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Hydrogeology of desert springs in the Panamint Range, California, USA: Geologic controls on the geochemical kinetics, flowpaths, and mean residence times of springs

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Abstract

Over 180 springs emerge in the Panamint Range near Death Valley National Park, CA, yet, these springs have received very little hydrogeological attention despite their cultural, historical, and ecological importance. Here, we address the following questions: (1) which rock units support groundwater flow to springs in the Panamint Range, (2) what are the geochemical kinetics of these aquifers, and (3) and what are the residence times of these springs? All springs are at least partly supported by recharge in and flow through dolomitic units, namely, the Noonday Dolomite, Kingston Peak Formation, and Johnnie Formation. Thus, the geochemical composition of springs can largely be explained by dedolomitization: the dissolution of dolomite and gypsum with concurrent precipitation of calcite. However, interactions with hydrothermal deposits have likely influenced the geochemical composition of Thorndike Spring, Uppermost Spring, Hanaupah Canyon springs, and Trail Canyon springs. Faults are important controls on spring emergence. Seventeen of twenty-one sampled springs emerge at faults (13 emerge at low-angle detachment faults). On the eastern side of the Panamint Range, springs emerge where low-angle faults intersect nearly vertical Late Proterozoic, Cambrian, and Ordovician sedimentary units. These geologic units are not present on the western side of the Panamint Range. Instead, springs on the west side emerge where low-angle faults intersect Cenozoic breccias and fanglomerates. Mean residence times of springs range from 65 (± 30) to 1,829 (± 613) years. A total of 11 springs have relatively short mean residence times less than 500 years, whereas seven springs have mean residence times greater than 1,000 years. We infer that the Panamint Range springs are extremely vulnerable to climate change due to their dependence on local recharge, disconnection from regional groundwater flow (Death Valley Regional Flow System - DVRFS), and relatively short mean residence times as compared with springs that are supported by the DVRFS (e.g., springs in Ash Meadows National Wildlife Refuge). In fact, four springs were not flowing during this campaign, yet they were flowing in the 1990s and 2000s.

KEYWORDS

crenobiontic, Death Valley, desert springs, fault hydrogeology, geochemical kinetics, Panamint Range, residence times, spring vulnerability

1 | INTRODUCTION

Over 2,000 springs emerge in the southern Great Basin, which includes Death Valley, within one of the most geologically complex and climatologically extreme environments on Earth (Bedinger & Harrill, 2012; Sada & Pohlmann, 2007). These springs developed in response to tectonic processes at long timescales (Ma) and climate variations at shorter timescales (ka). Today, these springs are biodiversity hotspots (Myers & Resh, 1999; Souza, Siefert, Escalante, Elser, & Eguarte, 2012; Stevens & Meretsky, 2008). These springs are known to support more than 40 taxonomically distinct crenobiontic (obligate spring-dwelling species) vertebrates and invertebrates (Keleher & Sada, 2012; Sada, Britten, & Brussard, 1995; Sada & Vinyard, 2002; Thomas et al., 2013) and over 500 known distinct species of *Bacteria* and *Archaea* (Thomas et al., 2013).

The Panamint Range consists of the Owlshhead Mountains to the south, central Panamint Mountains, and Cottonwood Mountains to the north. However, the focus of this study is on springs emerging in the Panamint Mountains. We use the terminology “Panamint Range” throughout this article to be consistent with Gleason, Frisbee, Rademacher, Sada, and Meyers (2019) and since Panamint Range is often used informally in reference to the Panamint Mountains. The Panamint Range located partially within Death Valley National Park (Figure 1), hosts over 180 springs that developed within this tectonic-

climatic framework. While some mountain ranges in the southern Great Basin contribute groundwater flow to the Death Valley Regional Flow System (DVRFS), the tectonic processes that formed the Panamint Range also disconnected it from the DVRFS (Belcher, Sweetkind, Faunt, Pavelko, & Hill, 2017). Consequently, the springs emerging in the Panamint Range are dependent upon local recharge (Gleason et al., 2019). Unfortunately, our understanding of groundwater flow processes post-recharge in the Panamint Range is extremely limited. Here, we build upon the work of Gleason et al. (2019) and address the following questions: (1) which rock units support groundwater flow to springs in the Panamint Range, (2) what are the geochemical kinetics of these aquifers, and (3) and what are the residence times of springs in the Panamint Range?

We answer questions 1 and 2 using a combination of geochemical analyses that include an evaluation of dissolved major-ion concentrations and strontium isotope ($^{87}\text{Sr}/^{86}\text{Sr}$) compositions. We answer question 3 using a combination of tritium (^3H) and radiocarbon (^{14}C) age-dating, and an isotopic chronometer built using chlorine-36 (^{36}Cl). Although limited data were available on the Panamint Range springs prior to our study (see Faunt, D'Agnese, & O'Brien, 2010; King & Bredehoeft, 1999), the data presented here represent a spatially comprehensive quantification of groundwater flow and geochemical processes in the Panamint Range. One of the broader overarching goals of the current research in the southern Great Basin is to test the

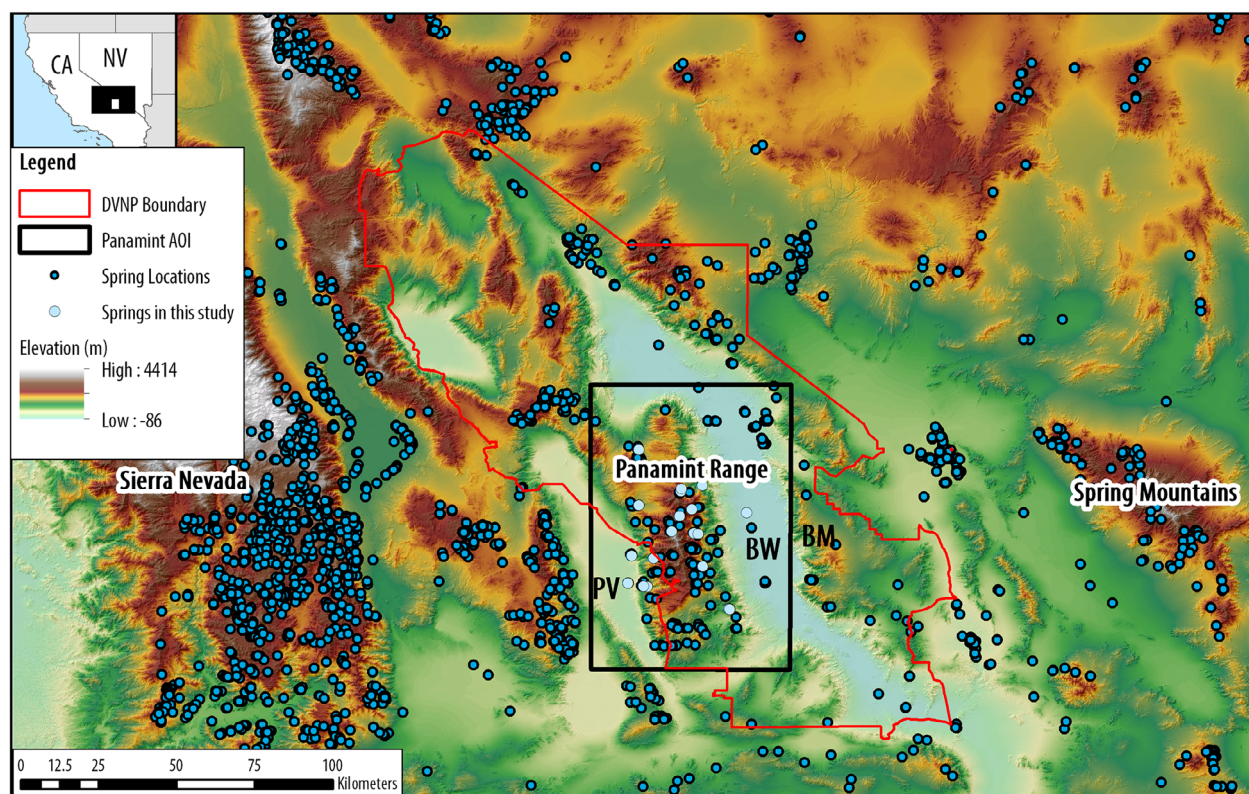


FIGURE 1 Regional map showing the location of the Panamint Range, surrounding mountain ranges, and spring locations (the black box in the inset shows the location of the study area relative to a basemap of the USA). The light blue circles represent the locations of springs sampled in this study. Information on the location of springs is sourced from the USGS National Hydrography Database (<https://nhd.usgs.gov>)

