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On the emergent scale of bedrock groundwater contribution to headwater mountain streams

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We investigated the contribution of bedrock groundwater to streamflow as a function of catchment scale in a headwater stream. Synoptic surveys were conducted during hydrologically important periods of the year using multiple environmental tracers in stream water, soil water, and bedrock groundwater, along a first-order montane stream, in west-central Montana. Sampled analytes included ²²²Rn, used to constrain total subsurface flux, and major and minor elements, used in end-member mixing analysis (EMMA) to identify the contributions of soil and bedrock groundwater to the stream. Partitioning between soil-derived and bedrock-derived groundwater was then analyzed as a function of the incremental and accumulated sub-catchment sizes. Radon results indicated that subsurface water contributions accounted for the majority of streamflow at all surveyed times. EMMA results revealed that the bedrock groundwater contribution to streamflow varied between 26% during peak snowmelt and 44% during late summer. Streamflow generation was dominated by soil groundwater contribution along the entire reach, but the bedrock groundwater contribution increased consistently with accumulated sub-catchment size. However, groundwater contributions were not well-correlated with incremental sub-catchment size. The scale at which increased bedrock groundwater discharge can be correlated with sub-catchment size appears to be >1 km² for our study. Our results are consistent with a conceptual model where streamflow is predominantly generated by a 3D subsurface nested flow system. Local subsurface heterogeneities control the stream source at local scales but begin to average out at scales >2 km². Our study indicates that, while soil groundwater is the dominant source, bedrock groundwater remains an important and predictable contributor to streamflow throughout the year, even in a snow-dominated, mountainous headwater catchment.

KEYWORDS

groundwater—surface water interactions, streamflow generation, mountain hydrology, environmental tracers, mountain aquifer

Introduction

Upland mountain watersheds host active hydrologic systems that play important roles in the storage and release of water to the larger stream network (Hale et al., 2016; Hale and McDonnell, 2016; Liu et al., 2004; Williams et al., 2015). These groundwater systems are locally important sources of streamflow (Asano et al., 2022; Gabrielli et al., 2012; Uchida et al., 2003) and recharge regional groundwater systems, which discharge to larger streams

in the lowlands (Beisner et al., 2018; Fan et al., 2013; Frisbee et al., 2011; Gardner et al., 2011; Gleeson and Manning, 2008; Smerdon et al., 2012). The amount of bedrock groundwater discharge in montane streamflow is important for conceptualizing watershed function and predicting hydrologic resource availability in the future (Condon and Maxwell, 2019; Meixner et al., 2016; Williams et al., 2015). However, groundwater flow systems are marked by extreme heterogeneity in hydraulic characteristics and the resulting flow path length and rate; as a result, groundwater discharge volumes and locations are all highly variable (Beisner et al., 2018; Gardner et al., 2011; Hale et al., 2016; Harrington et al., 2013; Smerdon and Gardner, 2021). Neither the spatial scale at which 3D groundwater flow becomes an important and predictable source of streamflow nor the seasonal variation in these processes are well-documented at present.

Subsurface water is an important source of streamflow, even in snow-dominated watersheds (Liu et al., 2004). Subsurface sources are commonly conceptualized as originating from both bedrock (weathered bedrock and saprolite) and soil reservoirs (Anderson et al., 1997a; Carroll and Williams, 2018; Hale et al., 2016; Hale and McDonnell, 2016; Salve et al., 2012; Torres et al., 1998). In this study, we refer to shallow groundwater in the soil zone as soil groundwater and deeper, fractured-bedrock groundwater as bedrock groundwater. The relative importance of soil and bedrock reservoirs depends on the soil and bedrock hydraulic characteristics (Flint et al., 2008; Gardner et al., 2020; Katsuyama et al., 2005; Pfister et al., 2017; Uchida et al., 2002, 2006). The amount of groundwater in a watershed significantly influences dynamic properties, such as hydrograph variability (Asano et al., 2022; Gardner et al., 2010; Pfister et al., 2017) and mean transit time (Asano et al., 2002; Asano and Uchida, 2012; Hale et al., 2016; Pfister et al., 2017; Uchida et al., 2006).

Soil water lateral flow and bedrock groundwater circulation are tightly coupled (Appels et al., 2015; Carroll et al., 2019; Gardner et al., 2020; Tromp-van Meerveld et al., 2007; Uchida et al., 2002). As a result of this coupling, the location and amount of bedrock recharge are controlled by subsurface permeability distribution, hillslope position, and bedrock topography, which affect the lateral soil moisture distribution and location of vertical drainage (Appels et al., 2015; Gardner et al., 2020). The relative contributions of soil and bedrock groundwater to stream water are expected to vary temporally (Spencer et al., 2021). The celerity of subsurface response depends on the soil and bedrock hydraulic properties (Anderson et al., 1997a,b).

Groundwater discharge location, rate, and flow path source can be highly localized and variable because they depend upon the local geology and surface characteristics (Beisner et al., 2018; Bergstrom et al., 2016; Gardner et al., 2011; Harrington et al., 2013; Smerdon et al., 2012). In addition, the localized bedrock and soil water contributions from hillslopes are buffered by the local alluvial groundwater system, and they can be mixed and transported within this system before reaching the stream (Jencso et al., 2010). Despite the spatial heterogeneity at local scales, it is clear that the contribution from deeper bedrock groundwater systematically increases with increasing drainage area at regional scales (Asano et al., 2022; Frisbee et al., 2011, 2012). Since there is observed heterogeneity in the amount and location of groundwater discharge occurring locally, which contrasts with predictable increases in

groundwater discharge at larger, integrated scales, it appears that groundwater discharge in streams is an emergent property.

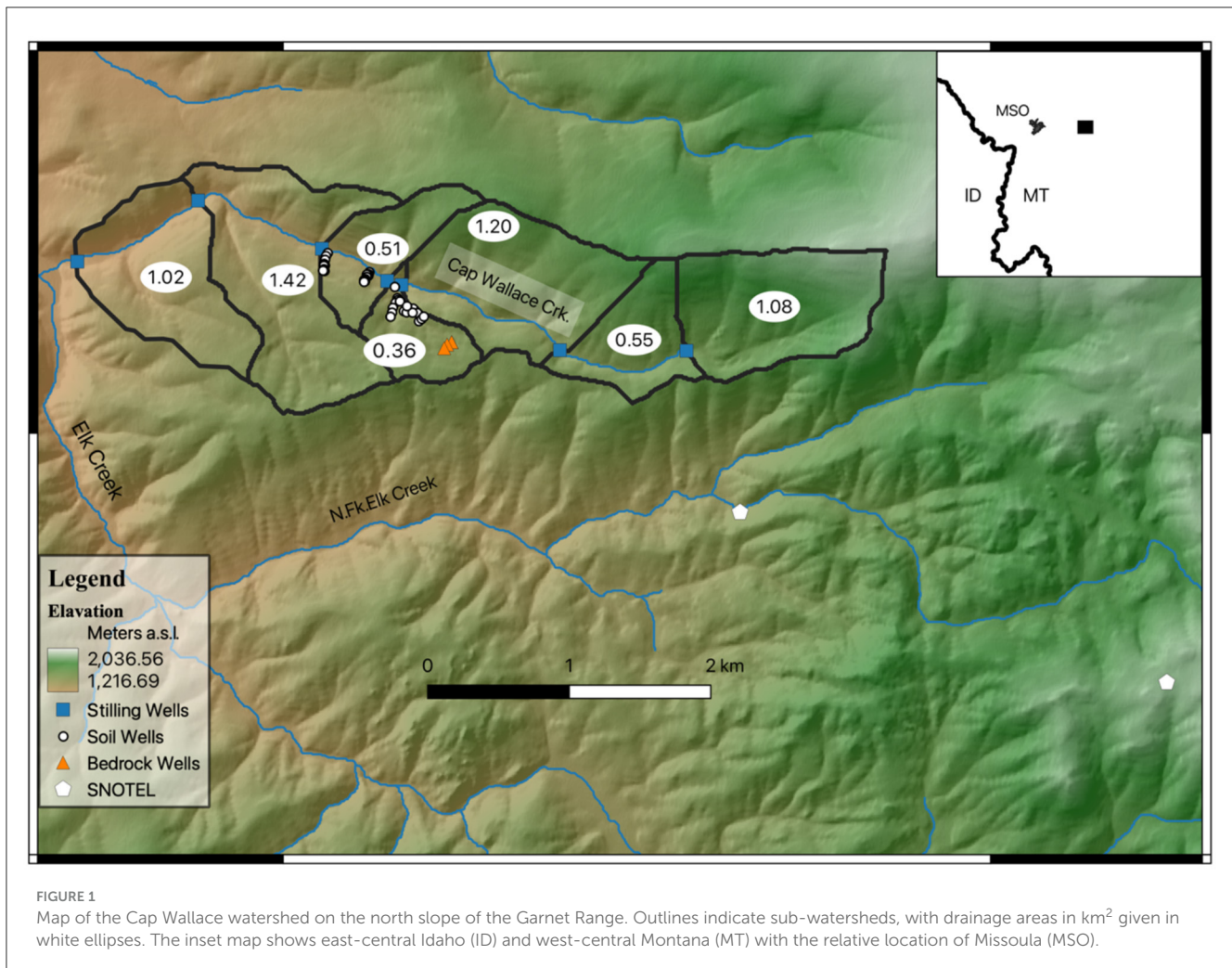
Is there, in fact, a scale at which groundwater discharge volume becomes more predictable? If so, does this scale change temporally? In this article, we investigate how the amount and location of bedrock and soil groundwater discharge vary as a function of space and time in a montane, first-order watershed in west-central Montana. The study objectives are to constrain the amount of bedrock vs. soil groundwater discharge along a stream reach as a function of seasonal changes in catchment wetness and increasing spatial scale. Leveraging a high-resolution network of gauging stations, we conducted a multi-tracer interrogation of the incremental sources of streamflow generation. We estimate the total volume of subsurface discharge and distinguish the contributions of soil and bedrock groundwater discharge along the stream. We then explore relationships of the bedrock groundwater contribution as a function of the incremental and integrated upland characteristics.

