in August, the contribution to spring discharge is at its maximum. In the winter season, especially in October, there is also significant spring discharge. However, during the pre-monsoon season, the contribution of baseflow groundwater to spring recharge is minimal. The catchment area is primarily influenced by the ISM, which delivers moisture from June to September. During this period, the moisture seeps into the soil, reaching deeper layers through the root zone, aided by the soil's permeability. This process contributes to groundwater baseflow and gradually replenishes the springs during winter, while only limited recharge takes place in the pre-monsoon season.

The key highlights and factors regulating the isotopic composition of springs and precipitation are summarized as follows:

- (i) The deviation of $\delta^{18}O$ and δD slopes from the GMWL in pre-monsoon precipitation samples shows a lower slope. This is due to high potential evaporation rates, which cause falling raindrops to evaporate, and low

relative humidity in the region, which further favours the lower slope. The notably gentle slope of the δD – $\delta^{18}O$ relation for spring water during winter, in contrast to other seasons, highlights several hydrological processes occurring within the catchment area. During the monsoon season, springs experience a significant contribution from direct precipitation ($82 \pm 21\%$), while winter brings additional moisture from the westerlies. This precipitation infiltrates the soil at varying depths and elevations, allowing for extended evaporation over time. As a result, the increased interaction under lower humidity conditions contributes to the observed lower slope of spring water in winter.

- (ii) The $\delta^{18}O$ altitude effect ($0.06\text{‰}/100\text{ m}$) corresponds with findings in other Himalayan catchments. However, in precipitation, this effect is approximately 5 times greater ($0.3\text{‰}/100\text{ m}$) compared to spring water ($0.06\text{‰}/100\text{ m}$).
- (iii) The $\delta^{18}O$ – δD slopes and d -excess values of the samples at different elevation transects of the study area seem to be affected by the differential magnitude of various hydrological processes. During the pre-monsoon and monsoon seasons, the spring water seems to undergo evaporative enrichment, with the highest magnitude at elevations of 1000–1500 m.
- (iv) Tritium concentration data indicates that 41% of the springs are over 30 years old. These springs are less sensitive to water shortages due to annual rainfall variations, longer subsurface flow paths, and greater storage, allowing sustained year-round discharge. In forested catchments, this is commonly associated with enhanced recharge regulation through increased infiltration and storage. Meanwhile, 17% of the springs are 1–2 years old, reflecting recent monsoon activity.

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Disclosure statement

No potential conflict of interest was reported by the authors.

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