hypothesis that the desiccation of desert springs will proceed from springs with the shortest groundwater residence times to the longest. In other words, the robustness of the hydrological systems (i.e., resistance to desiccation of springs) should be positively correlated with the residence time of groundwater in the system. The impacts of this desiccation are illustrated in the schematic diagram shown in Figure 2. Here springs with the shortest residence times have few stress-intolerant crenobiontics, while springs with long residence times have many stress-tolerant crenobiontics. Under aridification, springs with short residence times are the least resistant to hydrogeological change and will therefore be the most vulnerable (and their aquatic ecosystems will likewise be the most susceptible to desiccation). We assess the results of this study in the context of this conceptual model. Our conceptual framework will inform the management of desert springs regionally and increase attention on the vulnerability of desert springs globally.

2 | STUDY AREA

2.1 | Geology of the Panamint Range

Tectonic extension in the southern Great Basin uplifted and disconnected mountain ranges in the area and created basin and range topography starting about 14 Ma (McQuarrie & Wernicke, 2005; and references therein). At about 3 Ma, rapid tectonic extension (pull-apart basin extension) created the long, narrow valleys oriented

north-south bounded by high mountain ranges that are observed today (Knott et al., 2008; Norton, 2011; Phillips, 2008). These tectonic processes affected the climatology of the region which, in turn, impacted continental sedimentation (Chapin, 2008), created the southern Sierra Nevada rain shadow (Henry, 2009; Winograd, Szabo, Coplen, Riggs, & Kolesar, 1985), and modified the regional hydrological system (Knott et al., 2008; Phillips, 2008). Climate variations, namely, glacial-interglacial cycles (Winograd et al., 1992), have been the major driver of hydrological change in the region for the last 500 ka (Jayko, 2009; Knott et al., 2008; Phillips, 2008). Surface water drainages and groundwater flowpaths adjusted in response to the changes in topography and climatology. As glaciers advanced in the Sierra Nevada, runoff increased and the river-lacustrine system became connected (Knott et al., 2008). The Last Glacial Maximum represents the last period that the Death Valley river-lacustrine system was likely interconnected. As the glaciers receded and the climate warmed following the Last Glacial Maximum, the interconnected river-lacustrine system dried leaving behind numerous isolated spring systems. The resultant hydrologic fragmentation occurring over 3 to 2 Ma created conditions favorable to genetic isolation and the evolution of new species (Echelle, 2008; Echelle et al., 2005; Hershler & Liu, 2008; Smith et al., 2002).

The Panamint Range is a product of the tectonic forces described above and the springs that emerge in the Panamint Range are a product of coupled tectonic-climatic-hydrologic processes. As a consequence, the Panamint Range, located between Panamint Valley to the west and Death Valley to the east, is steep and rugged (Figure 1).

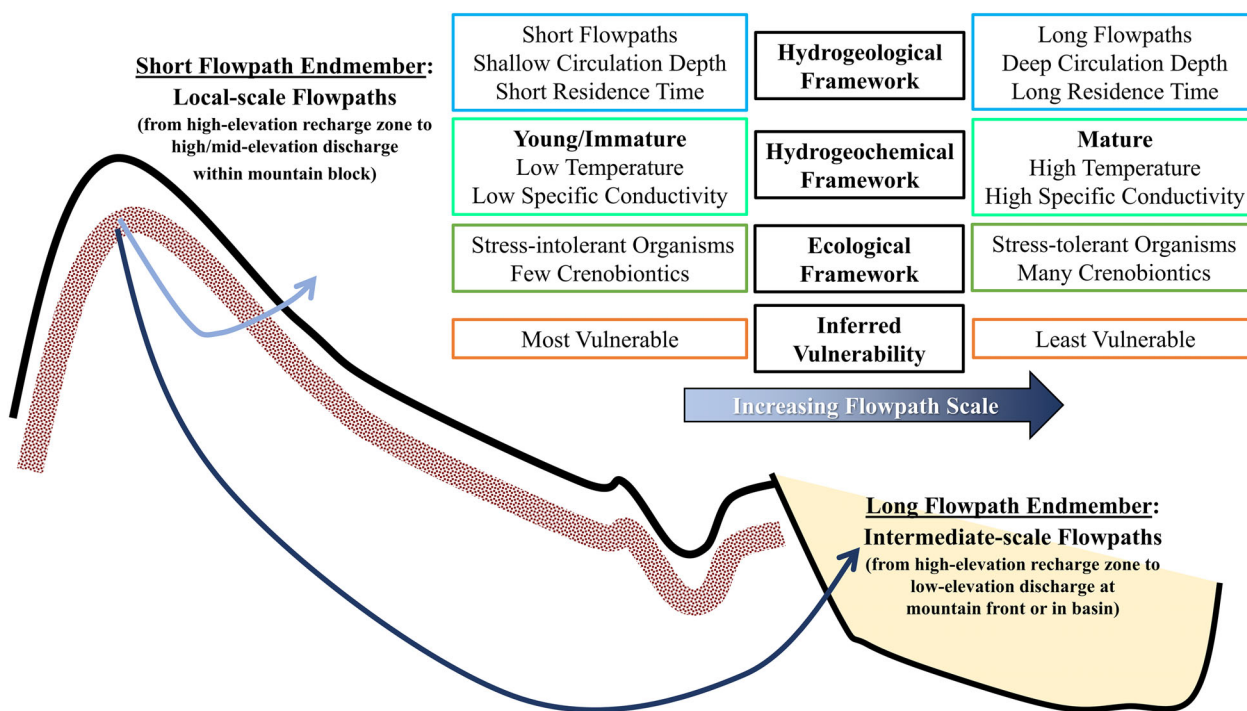


FIGURE 2 Schematic diagram showing the relationship between the ecological framework and the hydrogeochemical and hydrogeological frameworks. The inferred vulnerability of the springs to aridification is shown in the lower boxes. The arrow shows the increase in flowpath lengths from short (local-scale) flowpaths to long (intermediate-scale) flowpaths

Telescope Peak (UTM: 11S, 491,979 mE, 4,002,807 mN) is the highest point in the Panamint Range (3,366 mrsi [meters relative to sea level]). Badwater, located 29.8 km to the east of the Panamint Range, is the lowest point in North America (−85 mrsi, UTM: 11S, 516,218 mE, 4,011,516 mN). Due to the regional tectonic history, the geology of the Panamint Range is extraordinarily complicated with complexly and pervasively folded, fractured, and faulted rocks spanning nearly 2 Ga (Figures 3 and S1; Albee, Labotka, Lanphere, & McDowell, 1981; Brush et al., 2019; Hunt & Mabey, 1966; Petterson, Prave, Wernicke, & Fallick, 2011; Workman et al., 2016). Rather than provide a highly detailed geologic map in the main text, we provide a simplified geologic map of the Panamint Range where geologic units are lumped according to geologic age (Figure 3). However, a detailed geologic map modified from Workman et al. (2016) is provided in Figure S1 along with detailed descriptions of geologic units in the Supporting Information. The oldest basement rock of the Panamint Range is the middle Proterozoic (1.7 Ga) mylonitic gneiss of the World Beater Complex, which is intruded by 1.4 Ga quartz monzonite (Albee et al., 1981; Petterson et al., 2011; Figures 3 and S1). The World Beater Complex is overlain by the Mesoproterozoic to Neoproterozoic Pahrump Group (Crystal Spring, Beck Spring, and Kingston Peak; Albee et al., 1981; Petterson et al., 2011). The World Beater Complex and Pahrump Group are considered confining units in the DVRFS (Sweetkind, Belcher, Faunt, & Potter, 2010).