Study area

The Cap Wallace Watershed (CWW) is located in west-central Montana, ~56 km northeast of Missoula on the north slope of the Garnet Range (Figure 1). The catchment drains 6 km² of forested land and is a tributary of the Blackfoot River. The CWW ranges in elevation from 1,163 m at the confluence of Cap Wallace Creek (CWC) and Elk Creek to 1,918 m at the ridge tops. Vegetation consists primarily of second-growth western larch (*Larix occidentalis*) on the north-slope mid and upper elevations, Douglas-fir (*Pseudotsuga menziesii*) on north-facing, low-elevation slopes, and Ponderosa Pine (*Pinus ponderosa*) on south-aspect slopes. North-aspect slopes have well-developed soils, whereas south-aspect slopes lack soil coverage and are generally covered by steep talus fields. Due to topographic variation in the radiation balance and resulting microclimatic effects (Holden et al., 2011a,b), south-aspect slopes remain significantly drier throughout the year (Hoylman et al., 2019).

At the Lubrecht Flume SNOTEL station (site #604; 1,425 m), the average annual precipitation over the record period of 10 years is 514 mm. Nearly 50% of precipitation falls as snow, classifying the Cap Wallace as a snowmelt-dominated watershed. In 2017, temperatures ranged from −27.6°C to 34.2°C with an annual average temperature of 6.1°C (Figure 2). Peak snow water equivalent typically occurs in late March, with snow melt taking place during April, May, and early June. May and June are typically cool and rainy, while July, August, and September are generally hot and dry, with precipitation beginning again in October.

The Garnet Range is part of the Northern Rocky Mountains and was formed through folding and thrusting associated with the Jurassic Sevier and Cretaceous Laramide Orogenies (140–55 Ma). Subsequent extensional normal faulting was initiated in the early Eocene as a result of Basin and Range tectonics, triggering the uplift of metamorphic core complexes in the Northern Rockies (Portner et al., 2011). The CWW is situated on the flank of a large granitic batholith called the Garnet Stock, which intruded the area during the Late Cretaceous (Lonn et al., 2010). CWC follows an east–west trending normal fault. Quaternary glacial



and alluvial deposits cover the narrow valley bottom and stream floor, while the hillslopes are composed of gravelly, silt-loam soils (Soil Survey Staff, Natural Resources Conservation Service, United States Department of Agriculture, n.d.) that overlie several kilometers of argillites, siltites, and quartzites of the Precambrian metasedimentary Belt Supergroup. Soils are classified as Typic Haplustalfs (well-draining with a lower clay horizon), with depths ranging from 0.5 to 2 m that are thicker in hollows and thinner in upper elevation ridges (Hoylman et al., 2018). The north-aspect slope is composed of the Garnet Range Formation, a siltstone with hummocky sand lenses, which lies unconformably across the valley from the stratigraphically lower Bonner Quartzite, a massive, slightly metamorphosed, arkosic sandstone that makes up the south-aspect slopes (Lonn et al., 2010). These formations have been subject to tectonism associated with compression from major mountain-building events and subsequent active extension due to gravitational relaxation, leading to a dense fracture network within the bedrock (Brenner, 1968; Gardner et al., 2020).

Methods

We measured multiple environmental tracers in soil, groundwater, and stream water to perform hydrograph separation

along CWC. We conducted synoptic sampling along CWC for discharge of the stream and dissolved ²²²Rn and major ions at four hydrologically important times: (1) early-spring baseflow (March), (2) the rising limb of spring snowmelt freshet (April), (3) the falling limb of freshet (May–early June), and (4) summer baseflow (late June–October).

Field methods

Cap Wallace Watershed is instrumented with 36 shallow soil wells, three groundwater wells, and seven stilling wells for measuring stream discharge. Shallow soil wells are 1–2 m deep, fully screened, and pounded with a steel rod until refusal. Bedrock groundwater wells were drilled to depths of 10–20 m using a man-portable drilling rig. Bedrock wells were fully screened, and the annulus was filled with the sand pack from the bottom to 7 m below the land surface and then cased and the annulus sealed with bentonite to the land surface. Stilling wells were spaced at ~1 km intervals, strategically located downstream of converging sub-watersheds with large contributing areas to estimate the hillslope contribution from sub-watersheds along the stream (Figure 1). Field data collection began during the winter of 2017. Solinst pressure transducers were installed in stilling, soil,

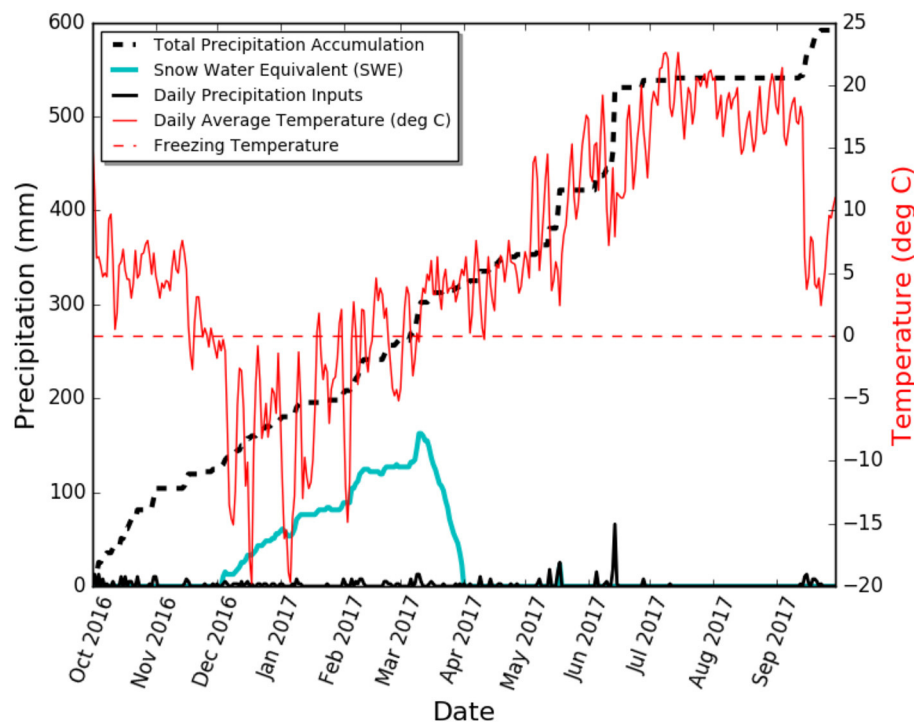


FIGURE 2

Lubrecht climate data for the 2017 water year. Data source: Lubrecht SNOTEL (site #604). 2017 was a heavy precipitation year with many days below freezing.

and groundwater wells to continually monitor and record stage measurements along the stream and contributing hillslopes.

Throughout the spring, summer, and fall of 2017, synoptic discharge measurements were made repeatedly at each stilling well using dilution gauging (Gooseff and McGlynn, 2005). Replicate measurements were collected at multiple stilling well locations during each sampling period. On average, replicates of stream discharge differed by $\pm 4\%$ of one another, but during baseflow, one replicate recorded a discharge value of 54% lower than the original measurement (28 August 2017). There are several potential reasons for this discrepancy. In particular, this measurement was collected at one of the lowest recorded flows. Low flows have been shown to complicate the dilution gauging technique when hyporheic exchange occurs (Moore, 2004). Because this replicate was identified as an outlier, this baseflow measurement was not used in our determination of the average uncertainty in replicate measurements (5%); however, it is clear that as we approach low flows, our ability to accurately measure the flow is reduced.

During synoptic discharge measurement campaigns, each stilling well location was sampled for ^{222}Rn and major ions (Ca^{2+} , K^+ , Mg^{2+} , Na^+ , F^- , Cl^- , NO_3^- , SO_4^{2-}). Sample bottles were rinsed three times with sample water. Samples for ^{222}Rn analysis were collected underwater in airtight 250 mL glass bottles to avoid exchange with the atmosphere. Major ion samples were filtered using a 0.45- μm disposable filter and stored in acid-washed polyethylene bottles. Cation samples were acidified with 0.2 mL of nitric acid. Field parameters, including pH, temperature, oxidation-reduction potential, and electric conductivity, were collected using

an *In-Situ* Aqua TROLL 600 sonde. Precipitation samples were collected at the Jones Pond Montana Mesonet site ~ 10 km away and were measured throughout the year. Integrated snowmelt samples were collected during the melt season using a snow core to collect complete snowpack samples and melted in the laboratory refrigerator. Water samples were collected from soil and groundwater wells, using a peristaltic pump in April 2017. Prior to collection, wells were purged until field parameters remained constant ($\sim 2\text{--}3\times$ the volume of the well).

Laboratory methods

^{222}Rn samples were analyzed within 48 h of collection using a Durrige Instruments RAD7 spectral alpha-decay detector. A sample of 250 ml was analyzed using the WAT-250 protocol, with 5 min of degassing (bubbling) followed by five rounds of counting, where each counting round was 5-min long. The RAD7 was purged for >20 min between each sample. Concentration was determined from the last four counting rounds (20 min total). The obtained concentrations were corrected to account for decay during the period between sample collection and analysis using the formula:

$$C_{tc} = C_i e^{\left(\frac{t}{132.4}\right)} \quad (1)$$

where C_{tc} is the time-corrected radon concentration in Bq/L, C_i is the initial radon concentration prior to correction (Bq/L),

and t is the time elapsed between collection and analysis in hours. Samples for the analysis of anions were run on a DIONEX DX500 (IC) and cations on a Perkin-Elmer OPTIMA 5300 (ICP-OES). Detection limits and accuracy for IC and ICP-OES analyses are presented in Table 1. All ^{222}Rn and major ion samples were analyzed at the University of Montana's Environmental Biogeochemical Laboratory.

Modeling approach and data analysis

Two different modeling techniques were used to identify and estimate the sources of streamflow. First, in-stream radon concentrations were used to estimate total subsurface discharge. End-member mixing analysis (EMMA) was then conducted on major ions in stream, soil, snowmelt, and groundwater to estimate the appropriate number and type of end members. The obtained end members were then used in a mass-balance mixing model to quantify end-member contributions to the stream. Mixing fractions were compared with landscape topographic characteristics. The effects of different internal landscape characteristics on spatial heterogeneity and duration of subsurface flow were quantified seasonally at spatial scales ranging from 0.4 to 6 km².

Radon modeling

^{222}Rn is advantageous as an analyte because it is not present in the atmosphere and is non-existent in overland flow, direct snowmelt, and precipitation. Meanwhile, all subsurface waters with a residence time of over a few hours have measurable ^{222}Rn , and all waters with a residence time over ~ 2 weeks achieve secular equilibrium and have relatively constant ^{222}Rn concentration for a given lithology. Thus, ^{222}Rn concentrations can be used to isolate subsurface streamflow sources. Estimating subsurface inflow from ^{222}Rn requires a model that accounts for subsurface discharge, gas exchange with the atmosphere, and radioactive decay. A 1D stream transport model was created to simulate longitudinal radon activity (Cook et al., 2003) using the equations:

$$\frac{\partial Q}{\partial x} = I(x) - L(x) - E(x) \quad (2)$$

$$Q \frac{\partial c}{\partial x} = I(c_i - c) + wEc - kwc - dw\lambda c \quad (3)$$

where c is the concentration of radon in the stream (Bq/L), c_i is the concentration in bedrock or soil groundwater (Bq/L), I is the groundwater inflow rate (m³/m/day), w is the stream width (m), d is the mean stream depth (cross-sectional area/width) (m), k is the gas exchange velocity (m/day), λ is the radon decay coefficient (per day), Q is the stream discharge (m³/day), E is the evaporation rate (m/day), and L is the stream extraction rate from pumping or diversions (m³/m/day). Equation 2 is the mass balance equation for water in the stream, and Equation 3 is the 1D steady-state equation for advective solute transport in the stream with groundwater inflow, atmospheric gas exchange, and in-stream radioactive decay. We assumed an atmospheric ^{222}Rn

concentration of zero, no production of ^{222}Rn within the stream, and steady-state flow conditions.

The ultimate goal of the ^{222}Rn fitting was to assess the amount of subsurface-derived discharge vs. overland discharge, assuming that soil and groundwater with residence times over 2 weeks have similar radon concentrations and overland flow has zero radon concentration. The distributed total groundwater discharge (soil plus bedrock) was estimated by fitting the observed ^{222}Rn stream discharge. Radon and discharge values were fit by varying stepwise groundwater discharge (Gardner et al., 2011). Groundwater inflow was assumed to be constant over 500 m steps, which resulted in 10 unknown groundwater inflow steps along the stream reach. The number and location of groundwater steps were held constant at all sampling times. Steps in the radon concentration correspond to boundaries of the stepwise constant groundwater inflow. The total number of observations for inversion is 14, resulting in an overdetermined system.

To quantify incremental gains from subsurface sources at each stilling well, we calculated the ratio of subsurface discharge to total streamflow gain (F_{ss}):

$$F_{ss} = \frac{Q_{ss1} - Q_{ss2}}{\Delta Q_t} \quad (4)$$

where Q_{ss1} is the upstream modeled subsurface discharge (L/s), Q_{ss2} is the downstream modeled subsurface discharge (L/s), and ΔQ_t is the measured difference in streamflow from upstream to downstream (L/s).

The parameters used for radon modeling are shown in Table 2. Those that were not directly measured in the field were estimated based on a series of equations also detailed in Table 2. Following the approach of Cook et al. (2003), a sensitivity analysis was conducted by varying each individual model parameter by $\pm 50\%$, while holding the other parameters constant to observe the associated effect on estimated subsurface inflow. Particular interest was given to the parameters not measured in the field since they were the largest source of uncertainty in the model.

As the gas transfer velocity (k) is difficult to constrain in low-order mountainous streams due to their variable geometries, flow velocities, and temperatures, special attention was given to the effect of k on estimated subsurface inflow rates. Four common equations that depend on slope, velocity, and depth (Raymond et al., 2012) were used to approximate the gas transfer velocity range.

End-member mixing analysis

End-member mixing analysis was performed following the approach of Hooper (2003). Details of the specific application of EMMA in this article can be found in Supplementary material. Here, we used four tracers (calcium, magnesium, sodium, and chloride, see Results) to constrain a two-end-member mixing model. The end members chosen were the reprojected average bedrock and soil groundwater compositions from all samples throughout the year. End-member concentrations were held constant over time. However, most bedrock and all soil groundwater samples were collected in spring when water tables

TABLE 1 Limits of detection and precision of field and laboratory duplicates for ion chromatography (IC) and inductively coupled plasma emission spectroscopy (ICAP-OES) analyses.

Parameter	Ca	K	Mg	Na	F	Cl	SO ₄	NO ₃
Limit of detection (mg/L)	0.102	0.51	0.102	0.51	0.015	0.100	0.100	0.100
Error % (lab)	±1.35	±1.40	±1.20	±1.12	±3.86	±2.31	±1.78	±0.21
Error % (field)	±3.20	±1.80	±3.90	±3.30	±6.57	±1.86	±0.19	±5.14

Errors associated with field and laboratory duplicates are within acceptable range for geochemical interpretation.

TABLE 2 Details of the parameters used in ²²²Rn modeling.

Parameter	Units	Value	Method of collection or estimation
In-stream radon concentration (<i>c</i>)	Bq/L	Avg. 0.907; 0.074–4.64 ±15%	Measured in field and analyzed using the RAD7 Alpha Decay Spectrometer at the University of Montana; spatially and temporally variable
Total river length (<i>x</i>)	km	5	Measured in field
Evaporation rate (<i>E</i>)	mm/day	0	Assumed negligible
Initial radon concentration (<i>c</i>)	Bq/L	Avg. 0.37; 0.12–0.60 ±15%	Concentration of most upstream sampling location (variable); measured in field
Atmospheric radon concentration	Bq/L	0	
Radon decay coefficient (<i>λ</i>)	/day	0.18	Constant
Gas transfer velocity (<i>k</i>)	m/day	Avg. 16.6; 5.5–27.7	Estimated from the study of Raymond et al. (2012)
Groundwater inflow Concentrations (<i>c_i</i>)	Bq/L	32.0	Highest measured concentration in groundwater samples; measured in field and analyzed using the RAD7 Alpha Decay Spectrometer at the University of Montana
Stream width (<i>w</i>)	m	0.75	Mean value of widths measured in field
Stream depth (<i>d</i>)	m	0.15	Mean value of depths measured in field

were shallow enough to be sampled. Our end members are referred to here as soil groundwater and bedrock groundwater, where soil water represents shallow groundwater in the soil zone, and groundwater represents bedrock groundwater in fractured bedrock. We estimated uncertainty for the EMMA results by propagating the error due to EMMA misfit through the mixing calculation details, which can be found in [Supplementary material](#). Error bars on mixing plots are given as the mean relative error $\frac{\delta Q}{Q}$ for all samples, which was 13%.