The Pahrump Group is overlain by late Palaeozoic units (Noonday Formation [dolomite, limestone, sandstone], Johnnie Formation [shale to dolomite], Stirling Quartzite) and early Cambrian Wood Canyon Formation [dolomite, quartzite, shale]. Detachment surfaces are mapped in the Noonday Formation, Johnnie Formation, and Stirling Formation (Norton, 2011). Exposed on the east side of the Panamint Range are lower Cambrian Zabriskie Quartzite, Cararra Formation [limestone, dolomite], Bonanza King Formation [dolomite], and Nopah Formation [dolomite] along with Oligocene limestone and quartzite (Hunt & Mabey, 1966; Workman et al., 2016). This rock sequence is classified as the lower carbonate aquifer of the DVRFS (Sweetkind et al., 2010).

Late Mesozoic to Cenozoic intrusive and sedimentary rock units are found across the Panamint Range. Intrusive plutons range from the Cretaceous Hall Canyon pluton to the Miocene Little Chief Stock (Labotka, Albee, Lanphere, & McDowell, 1980). The 100.6 ± 7.6 Ma Skidoo granite (Hodges, McKenna, & Harding, 1990) and Neogene Hanaupah granite (Figure S1; Hunt & Mabey, 1966) are particularly important to our study since they may affect groundwater flowpaths in the central portion of the Panamint Range. In the northwest part of the Panamint Range are Cenozoic breccias and fanglomerates of the Nova Formation (Hunt & Mabey, 1966). Sweetkind et al. (2010) categorized all intrusive rocks as confining units whereas Cenozoic clastic rocks are categorized as aquifers.

The Panamint Range is extensively faulted, folded, and fractured (Figures 3 and S1; Albee et al., 1981; Brush et al., 2019; Cichanski, 2009; Hunt & Mabey, 1966; Workman et al., 2016). Numerous north–south trending normal faults and low-angle detachment faults are present (Workman et al., 2016). Detachment faults (east-dipping, low-angle normal faults) offset many rock units

including the Noonday Formation, Johnnie Formation, and Stirling Formation which crop out along the crest of the Panamint Range (Labotka & Albee, 1990; Workman et al., 2016). The eastern dip reflects the movement of the extensional allochthon on the low-angle detachment faults (Maxson, 1950; Miller, 1987; Norton, 2011). The geologic units cropping out on the northeastern side of the Panamint Range are nearly vertical and the normal faults associated with these units terminate at detachment faults (Miller, 1987). Slide breccias and thick breccias of the overlying rock are mapped at the detachment fault contact (Hamilton, 1988; Hunt & Mabey, 1966). The regional detachment fault is well exposed in the lower reaches of Hanaupah Canyon where it offsets Wood Canyon Formation and Proterozoic mylonitic gneiss (Hamilton, 1988; Hunt & Mabey, 1966). Numerous strike-slip faults of varying length are present in the Panamint Range (Albee et al., 1981; Hunt & Mabey, 1966; Wrucke, Stone, & Stevens, 2007). Sweetkind et al. (2010) treated fault-related permeability in different ways depending on the fault type in the DVRFS model. Faunt (1995), however, suggested that extensional faults (i.e., normal and detachment faults) are predominantly permeable due to the associated brecciation along the contact. By contrast, transform and compressional faults (i.e., strike slip, reverse, and thrust faults) are predominantly impermeable (Faunt, 1995).

2.2 | Hydrogeology of the Panamint Range

Regionally, mountain ranges, like the Panamint Range, receive higher amounts of precipitation than their adjacent basins. Recharge from mountain precipitation supports local groundwater flow within mountain blocks and can also contribute flow to the DVRFS (Sweetkind et al., 2010). Surface flow from the mountain block rarely reaches the basin but is critical to mountain-front recharge (MFR) on alluvial fans and bajadas, especially within Hanaupah Canyon and Surprise Canyon (Figure 4). Gleason et al. (2019) quantified the sources of recharge that support springs in the Panamint Range. They found that snow-melt conservatively accounts for 57 (± 9) to 79 (± 12) percent of recharge; rainfall accounts for the remainder. Recharge is largely confined to the Noonday Formation, Johnnie Formation, and Kingston Peak Formation because these formations crop out along the crest and highest elevations of the Panamint Range where the highest amount of snow accumulates.

2.3 | Climatology of the Panamint Range

Gleason et al. (2019) provide detailed descriptions of the climatology of the study area and only a summary is provided here. Meteorological data from a weather station near Emigrant Canyon Pass (1,225 mrsi) indicate that the average summer temperature ranges from 17 to 34°C and the average winter temperature ranges from 0 to 12°C. In comparison, average summer temperatures at Furnace Creek (−59 mrsi) in Death Valley range from 30 to 46°C and average winter temperature range from 4 to 19°C (Stachelski, 2013). Average rainfall