Terrain analysis

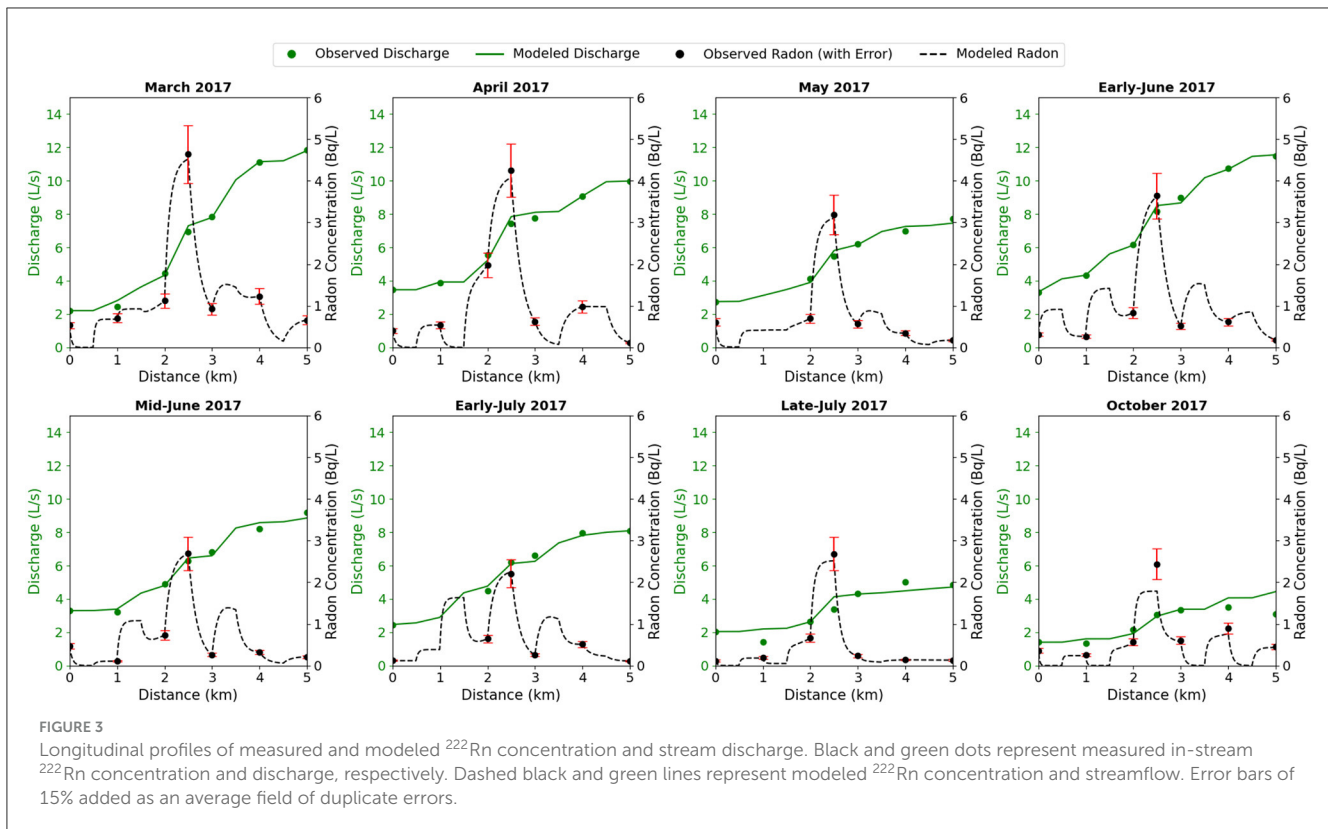
To explore the topographic characteristics that influence streamflow generation, the catchment was delineated into “sub-catchments.” All-terrain analyses were performed using the SAGA GIS software ([Conrad et al., 2015](#)). A 10-m resolution digital elevation model of the study area had sinks filled, and the stream network burned in. The CWW boundary was delineated by creating flow direction and accumulation grids to determine the contributing area for each stilling well. The watershed was then divided into seven sub-catchments based on those areas. In the process, each sequential downstream sub-catchment integrated all upstream sub-catchments; we refer to these sub-catchments as “accumulated” sub-catchments. “Incremental” sub-segments contributing between stilling wells were determined by subtracting the upstream sub-watershed from the downstream sub-watershed ([Figure 1](#)). Mean elevation, Topographic Wetness Index (TWI),

and upslope accumulated catchment area (UAA) were calculated for each accumulated and incremental sub-catchment. Linear regression was then performed to assess the modeled end-member discharge correlation to terrain attributes for both the incremental and accumulated areas.

To analyze the relationship between landscape and end-member discharge in an incremental sense, fractions of end-member discharge gained in each sub-catchment were calculated by quantifying the solute concentration of water entering the incremental reach (ΔC) at each stilling well along the reach using the following formula:

$$\Delta C = \frac{Q_d C_d - Q_u C_u}{\Delta Q} \quad (5)$$

where Q_d is downstream discharge (L/s), C_d is downstream solute concentration (mg/L), Q_u is upstream discharge (L/s), C_u is the upstream solute concentration (mg/L), and ΔQ is the net change in discharge from the upstream stilling well to the downstream stilling well (L/s). Incremental solute concentrations, such as calcium, magnesium, sodium, and chloride, were calculated for each tracer used in mixing models. These solute concentrations were applied in mass-balance mixing models ([Supplementary material](#)) to quantify fractional gains in groundwater and soil water from one sub-catchment to the next. Modeled end-member fractions were compared against the mean elevation and TWI values as well as the area for each incremental sub-catchment to evaluate the relationships between internal catchment characteristics and discharge. To evaluate these



relationships for the accumulated area, the elevation, TWI, and UAA values for the sub-watershed draining to each stilling well were compared to the original EMMA-modeled groundwater and soil water contributions at each stilling well location. Both accumulated and incremental analyses were conducted for each synoptic sampling campaign in order to evaluate seasonal change.

Hydrometric data from soil wells at various landscape positions were used as a proxy for determining when the catchment was “wet,” as the data indicated the interconnectedness of the water table over time. Full hydrographs for soil and groundwater over the study period can be found in [Gardner et al. \(2020\)](#). Soil water connection was evaluated through binary plots of saturated vs. unsaturated conditions.

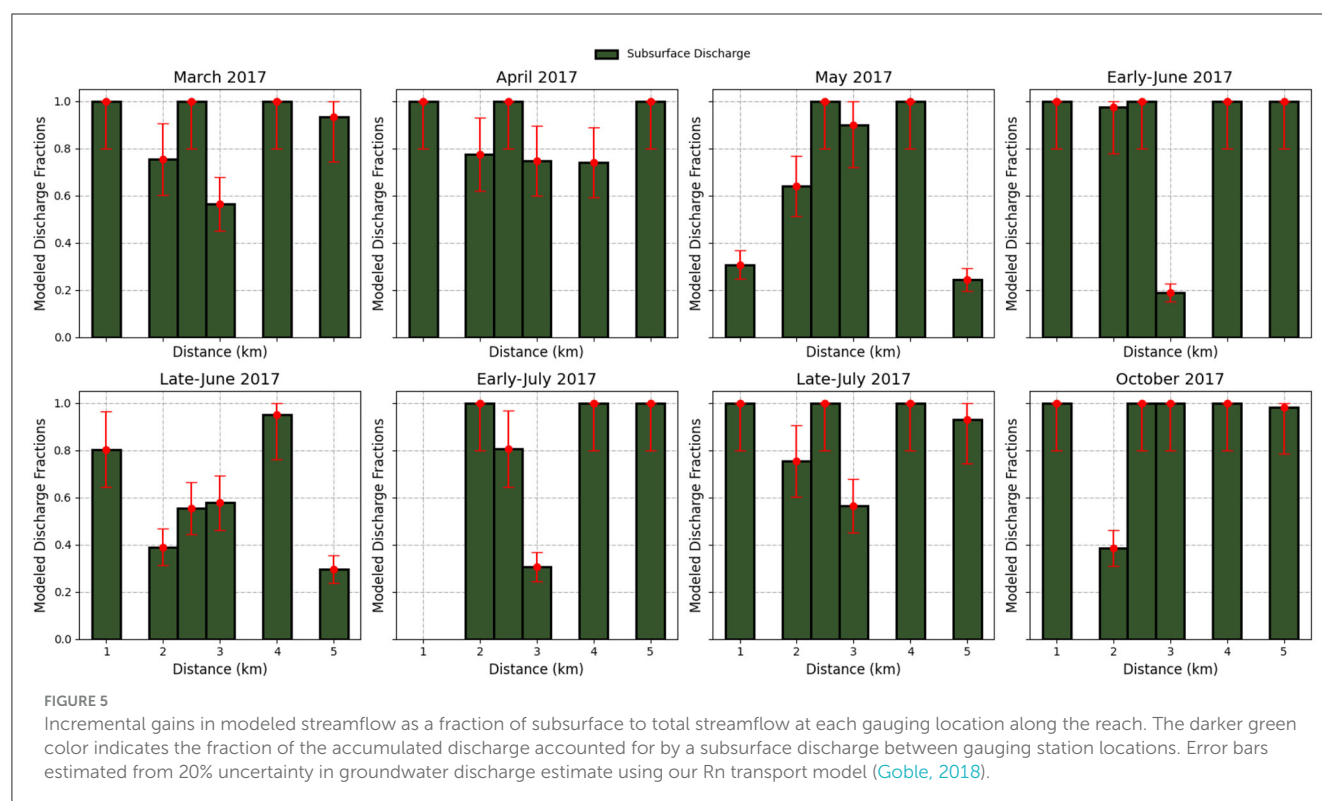
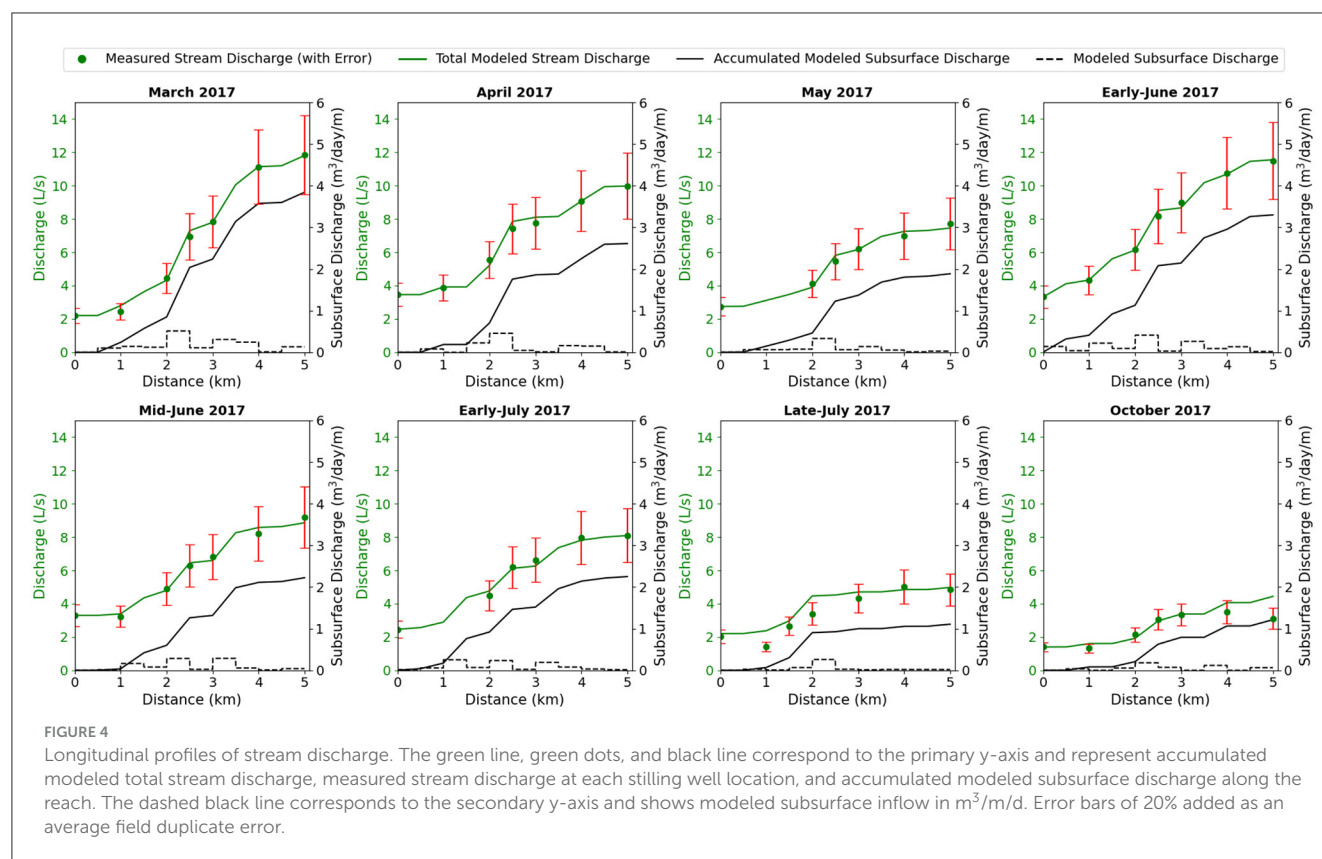
Results

Radon concentrations ranged from 0.074 to 4.25 Bq/L and showed the same general spatial pattern across varying discharges ([Figure 3](#)). During each sampling period, a significant increase in radon activity was observed approximately halfway down the observed reach (~2.5 km). This increase coincides with the location of a large contributing hillslope area from the south ([Figure 1](#)). The difference between duplicate samples ranged from 0 to 40%, with low-concentration samples exhibiting larger variability, consistent with Poisson statistics of radioactive decay. For example, at a concentration of 0.074 Bq/L (corresponding duplicate 0.111 Bq/L), there is a 40% variability, while at a

concentration of 2.527 Bq/L (duplicate 2.597 Bq/L), there is a 2.7% variability.

[Figure 4](#) shows total subsurface discharge estimates along with modeled and observed total stream discharge. The longitudinal profile of the accumulated subsurface discharge closely follows the increase in total stream discharge. For each interval, subsurface discharge accounts for 19–100% of the incremental gains in streamflow, with an average of 81% ([Figure 5](#)). In May and mid-June, the months of highest rainfall, the influence of non-subsurface discharge is strongest, with subsurface discharge accounting for, on average, 68% and 59%, respectively, of the incremental gains in streamflow. Non-subsurface discharge generation estimates predominantly occur near the center of the reach (2.5–3 km downstream of the headwaters) at the convergence of two large hillslopes.

Reprojected mean groundwater and soil water end-member concentrations are provided in [Table 3](#). The two end members chosen, spanned the sample variance and the residual error in modeled vs. observed concentration showed no structure. The fractions of soil and bedrock groundwater in stream water samples are shown in [Figure 6](#). The bedrock groundwater discharge fraction ranged from 0 to 0.45, with an average of ~0.2. In general, bedrock groundwater becomes a larger component of streamflow as the distance downstream increases, with the highest fraction of groundwater discharge occurring at the farthest downstream gauging location. [Figure 7](#) shows the percentage of bedrock groundwater at the outlet as a function of time. Groundwater at the outlet fluctuated between 26 and 44% and averaged 38% of total streamflow. The fraction of bedrock



groundwater to total streamflow at the outlet was highest during the late spring (2 June 2017) and lowest during mid-spring (19 May 2017).

Table 4 summarizes linear regression results for incremental and accumulated terrain characteristics with the bedrock groundwater mixing fraction. Linear regression of cumulative

TABLE 3 Solute concentrations used in EMMA mass-balance mixing models.

Tracer concentration (mg/L)	Groundwater	Soil water
Calcium	7.94	2.72
Chloride	1.22	1.30
Magnesium	3.93	1.84
Sodium	8.47	2.39

mean terrain indices and soil and bedrock groundwater fractions revealed strong relationships with both cumulative average elevation and cumulative upslope accumulated area (Figure 8, Table 4). R^2 values ranged from 0.52 to 0.98 over the 7-month study period. While the R^2 values ranged from 0.05 to 0.81 for the incremental terrain characteristics, they tended to switch drastically from high to low values throughout the year. No systematic trend was found relating incremental end-member discharges with incremental TWI and area (Figure 9, Table 4).

Figure 10 plots the r^2 values from cumulative and incremental regression with binary soil well responses as a function of time. Soil wells with high TWI values tended to stay continually saturated, whereas those with lower TWI values exhibited more transient-like behavior in response to climatic forcing. In both cases, the r^2 value was highest (0.98 and 0.81, respectively) in early July, ~3 weeks after the largest catchment “wet-up.”