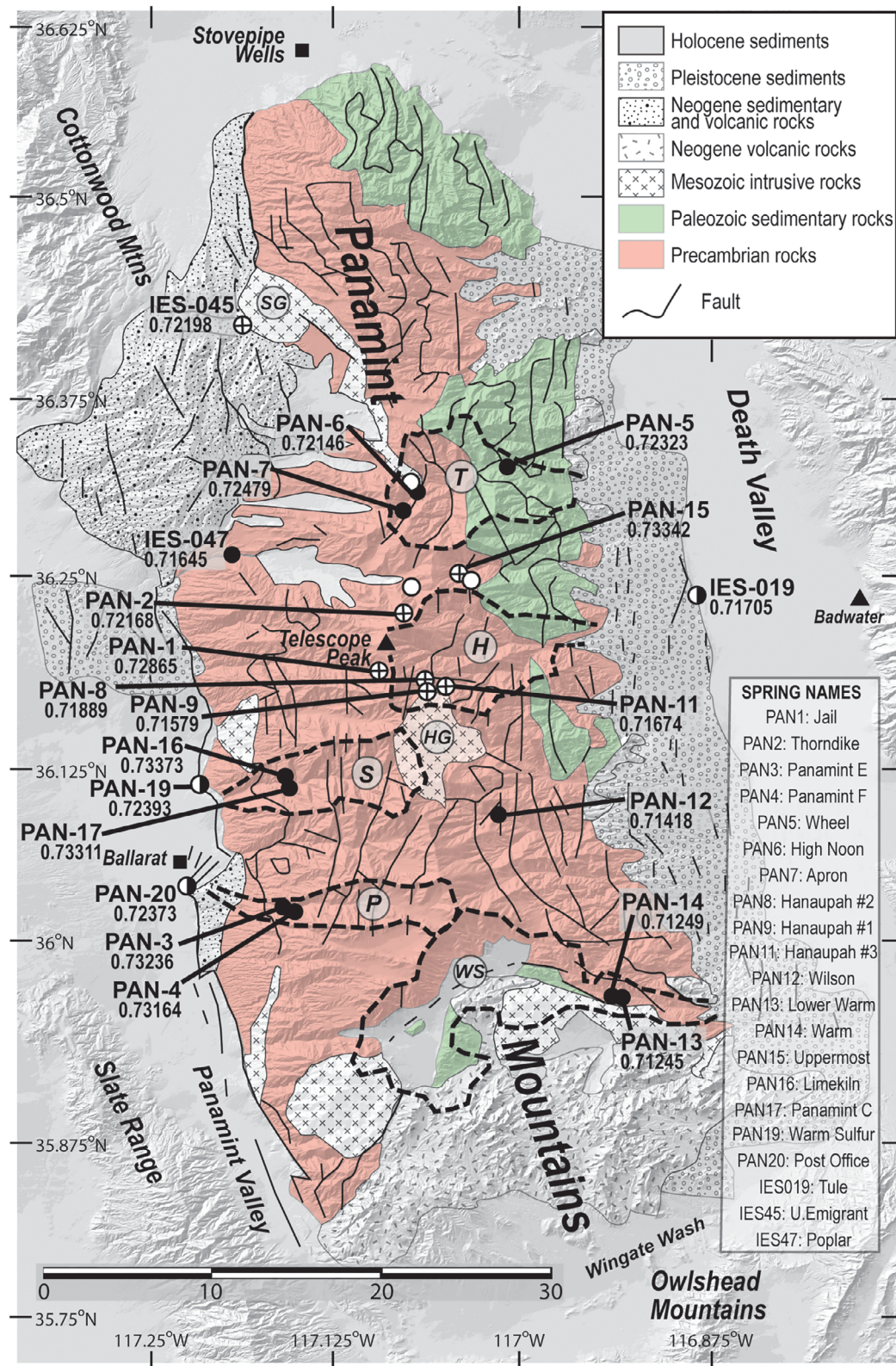


FIGURE 3 Generalized geologic map of the Panamint Mountains showing springs with $^{87}\text{Sr}/^{86}\text{Sr}$ ratio. Circles are locations of springs: dry (white); basin, sodium-chloride-water, springs (half white/half black); high elevation, calcium-bicarbonate-water, springs (white with cross); and calcium-sulfate-water springs (black). Dashed outlines are drainage basins of canyons referred to in text: Hanaupah Canyon (H), Pleasant Canyon (P), Surprise Canyon (S), Trail Canyon (T), and Warm Springs Canyon (WS). Black squares are semi-developed areas. A detailed geologic map is available in Figure S1

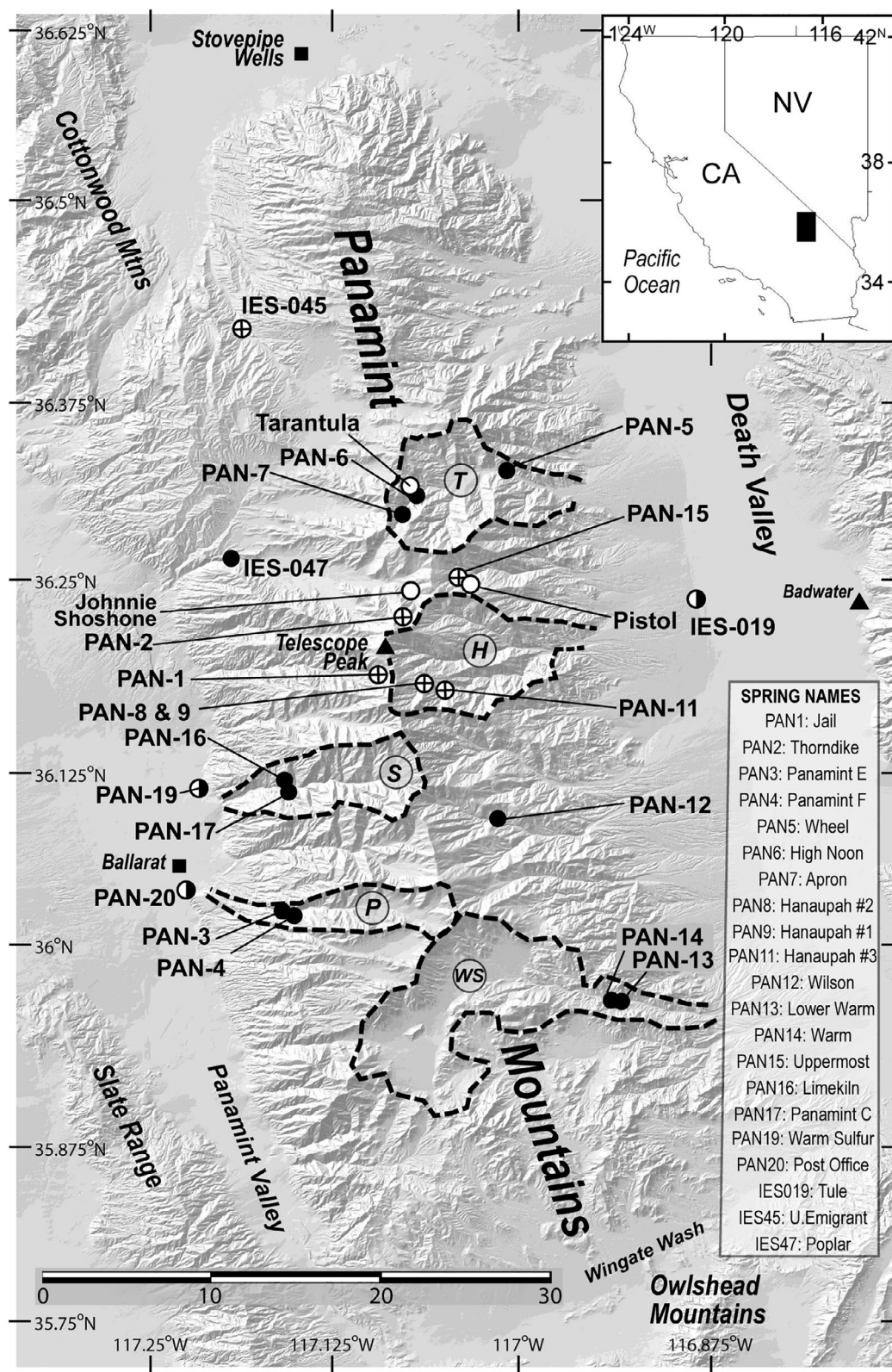


FIGURE 4 Shaded relief map of the Panamint Mountains and surrounding area

in the western Panamint Valley floor ranges from 8 to 10 cm year⁻¹, whereas Badwater Basin to the east averages less than 5 cm year⁻¹ (Wauer, 1964). PRISM data show that the crest of the Panamint Range has an average annual precipitation of 48.6 cm year⁻¹

(PRISM, 2004). Snowfall is common along the crest of the Panamint Range in the winter months (December, January, and February), however, the snowpack is not as thick as it is in the southern Sierra Nevada. Intense thunderstorms can occur in early spring and early fall,

but minimal recharge is attributed to these storms based on stable isotope data (Gleason et al., 2019).

3 | METHODS

3.1 | Sampling of springs

A total of 18 springs (Figure 3; Table 1) were sampled during a field campaign from May 23, 2017 to June 2, 2017. Samples of spring water were collected for stable isotopic analyses (presented in Gleason et al., 2019), geochemical and isotopic analyses (presented here), and microbiology and benthic macroinvertebrate communities (planned for a future work). All water samples were collected using a portable peristaltic pump (Miller & Frisbee, 2018) and Masterflex silicon tubing. One end of the tubing was placed in the spring orifice,

where possible, and the other was attached to a 0.22 μ m polyethersulfone membrane Sterivex-GP pressure filter. The tubing was flushed for 10 min before attaching the filter to collect samples. When the spring emergence was not accessible, samples were collected in the spring run downslope of, and as close as possible to, the spring emergence. Samples of water were stored unrefrigerated in the field due to the remote location of the study site and were promptly refrigerated upon return to the lab (1–5 days after collection). A 50% ethanol (C_2H_5OH) solution was used to disinfect all equipment (tubing, filters, and buckets), shoes, and clothes of the research team between each spring sampling site. Data from three additional springs, Tule Spring (IES-019), Upper Emigrant Spring (IES-045), and Poplar Spring A (IES-047), are included in this analysis since they emerge in the Panamint Range, but were sampled as part of the larger project funded by the NSF Integrated Earth Systems Program. These three springs were sampled in 2016 and 2017.

TABLE 1 Listing of spring IDs (specific to this project) and names (*denotes an informal name—spring is not named on the quadrangle), spring description (seep [diffuse emergence], small spring [<20 cm wide spring run], and moderate spring [<1 m wide spring run]), quadrangle topographic map showing spring location (7.5' Topo), drainage where spring emerges, rock units, rock type, contact type, and reference (Ref.)

Spring ID	Spring name	Spring description	7.5' Topo	Drainage	Rock units	Rock type	Contact type	Ref.
PAN 1	Jail Spring	Small spring	T.P. ¹	Jail Canyon	Zn/Zj	LS	Depos.	ALLM
PAN 2	Thorndike Spring*	Small spring	T.P. ¹	Mahogany Flat	Zj	LS	None	ALLM
PAN 3	Unnamed Panamint Spring E*	Small spring	B. ¹	Pleasant Canyon	Zk	Qtz/LS	Depos.	ALLM
PAN 4	Unnamed Panamint Spring F*	Small spring	B. ¹	Pleasant Canyon	Xmi	Gnss	None	ALLM
PAN 5	Wheel Spring*	Seep	W.P. ⁰	Trail Canyon	Cc	LS	None	HM
PAN 6	High Noon Spring*	Small spring	W.P. ⁰	Trail Canyon	Kis/Cn	G/LS	Fault	HM
PAN 7	Apron Spring*	Small spring	W.P. ¹	Trail Canyon	Cn	LS	Fault	HM
PAN 8	Main Hanaupah Spring # 2*	Moderate spring	T.P. ¹	Hanaupah Canyon	Zn (RM)	DS	None	ALLM
PAN 9	Main Hanaupah Spring # 1*	Moderate spring	T.P. ⁰	Hanaupah Canyon	Zk/Zn	MS/LS/C	Fault	ALLM
PAN 11	South Hanaupah Spring # 3*	Moderate spring	T.P. ⁰	Hanaupah Canyon	Zj	LS	Fault Block	ALLM
PAN 12	Wilson Spring*	Small spring	P. ¹	Johnson Canyon	Zj	LS	Fault	ALLM
PAN 13	Lower Warm Spring A*	Small spring	A.S. ¹	Warm Spring Canyon	PIPb/Jis	BVT	Noncon.	WSS
PAN 14	Lower Warm Spring B*	Moderate spring	A.S. ¹	Warm Spring Canyon	PIPb/Jis	BVT	Noncon.	WSS
PAN 15	Uppermost Spring*	Seep	W.P. ⁰	Death Valley Canyon	Zj/Zs	LS/Qtz	None/fault	HM
PAN 16	Limekiln Spring	Small spring	B. ¹	Surprise Canyon	Zj	LS	None	ALLM
PAN 17	Unnamed Panamint Spring C*	Small spring	B. ¹	Surprise Canyon	Zs	Qtz	Fault	ALLM
PAN 19	Warm Sulfur Spring	Seep	B. ¹	Panamint Valley	Qay	MS/SS	Depos.	ALLM
PAN 20	Post Office Spring	Seep	B. ¹	Panamint Valley	Qay	MS/SS	Depos.	ALLM
IES-019	Tule Spring (March 2016)	Seep	H.C. ¹	Death Valley	Qay	SP/Carb	Depos.	HM
IES-019	Tule Spring (March 2017)	Seep	H.C. ¹	Death Valley	Qay	SP/Carb	Depos.	HM
IES-045	Upper Emigrant Spring	Small spring	E.C. ¹	Emigrant Canyon	NF/Kis	SS/G	Fault	HM
IES-047	Poplar Spring*	Seep	E.P. ¹	Wildrose Canyon	Zk brecc	MS/C	?	ALLM

Note: The 7.5' Topo column contains the following abbreviations: T.P. = Telescope Peak, B. = Ballarat, W.P. = Wildrose Peak, P. = Panamint, A.S. = Anvil Spring Canyon West, H.C. = Hanaupah Canyon, E.C. = Emigrant Canyon, and E.P. = Emigrant Pass. The superscripts in the 7.5' Topo column indicate: 0 = not marked on the quadrangle and 1 = marked on the quadrangle. Abbreviations for rock units include: Zn = Noonday Formation (RM = Redlands Member), Zj = Johnnie Formation, Zk = Kingston Peak Formation (brecc = breccia), Xmi = World Beater Complex, Cc = Carrara Formation, Kis = silicic intrusive rocks (including the Skidoo Granite), Cn = Nopah Formation, PIPb = Bird Spring Formation, Jis = silicic intrusive rocks (including tonalites), Zs = Stirling Quartzite, Qay = Quaternary playa and fan/bajada deposits, NF = Nova Formation. Rock types include: LS = limestone, Qtz = quartzite, Gnss = gneiss, G = granite, DS = dolostone, MS = mudstone, C = conglomerate, BVT = Butte Valley Thrust, SS = sandstone, SP = silty playa, and Carb = lake beds with carbonate. Contact types include: none (no contact marked), fault, Depos. = depositional, and Noncon. = nonconformity. References includes: ALLM = Albee et al. (1981), HM = Hunt and Mabey (1966), and WSS = Wrucke et al. (2007).

3.2 | Field chemistry and general chemistry analyses

A YSI Professional Plus (Quatro) multi-parameter probe was used to measure the following field chemical parameters (Table 2): temperature, pH, and electrical conductivity. Specific conductivity (SpC; $\mu\text{S cm}^{-1}$ corrected to 25°C) was calculated by the meter ($\text{SpC} = \text{EC} \times 1.91$; where 1.91 is the temperature coefficient for waters at 25°C). Total dissolved solids (TDS; mg L^{-1}) was calculated by the meter ($\text{TDS} = \text{EC} \times 0.65$). Calibration of the YSI probe was conducted once per day and between springs having elevated TDS or pH values using a three-point calibration for pH and conductivity.

Samples of filtered spring water were collected in 250 mL Nalgene HDPE bottles and submitted to the Analytical Chemistry Laboratory at the New Mexico Bureau of Geology and Mineral Resources for chemical analyses (Table 3). Cations (Na^+ , K^+ , Mg^{2+} , Sr^{2+} , and Ca^{2+}) were measured using a PerkinElmer Optima 5,300 DV ICP-OES according to EPA 200.7. Anions (Br^- , Cl^- , F^- , NO_2^- , NO_3^- , PO_4^{3-} , and SO_4^{2-}) were measured using a Dionex ICS-5000 IC according to

EPA 300.0. Alkalinity was completed according to EPA 310.1 and silica concentrations were provided according to SM 1030E. Table 3 shows the detection limits of each solute and charge balance for each spring.