Discussion

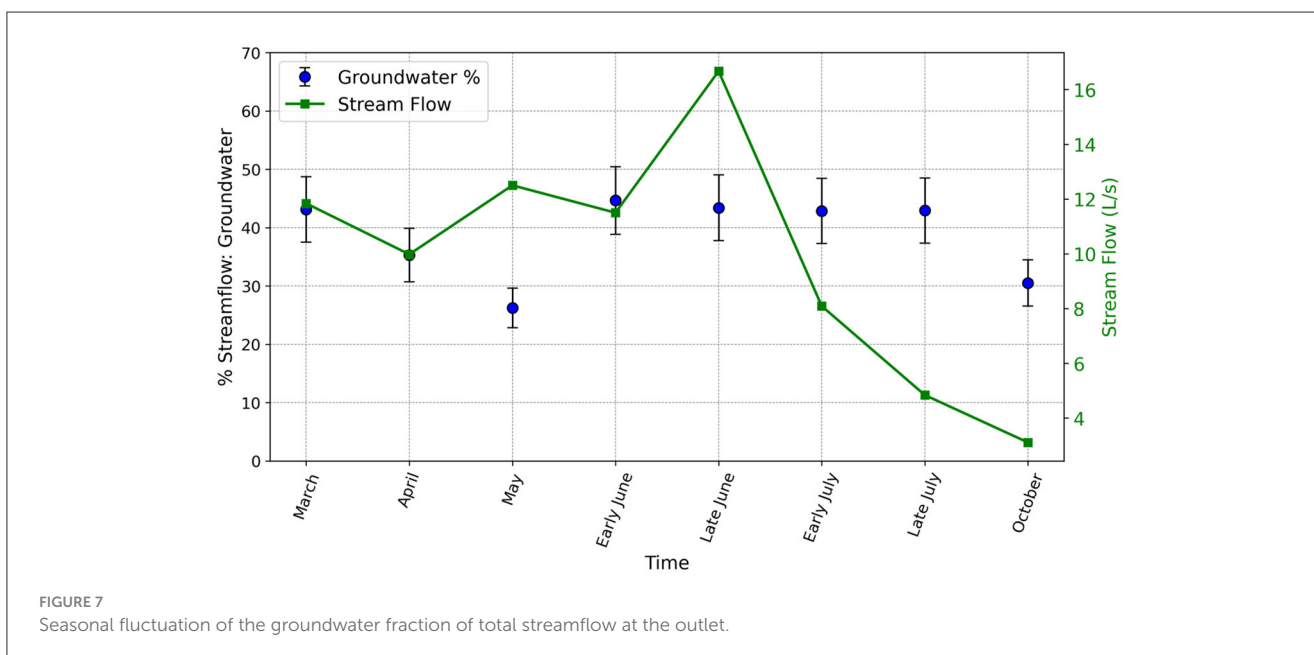
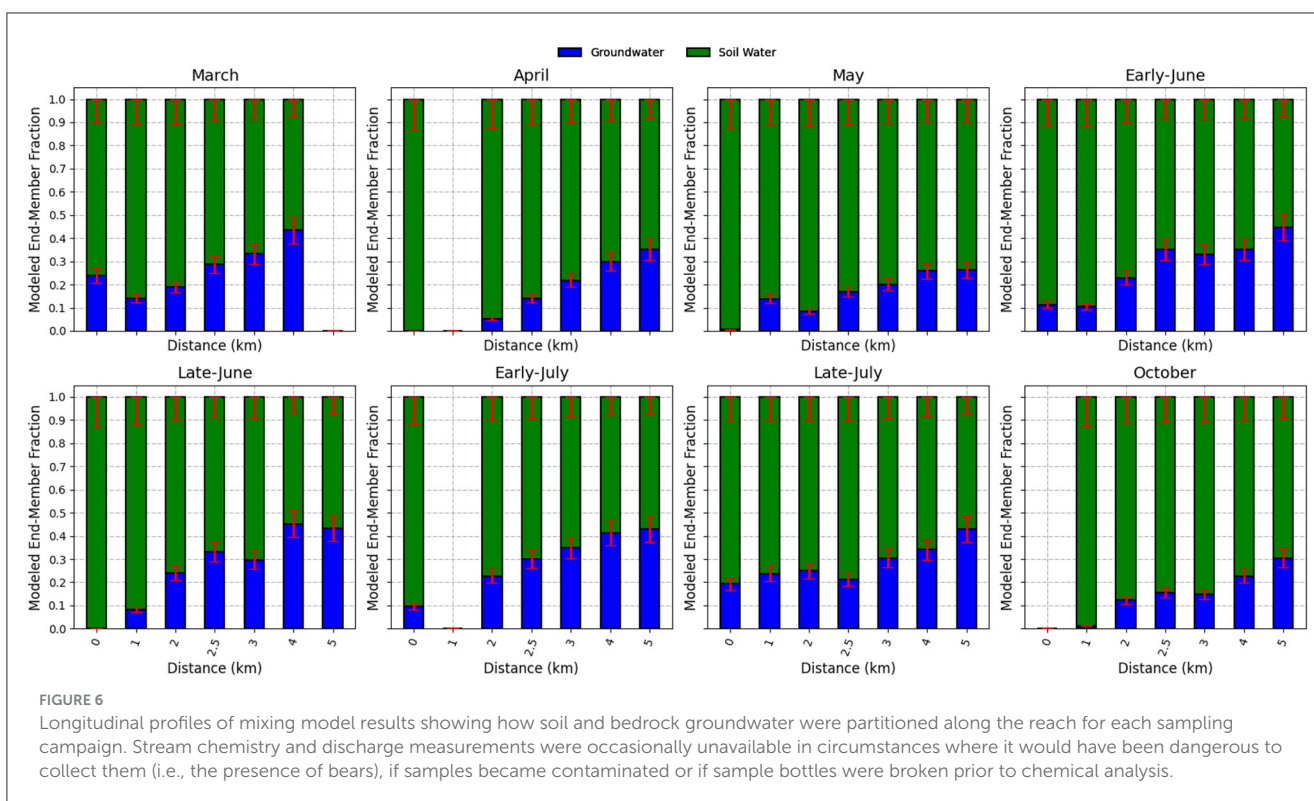
Radon results indicate the majority of runoff in CWC is derived from subsurface sources (soil and bedrock groundwater) throughout the year, even during the snowmelt cycle, consistent with prior observations in alpine catchments (Liu et al., 2004). These results may be expected in later parts of the season (late summer and fall) when the overland flow is expected to be zero in our catchment. However, the fact that subsurface flow dominates during snowmelt (March, April, and May) indicates that direct runoff through the snowpack is not a major source of water for our stream. Radon results are consistent with soil hydrometric data. Soil wells with low TWI values (generally located at higher elevations with less accumulated area) exhibit transient behavior to catchment wet-ups (Figure 10) (Gardner et al., 2020). Some wet-ups had durations on the order of hours to days, suggesting that soil water in these landscape positions is rapidly lost through lateral inter-flow and/or infiltration to the bedrock aquifer.

While incremental discharge source plots show that subsurface sources account for the majority of streamflow gains, there are some instances where non-subsurface sources are estimated to be more important in streamflow generation (Figure 5). Non-subsurface streamflow generation occurred throughout the year but had the strongest influence on stream chemistry during the late spring to mid-summer (May through late July). These instances occurred primarily in the middle section of the reach,

where multiple convergent zones from large hillslopes converge (Figure 3). Hydrometric data from soil wells indicate that the catchment was at its most saturated state during the May and June periods (Figure 10). Thus, during May and June, we hypothesize that non-subsurface sources could be due to saturation excess and overland flow in these large contributing drainage bottoms or soil and/or preferential/piping flow in the unsaturated zone. The fact that our radon modeling indicates some non-subsurface flow during the baseflow periods (late July and October) is likely due to model error. A probable source of error is the discharge of short residence time flow paths that have low R_n .

End-member mixing analysis provided an independent test of the discharge source findings, along with the ability to further divide stream-flow composition. The results indicate that the stream is generally composed of shallow saturated soil water at the headwaters and that shallow soil water is the dominant source of streamflow generation, but bedrock groundwater discharge increases with downstream distance, even during periods of snowmelt and heavy rainfall. Our bedrock groundwater discharge fractions are generally comparable with those of other mountain aquifer studies (Somers and McKenzie, 2020). The increasing influence of bedrock groundwater with downstream distance argues for a 3D watershed conceptual model consistent with those proposed by previous researchers evaluating groundwater flow in mountainous catchments in the western U.S. (Carroll et al., 2018; Frisbee et al., 2011, 2012). However, this study demonstrates that this emergent relationship exists in headwater catchments >2 km² in scale.

Temporal analysis indicates surprisingly low variability in modeled end-member proportions throughout the observed time period; however, early summer months have the highest proportion of bedrock groundwater discharge. Several large precipitation events (~60 mm/day) occurred in early June. These precipitation events could have rapidly filled the subsurface reservoirs, which could, in turn, drive increased discharge through a variety of mechanisms, including groundwater ridging, preferential recharge, and high fractured bedrock celerity. Rapid groundwater reactions to precipitation inputs are consistent with the experimental observations of Anderson et al. (1997a). However, our observed late-season bedrock groundwater ratios of streamflow are surprisingly low. Several previous hydrograph separation studies have assumed that baseflow is entirely composed of groundwater (Genereux et al., 1993; Pinder and Jones, 1969). Indeed, our results indicate that shallow soil groundwater is a substantial portion of streamflow during the late season. These results could be explained by several different mechanisms. For example, the substantial difference in the hydraulic diffusivity of soil and fractured bedrock can lead to situations where the fractured bedrock discharge exhibits a rapid response to infiltration, which is then followed by a long-term storage release from porous soil (Anderson et al., 1997b). Additionally, soil interflow and groundwater recharge and discharge have been shown to be tightly coupled at the scales of hillslopes (Gardner et al., 2020) and catchments (Carroll et al., 2019), and this interaction is capable of sustaining groundwater recharge and soil water throughflow throughout the year (Gardner et al., 2020).