3.3 | Environmental isotopes used to infer residence times and mixing processes

3.3.1 | Tritium analyses

Tritium (^3H , $t_{1/2} = 12.32$ years) was analyzed on all spring samples in the Panamint Range to quantify young groundwater mean residence times ($< \sim 300$ years) and assess mixing processes. Samples of spring water were collected in 1,000 mL Nalgene HDPE bottles and analyzed by the University of Miami Tritium Laboratory using electrolytic enrichment and low-level counting (reported uncertainty of ± 0.1 TU).

TracerLPM (Jurgens, Bohlke, & Eberts, 2012), a lumped parameter model (Maloszewski & Zuber, 1982; Stewart & Morgenstern, 2016),

TABLE 2 Elevation and field chemistry of Panamint Range springs

Spring ID	Spring name	Date sampled	UTM 11S mE	UTM 11S mN	Elev. (mrsi)	Temp. (°C)	pH	SpC ($\mu\text{S cm}^{-1}$)	TDS (ppm)
PAN 1	Jail Spring	May 24, 2017	491,216	4,005,046	2,433.5	9.0	7.98	226.9	147.9
PAN 2	Thorndike Spring	May 25, 2017	493,294	4,009,949	2,336.6	9.8	7.66	382.1	247.7
PAN 3	Unnamed Panamint Spring E	May 26, 2017	484,864	3,987,455	963.2	18.8	8.31	970.0	650.5
PAN 4	Unnamed Panamint Spring F	May 26, 2017	483,892	3,987,614	803.5	19.4	7.30	1,010	715.0
PAN 5	Wheel Spring	May 26, 2017	499,209	4,019,868	747.7	22.6	8.30	777.0	507.0
PAN 6	High Noon Spring	May 27, 2017	493,756	4,018,743	1,418.5	17.3	7.94	2,234	1,449
PAN 7	Apron Spring	May 27, 2017	493,581	4,017,809	1,606.0	17.7	8.11	2,743	1,781
PAN 8	Main Hanaupah Spring # 2	May 28, 2017	496,928	4,004,586	1,264.9	15.1	8.54	265.4	172.9
PAN 9	Main Hanaupah Spring # 1	May 28, 2017	497,013	4,004,384	1,258.2	20.5	8.46	406.5	263.9
PAN 11	South Hanaupah Spring # 3	May 28, 2017	498,063	4,004,323	1,153.7	16.1	7.99	708.0	461.5
PAN 12	Wilson Spring	May 29, 2017	499,326	3,993,877	1,194.8	20.4	8.34	943.0	617.5
PAN 13	Lower Warm Spring A	May 30, 2017	506,301	3,980,186	754.7	34.4	7.64	654.0	422.5
PAN 14	Lower Warm Spring B	May 30, 2017	506,108	3,980,234	759.6	34.3	7.82	667.0	429.0
PAN 15	Uppermost Spring	May 31, 2017	496,410	4,012,068	1,632.5	16.5	8.45	796.0	455.0
PAN 16	Limekiln Spring	June 1, 2017	486,446	3,996,617	1,223.1	19.4	8.13	765.0	494.0
PAN 17	Unnamed Panamint Spring C	June 1, 2017	486,503	3,996,454	1,206.4	16.5	7.99	747.0	475.0
PAN 19	Warm Sulfur Spring	June 1, 2017	480,753	3,997,248	317.6	32.0	7.86	3,791	2,464
PAN 20	Post Office Spring	June 2, 2017	479,772	3,988,537	320.6	18.7	7.72	8,922	5,798
IES-019	Tule Spring (March 2016)	March 17, 2016	510,652	4,010,962	-77.4	26.8	7.54	5,127	3,335
IES-019	Tule Spring (March 2017)	March 14, 2017	510,652	4,010,962	-77.4	27.4	7.35	3,181	2074
IES-045	Upper Emigrant Spring	May 19, 2016	482,674	4,031,167	1,230.7	19.8	7.13	931.0	604.5
IES-047	Poplar Spring	March 13, 2017	482,637	4,013,420	1,224.8	17.7	7.40	1,070	695.5
Dry	Pistol Spring	May 25, 2017	496,607	4,012,059	-	-	-	-	-
Dry	Johnnie Shoshone Spring	May 25, 2017	493,533	4,011,415	-	-	-	-	-
Dry	Tarantula Spring	May 25, 2017	493,750	4,018,362	-	-	-	-	-
Dry	Unnamed Panamint Spring G	May 25, 2017	483,306	3,987,686	-	-	-	-	-

Note: "Elev" is the elevation of the spring emergence, "SpC" is specific electrical conductivity corrected to 25°C and "TDS" is total dissolved solids.

TABLE 3 General chemistry

Spring ID	Ca ²⁺ [0.05]	Mg ²⁺ [0.05]	K ⁺ [0.05]	Na ⁺ [0.05]	Sr ²⁺ [0.005]	Br ⁻ [0.1, 0.01]	Cl ⁻ [0.1]	F ⁻ [0.1]	SO ₄ ²⁻ [1.0]	HCO ₃ ⁻ [5.0]	Alk. [5.0]	Si [0.05]	C.B. (%)
PAN 1	59.9	5.66	2.42	9.24	0.173	0.024*	3.28	ND	67	158	130	16.1	-2.1
PAN 2	62.8	3.88	2.61	14.4	0.279	ND	7.44	ND	8.28	231	190	13.1	-0.45
PAN 3	146	41.1	5.62	35.1	0.811	0.20	30.9	0.19	347	258	211	28.5	0.02
PAN 4	125	41.7	5.78	35.6	0.800	0.19	32.5	0.14	352	191	156	26.2	-0.21
PAN 5	27.2	58.9	3.70	36.3	0.561	0.24	53.9	3.38	168	167	137	14.3	-0.29
PAN 6	352	136	8.63	39.7	3.00	0.171*	21.5	2.10	1,180	263	216	35.2	1.68
PAN 7	359	232	9.00	33.9	1.44	0.070*	18.3	2.83	1,510	349	286	38.8	1.01
PAN 8	35.4	11.1	1.44	4.42	0.108	0.011*	1.63	0.28	28.8	134	110	16.8	0.74
PAN 9	51.9	10.4	1.76	19.4	0.219	ND	6.95	1.67	49.2	191	156	27.4	-1.24
PAN 11	95.6	28.9	3.18	15.4	0.517	ND	6.70	0.80	137	302	247	37.2	-1.03
PAN 12	91.1	67.9	2.59	16.1	1.06	0.078*	11.7	0.24	282	271	222	20.6	1.07
PAN 13	64.0	21.4	3.21	33.5	0.810	0.13	25.9	0.55	179	130	107	34.2	-1.3
PAN 14	64.2	21.6	3.26	33.9	0.806	0.11	25.8	0.54	177	128	105	32.8	-0.55
PAN 15	40.2	72.6	2.85	23.0	1.85	ND	12.9	ND	41.0	485	397	32.7	-1.07
PAN 16	89.4	40.5	3.90	11.6	0.460	0.075*	8.35	0.97	214	224	184	21.3	-0.16
PAN 17	92.9	39.9	3.94	11.3	0.451	0.042*	7.57	1.19	204	257	210	20.7	-1.32
PAN 19	85.0	45.8	28.4	575	0.862	1.2	873	1.87	354	179	147	27.0	-1.86
PAN 20	720	202	59.9	1,170	8.78	2.2	1,520	0.86	2,420	537	440	34.7	1.51
IES-019	107	74.8	9.73	490	2.41	0.33	994	ND	141	115	94	26.2	0.26
IES-019	105	68.3	9.42	487	2.23	0.14	998	ND	143	109	89	27.5	-1.05
IES-045	77.4	58.4	1.07	50.3	0.89	ND	38.8	0.38	108	435	356	23.1	1.83
IES-047	190	58.6	9.21	58.3	1.68	0.17	23.9	0.89	503	326	267	32.2	1.63