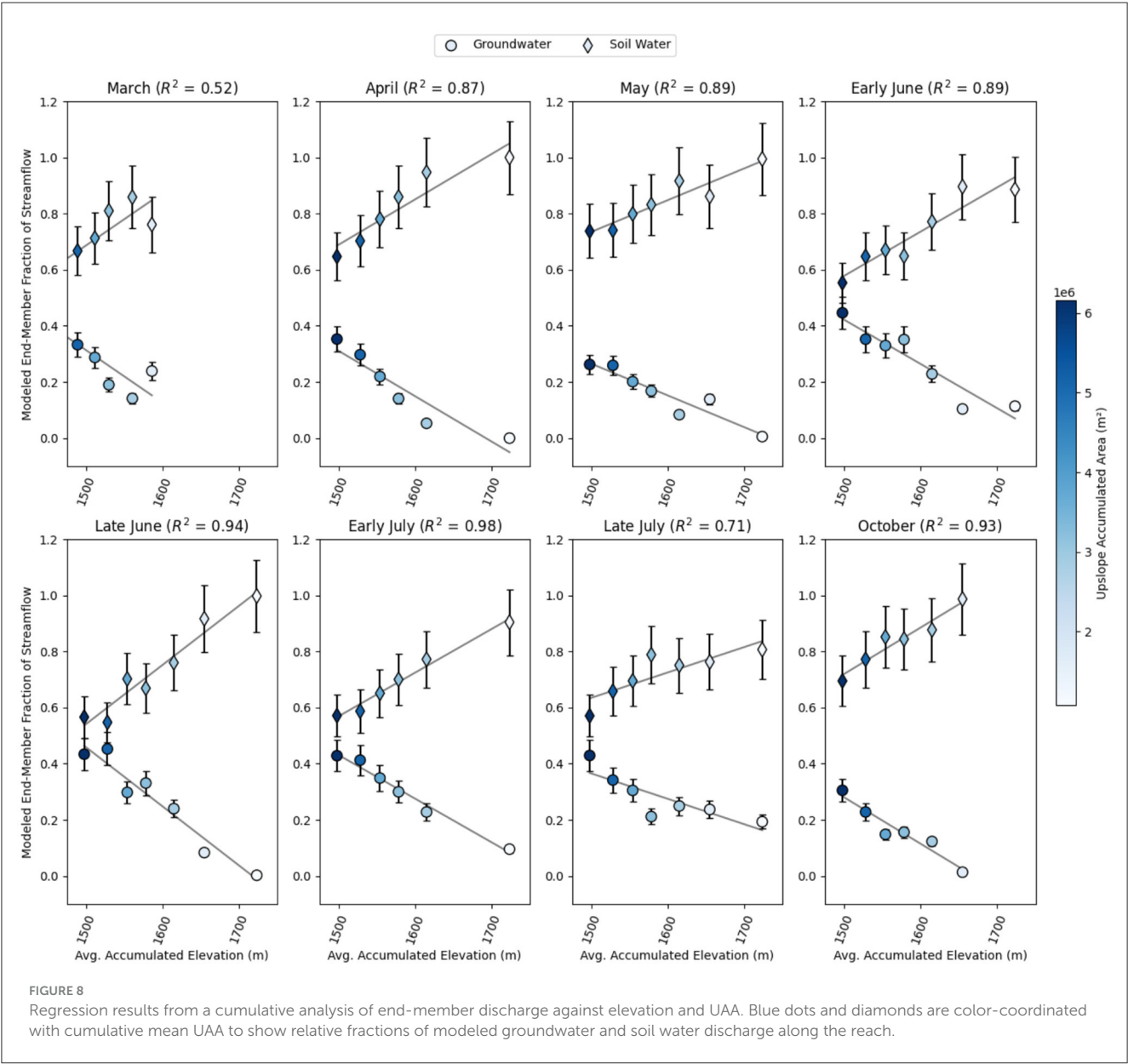


Soil groundwater remains the dominant source of subsurface discharge over our entire reach throughout the year. However, bedrock groundwater discharge becomes an increasingly higher fraction of discharge as the amount of accumulated area increases. These observed trends of high soil groundwater discharge and increasing bedrock groundwater contribution with increased catchment scale imply that streamflow represents an integration of individual hillslope responses and larger-scale deeper groundwater flow paths (Figure 11). We attribute the

increasing contribution of bedrock groundwater discharge with increasing accumulated area due to the integration of more heterogeneity and higher groundwater discharge zones (Asano et al., 2022; Frisbee et al., 2011). However, we demonstrate this relationship at a landscape scale that is orders of magnitude smaller than what has been observed in previous research. The relationship was generally consistent across variable streamflow states, with small temporal inconsistencies that are likely associated with dynamic catchment properties, such

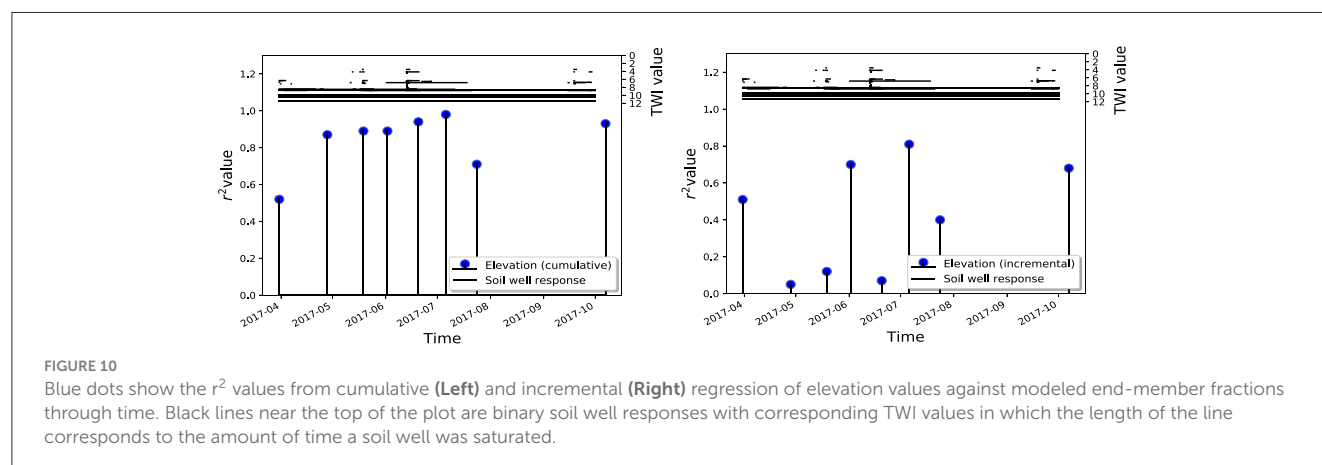
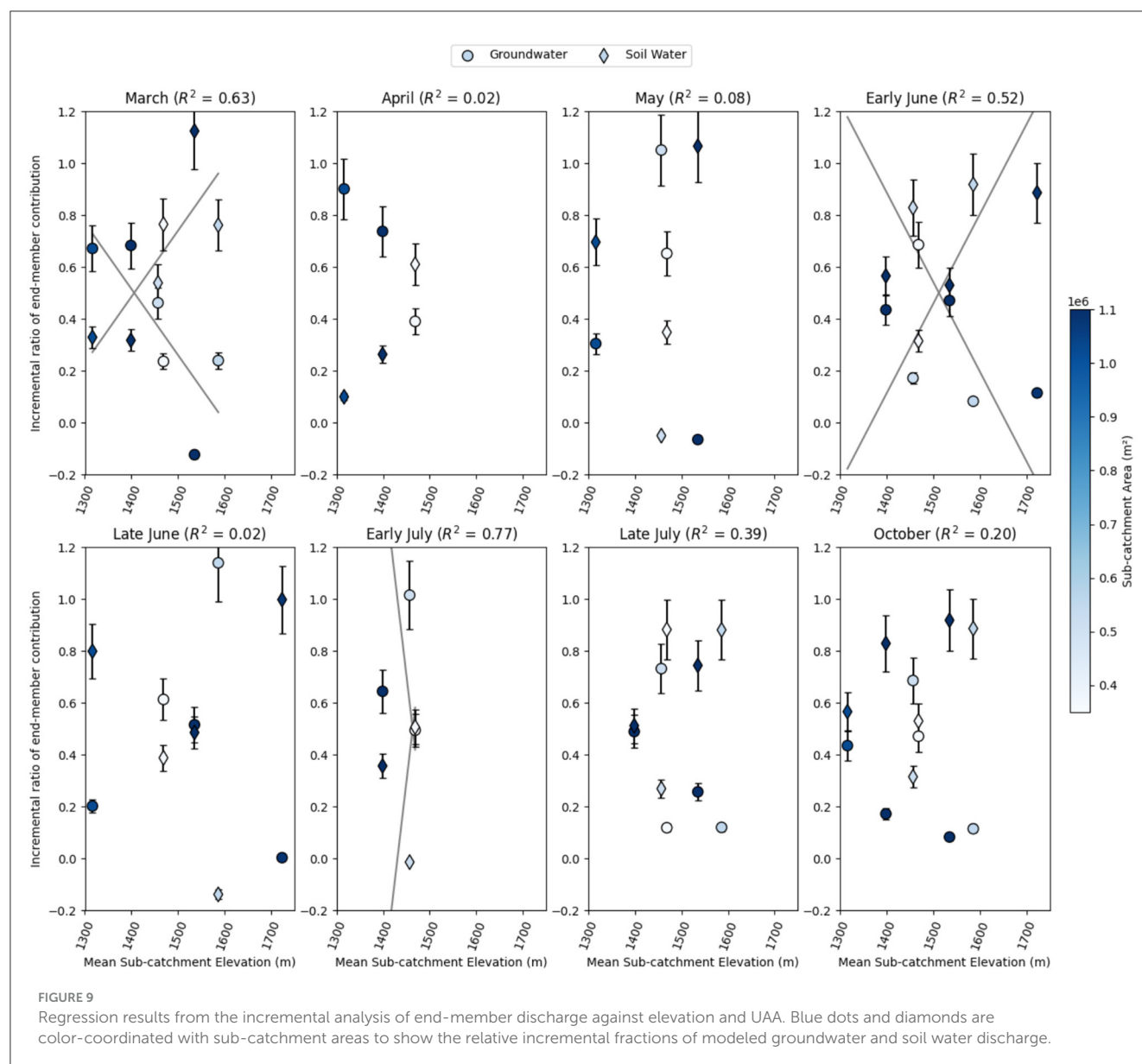
TABLE 4 Summary of r^2 values from cumulative and incremental regression against mean elevation, topographic wetness index, and sub-watershed area.