Note: All solute concentrations are listed in mg L⁻¹, solute detection limits are shown in brackets, "ND" is non-detect, "Alk." is alkalinity as CaCO₃, and "C. B." is charge balance. *Bromide was analyzed at the 0.1 mg L⁻¹ detection. If water samples were non-detects for Br⁻ at 0.1 mg L⁻¹, then those samples were analyzed at a lower detection limit of 0.01 mg L⁻¹. Concentrations with an asterisk are measured at the low detection limit and ND are non-detects at the low detection limit.

was used to estimate mean residence times of springs (Table 4). Residence times were calculated using TracerLPM only for springs having ³H concentrations greater than 1.0 TU. An input time series for Modesto, CA (provided in TracerLPM) was used for all Panamint Range springs due to its proximity to the study area (~ 397 km). A dispersion-type response function was chosen because this function approximates the effects of advection and dispersion in natural flow systems (Kreft & Zuber, 1978; Maloszewski & Zuber, 1982) and it is most applicable where mixing occurs within the aquifer (Jurgens et al., 2012). A dispersion parameter (DP), representing the inverse Peclet number (PE = ratio of advection to dispersion), is optimized in the model. A low DP indicates mostly advective processes, while a high DP indicates mostly dispersive processes. Uncertainty was assessed by varying the measured tritium concentration according to its reported analytical uncertainty (±0.09 TU). Changing the upper bound on mean age by ±100 was found to vary the mean residence time by 5–20 years.

Tritium is most appropriate for age-dating groundwaters ranging from 50 years (Clark & Fritz, 1997) up to 100 (Cartwright & Morgenstern, 2015) and possibly 200 years (Stewart & Morgenstern, 2016). In principle, mixing models should be able to provide mean residence times in the 200–300 year range, but are complicated by: (1) an increasingly large number of tritium-dead years is required to re-create low ³H concentrations (< 0.3 TU) and (2) the

hysteretic nature of the bomb-pulse implies that there may be more than one plausible residence time for a ³H concentration immediately before and after the bomb-pulse era. Here, we compare the results of the TracerLPM with a recharge-weighted, steady-state, backward-in-time tritium mixing model that utilizes some of the features of the models described in Rose (1993) and Wade (2002). Recharge is estimated annually by:

$$R_i = \left(\text{random}_0^{0.20} \right) \times P_{\text{avg}} \quad (1)$$

where; R_i is the estimated recharge (cm) for year i , the average annual precipitation (P_{avg} , cm) is assumed to be equal to 48.6 cm (Gleason et al., 2019), and the scaling coefficient is a random percentage ranging from 0 (0%) to 0.20 (20%) to account for natural variability in recharge from year to year. The upper limit of the scaling coefficient (0.20) is equivalent to the average MBR calculated for the Panamint Range using a chloride mass-balance approach (Gleason et al., 2019). Cumulative recharge (R_j) is then calculated according to:

$$R_j = \sum_{2017}^j R_i \quad (2)$$

where; R_j is the cumulative (backward in time) annual recharge (cm; recharge is accumulated from 2017 backward in time to year j). A

TABLE 4 ^3H data and residence times calculated using TracerLPM. Please note that residence times were not calculated for springs with ^3H concentrations less than 1.0 TU. $^3\text{H}_{\text{unc}}$ is the reported analytical uncertainty. σ_{rt} is the estimated uncertainty in residence time found by running each sample in TracerLPM using its reported analytical uncertainty (± 0.09 TU). DP is the estimated dispersion parameter. The following parameterization was held consistent for all springs and model runs: 1) the unsaturated zone travel time = 0 years, 2) the lower bound on the mean age = 1 year and the upper bound = 400 years (assuming that springs with $^3\text{H} > 1.0$ have residence times less than 400 years), and 3) the lower and upper boundaries for the DP parameter = 0.01 and 0.99, respectively. The residence times and uncertainties shown in italics are calculated from the backwards in time, mixing model discussed in this paper. PAN 8 has the most uncertain residence time because a ^3H concentration of 2.42 occurs once before the bomb pulse and twice after the bomb pulse and is a consequence of mixing backwards in time with a hysteretic ^3H time series. In general, uncertainty increases as the ^3H approaches 0 TU because an increasingly large number of tritium-dead years are required to mix and reduce the ^3H concentration of the well-mixed aquifer

Spring ID	Spring Name	^3H (TU)	$^3\text{H}_{\text{unc}}$ (TU)	Residence Time (yrs)	σ_{rt} (yrs)	DP
PAN 1	Jail Spring	1.87	0.09	124 (155)	15 (3)	0.91
PAN 2	Thorndike Spring	1.35	0.09	216 (218)	24 (3)	0.92
PAN 3	Unnamed Panamint Spring E	0.11	0.09	-	-	-
PAN 4	Unnamed Panamint Spring F	0.19	0.09	-	-	-
PAN 5	Wheel Spring	0	0.09	-	-	-
PAN 6	High Noon Spring	0.07	0.09	-	-	-
PAN 7	Apron Spring	0.82	0.09	304 (375)	60 (5)	0.77
PAN 8	Main Hanaupah Spring # 2	2.42	0.09	65 (18, 52, 123)	30 (2, 1, 1)	0.51
PAN 9	Main Hanaupah Spring # 1	1.00	0.09	237 (296)	30 (7)	0.64
PAN 11	South Hanaupah Spring # 3	1.06	0.09	180 (282)	23 (5)	0.40
PAN 12	Wilson Spring	0.41	0.09	-	-	-
PAN 13	Lower Warm Spring A	0	0.09	-	-	-
PAN 14	Lower Warm Spring B	0	0.09	-	-	-
PAN 15	Uppermost Spring	0.45	0.09	-	-	-
PAN 16	Limekiln Spring	0.53	0.09	-	-	-
PAN 17	Unnamed Panamint Spring C	0.55	0.09	-	-	-
PAN 19	Warm Sulfur Spring	0.11	0.09	-	-	-
PAN 20	Post Office Spring	0.65	0.09	-	-	-
IES-019	Tule Spring (March 2016)	0	0.09	-	-	-
IES-019	Tule Spring (March 2017)	0	0.09	-	-	-
IES-045	Upper Emigrant Spring	0.02	0.09	-	-	-
IES-047	Poplar Spring	0.15	0.09	-	-	-

Note: Please note that residence times were not calculated for springs with ^3H concentrations less than 1.0 TU. $^3\text{H}_{\text{unc}}$ is the reported analytical uncertainty. σ_{rt} is the estimated uncertainty in residence time found by running each sample in TracerLPM using its reported analytical uncertainty (± 0.09 TU). DP is the estimated dispersion parameter. The following parameterization was held consistent for all springs and model runs: (1) the unsaturated zone travel time = 0 years, (2) the lower bound on the mean age = 1 year and the upper bound = 400 years (assuming that springs with $^3\text{H} > 1.0$ have residence times less than 400 years), and (3) the lower and upper boundaries for the DP parameter = 0.01 and 0.99, respectively.