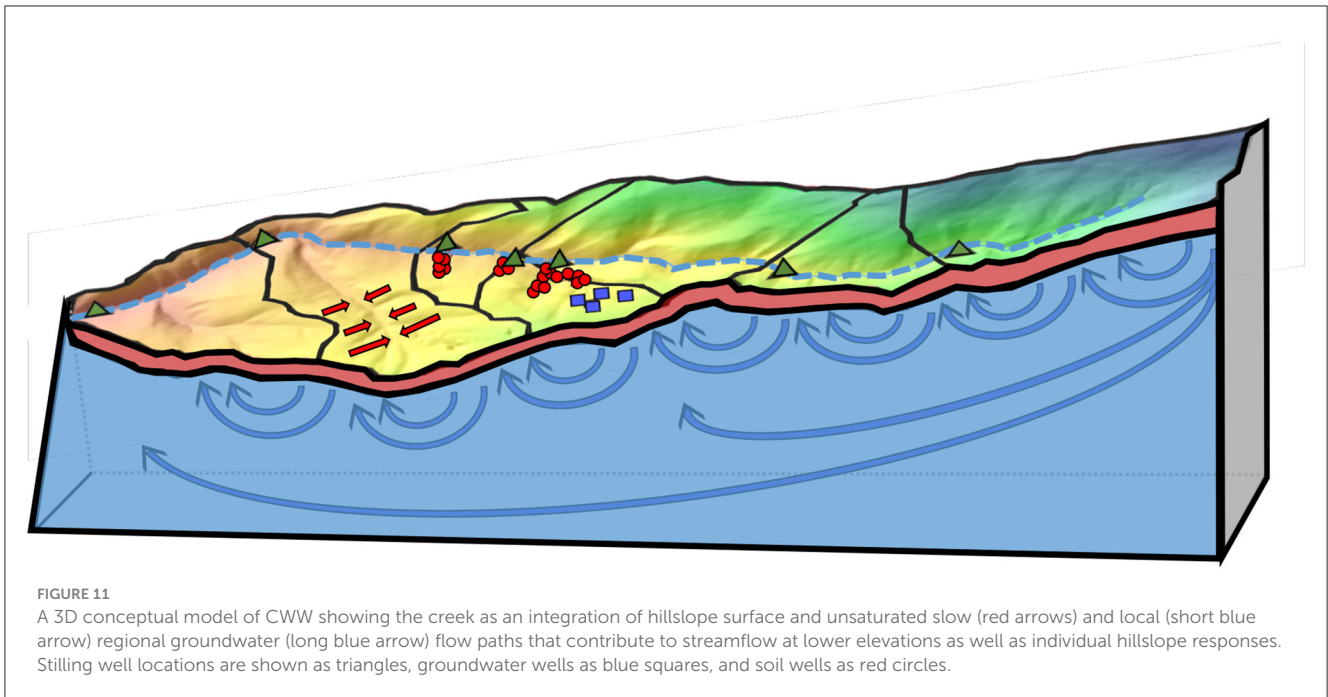
Type	Attribute	Mar	Apr	May	Early June	Late June	Early July	Late July	Oct
Incremental	Elev.	0.45	0.93	0.89	0.87	0.92	0.96	0.73	0.96
	TWI	0.47	0.22	0.59	0.78	0.93	0.74	0.35	0.69
	Area	0.09	0.05	0.00	0.00	0.04	0.00	0.11	0.23
Accumulated	Elev.	0.52	0.99	0.89	0.89	0.94	0.98	0.71	0.93
	TWI	0.66	0.96	0.80	0.95	0.97	0.98	0.62	0.90
	Area	0.71	0.91	0.80	0.86	0.89	0.92	0.85	0.96



as climate and soil moisture, highlighting the importance of antecedent conditions.

The lack of trend in bedrock groundwater discharge with incremental sub-watershed area suggests that a minimum scale exists below which it is difficult to observe trends in bedrock groundwater discharge with terrain metrics. This lack of trend could result from two proposed processes: (1) regional groundwater flow paths could contribute to streamflow in a non-linear fashion





due to geologic heterogeneities and (2) valley-bottom exchange of water. Linear trends in modeled soil and bedrock groundwater contributions against elevation and upslope accumulated area should not necessarily be expected, as a sub-catchment high in the watershed could export its regional groundwater to locations lower in the watershed. Thus, in the case of active regional groundwater flow, low-elevation sites would receive disproportionately more groundwater than high-elevation sites, potentially explaining the observed correlations with accumulated catchment area.

Caveats and limitations

Our estimates of the bedrock groundwater discharge fraction are affected by the total flow and our error in gauging low flows. For example, on 8 June 2017, a 0.25-L/s increase in streamflow was recorded when going from stilling well “CWSTW6” (located ~2.5 km downstream from headwaters) to stilling well “CWSTW2-upper” (~3 km downstream from headwaters). Over this same reach, a 0.15-L/s increase in modeled subsurface discharge was calculated, implying that 60% of discharge gained was obtained from the subsurface while 40% was obtained from the surface or unsaturated sources. However, the observed 0.25-L/s increase is small compared to the uncertainty in discharge measurements. In addition, estimated subsurface inflows are subject to uncertainty in model parameters. Groundwater inflow estimates derived from ^{222}Rn are sensitive to the gas transfer velocity (k), which is also one of the least constrained parameters in the radon transport model. Varying the gas transfer velocity by $\pm 50\%$ has a marked effect on estimated groundwater inflow: a 50% decrease in ^{222}Rn would result in an estimated gain of 0.0 L/s, implying that 100% of streamflow comes from the surface, while a 50% increase in ^{222}Rn would result in an estimated gain of 0.45 L/s, which was more

than what was actually measured. In addition, radon modeling is limited to detecting subsurface sources that have an appreciably high concentration of ^{222}Rn , which requires subsurface residence times of more than a couple of days, with secular equilibrium being reached only after ~2 weeks. It is possible that subsurface sources with brief residence times and thus low ^{222}Rn concentration could go undetected.

Overall, these uncertainties could contribute to uncertainty in the exact volume and proportion of subsurface discharge sources for a given reach, particularly at low flows for any one of our data interpretations. However, if the errors are of similar magnitude for each reach, then the overall pattern across the watershed should remain. In addition, we have the added benefit of combining discharge, soil hydrometric data, and multiple stream tracers, which helps minimize the influence of error in any single data source. Given these factors, we argue that our results are suitable for characterizing the watershed's behavior and developing conceptual models of streamflow generation.

Conclusion

In this study, multiple environmental tracers were used to estimate the discharge of soil and bedrock groundwater along a 5 km reach of a snowmelt-dominated mountainous catchment at a variety of hydrologically important times. ^{222}Rn modeling indicated that streamflow is generated predominantly from subsurface sources for all sampled periods. EMMA results indicated that streamflow chemistry could be explained by discharge from soil and bedrock groundwater. On average, bedrock groundwater comprised 38% of streamflow at the watershed outlet but fluctuated between a low of 26% in spring and a high of 44% in early summer. Bedrock groundwater discharge for individual stream segments was not well-correlated with landscape characteristics for incremental

sub-watersheds draining to the segment (01 km^2); however, it was well-correlated with the aggregated watershed landscape characteristics upstream of that segment ($> 2 km^2$). EMMA-derived mixing models showed that headwaters are dominantly composed of soil groundwater and that bedrock groundwater becomes an increasingly more important component to streamflow at larger catchment scales with lower elevations and higher UAA values, suggesting 3D topography-driven flow. For the accumulated watershed characteristics, correlations among landscape and end-member discharge were strong across variable states of catchment “wetness” conditions, and accumulated elevation or total upslope area were robust predictors of groundwater discharge during all seasons ($r^2 = 0.52$ – 0.98). Groundwater discharge can be expected to constitute a significant source of streamflow generation in upland and montane catchments throughout the year. Bedrock groundwater flow paths follow a 3D nested flow pattern where recharge originating in upper reaches can discharge in further downstream reaches according to the heterogeneous subsurface architecture. Groundwater discharge locations are strongly controlled by local heterogeneities and cannot be well-predicted from the incremental contributing area at scales below 1 km^2 . As the watershed scale exceeds 2 km^2 , the aggregated average groundwater discharge from these 3D flow paths increases with an increase in drainage area. These findings could be useful to other researchers in developing conceptual models of streamflow generation in other watersheds.

Data availability statement

The original contributions presented in the study are included in the article/[Supplementary material](#), further inquiries can be directed to the corresponding authors.

Author contributions

IV: Writing – original draft, Writing – review & editing. KJ: Conceptualization, Funding acquisition, Investigation, Project administration, Resources, Supervision, Writing – original draft, Writing – review & editing. ZH: Conceptualization, Investigation, Software, Writing – review & editing. RL: Conceptualization, Data

curation, Formal analysis, Investigation, Methodology, Writing – review & editing. WG: Conceptualization, Data curation, Formal analysis, Funding acquisition, Investigation, Methodology, Project administration, Resources, Software, Supervision, Validation, Visualization, Writing – original draft, Writing – review & editing.

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Conflict of interest

The authors declare that the research was conducted in the absence of any commercial or financial relationships that could be construed as a potential conflict of interest.

Generative AI statement

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Supplementary material

The Supplementary Material for this article can be found online at: <https://www.frontiersin.org/articles/10.3389/frwa.2025.1539177/full#supplementary-material>

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