decay-corrected ^3H concentration in recharge is calculated for year i according to:

$$T_i^* = T_i e^{-\lambda t} \quad (3)$$

where; T_i^* is the decay-corrected ^3H (TU) in precipitation for year i and is assumed to be equal to the decay-corrected ^3H in

recharge, T_i is the atmospheric ^3H for year i (we use the same atmospheric ^3H time series for Modesto, CA as used in TracerLPM), t is time elapsed since recharge calculated backward in time from 2017 to year j , and λ is the decay constant for ^3H ($0.05626 \text{ year}^{-1}$; see Rose, 1993). The calculated decay-corrected ^3H for each year i is then scaled by the calculated annual recharge according to:

$$T_{Ri} = R_i \times T_i^* \quad (4)$$

where; T_{Ri} is the recharge-scaled annual ^3H ($\text{TU} \cdot \text{cm}$) for year i . The recharge-scaled annual ^3H is then accumulated backward in time from 2017 to year j according to:

$$T_j = \sum_{2017}^j R_i T_i^* \quad (5)$$

The accumulated recharge-scaled ^3H ($\text{TU} \cdot \text{cm}$) through year j is then divided by the accumulated recharge (cm) through year j to provide a recharge-weighted ^3H concentration (T_{est}) in the well-mixed aquifer. The mean residence time (mrt) of the aquifer is then found by increasing the year j backward in time until T_{est} is equal to T_{meas} (the measured ^3H concentration of the spring). Thus, when T_{est} is equal to T_{meas} , then mrt is equal to $2017 - j$. An average mrt is calculated by running the model 50 times (i.e., generating 50 random sets of recharge percentages). This mixing model is relatively robust within the past 100 years; however, uncertainty increases with increasing residence time since an increasingly large number of tritium-dead years are required to mix and reduce the ^3H in the aquifer. We compare the results of the TracerLPM model with this mixing model to assess the uncertainty in ^3H residence times calculated between 100 and 300 years.

3.3.2 | Radiocarbon analyses

Radiocarbon (^{14}C , $t_{1/2} = 5,730$ years) samples were collected at all springs. However, the orifices of some springs were inaccessible, therefore samples were collected in the spring runs. These samples show evidence for atmospheric equilibration and are not presented here. Radiocarbon data are only shown for four springs where samples could be collected directly from the spring emergence: Jail Spring, Lower Warm Spring A, Lower Warm Spring B, and Tule Spring (Table 5). One liter of unfiltered spring water was collected in Nalgene HDPE bottles for ^{14}C and stable carbon (^{13}C) analyses. Bottles were tightly capped and taped closed. Radiocarbon analyses were completed by the University of Arizona AMS Laboratory. The analytical uncertainty for $\delta^{13}\text{C}$ was ± 0.1 and the analytical uncertainty for the percent modern carbon (pmC) ranged between ± 0.12 and 0.26 . Several radiocarbon correction models are available to correct measured radiocarbon activities (see Han & Plummer, 2016). We compared the results of three models: Fontes and Garnier (1979), Han and Plummer (2013, 2016), and Ingerson and Pearson Jr. (1964) using Netpath XL (Parkhurst & Charlton, 2008). The Ingerson & Pearson (referred to as *I&P* hereafter) is based solely on $\delta^{13}\text{C}$ mixing processes. In comparison, the Fontes and Garnier (*F&G*) and Han & Plummer (*H&P*) models both combine $\delta^{13}\text{C}$ mixing processes with a geochemical mass-balance that accounts for fractionation between gas, liquid, and

mineral phases of dissolved inorganic carbon. We chose these three models because (1) the *I&P* model is simple and it is nearly impossible to measure $\delta^{13}\text{C}$ in all the carbonate facies present in the Panamint Range (it allows us to use broad endmembers in the evolution of $\delta^{13}\text{C}_{\text{DIC}}$ as it moves through the system), and (2) the relatively more complex *F&G* and *H&P* models provide a more comprehensive way to account for the geochemical stoichiometry of the system based on what is known about the mineralogy of the geologic units. The data used to correct the water samples are provided in Table 5 and associated caption. Uncertainty in radiocarbon residence times was assessed by varying the $\delta^{13}\text{C}_{\text{soil}}$ term by $\pm 1\text{‰}$ since it is considered the most uncertain term in the correction model.

3.3.3 | Chlorine-36 analyses and creation of the $^{36}\text{Cl}/\text{Cl}$ chronometer

Chlorine-36 (^{36}Cl , $t_{1/2} = 301$ ka) was analyzed on all spring samples to quantify residence times of spring waters and identify the influence of mixing with brines and/or hydrothermal deposits (Table 6). Samples of filtered spring water were collected in 1,000 mL Nalgene HDPE bottles. Chlorine-36 ratios ($^{36}\text{Cl}/\text{Cl}$) were analyzed by the Purdue Rare Isotope Measurement Laboratory (PRIME Lab) using an accelerator mass spectrometer with a relative measurement uncertainty between $\pm 2.3\%$ and 4.5% .

The majority of springs could not be dependably age-dated using radiocarbon because they were sampled in the spring run. Many of these springs also had low ^3H concentrations (< 0.7 TU). Therefore, a $^{36}\text{Cl}/\text{Cl}$ chronometer was created by fitting a power-law trendline to the relationship between mean residence time (years) and measured $^{36}\text{Cl}/\text{Cl}$ ($\times 10^{-15}$) of seven springs that had acceptable ^3H and ^{14}C residence times (Figures S2–S4). The $^{36}\text{Cl}/\text{Cl}$ chronometer is given by: $\text{Residence time (years)} = 234,160 \times (^{36}\text{Cl}/\text{Cl})^{-0.878}$; $r^2 = 0.96$ (see Figures S3 and S4). The oldest residence times of the trendline were established using the ^{14}C residence times for Lower Warm Spring A and Lower Warm Spring B (Table 5). The youngest residence times of the trendline were established using ^3H residence times for Jail Spring, Thorndike Spring, Apron Spring, Main Hanaupah Spring # 1, Main Hanaupah Spring # 2, and South Hanaupah Spring # 3 (Table 4).

The robustness of the $^{36}\text{Cl}/\text{Cl}$ chronometer was assessed using Cl^-/Br^- ratios. The selection of a Cl^-/Br^- representative of groundwater “unaffected” by salts is dependent upon local geology, land-use, and anthropogenic activities (Davis, Whittemore, & Fabryka-Martin, 1998; Panno et al., 2006). Gleason et al. (2019) selected a conservative Cl^-/Br^- of 200; however, additional geochemical analyses indicate that unaffected groundwater in the Panamint Range has a Cl^-/Br^- less than 270. A total of 12 springs have Cl^-/Br^- ratios less than 270 (Davis et al., 1998). Springs in the study area that have Cl^-/Br^- greater than 270 are either influenced by mixing with basin brines, mixing with evaporite deposits in their spring run, or are likely impacted by hydrothermal deposits. The $^{36}\text{Cl}/\text{Cl}$ chronometer is site

TABLE 5 Carbon isotope data for selected springs ($\delta^{13}\text{C}$ and ^{14}C data and residence times for springs)

Spring ID	Spring name	pmC (uncertainty)	Measured $\delta^{13}\text{C}$ (‰)	Calculated $\delta^{13}\text{C}$ (‰)	I & P R.T. (years)	F & G R.T. (years)	H & P R.T. (years)*
PAN 1	Jail Spring	0.7654 (0.0026)	-9.1	-9.09	Modern (<150 years)	Modern (<150 years)	Modern (<150 years)
PAN 13	Lower Warm Spring A	0.3408 (0.0013)	-4.3	-4.30	1,409	1,421	1,412
PAN 14	Lower Warm Spring B	0.3028 (0.0012)	-4.8	-4.79	1,829	1,842	1,832
IES-019	Tule Spring (March 2016)	0.7763 (0.0028)	+0.1	-	Modern (<150 years)	-	-
IES-019	Tule Spring (December 2016)	0.5367 (0.0017)	-9.5	-	1,369	-	-
IES-019	Tule Spring (March 2017)	0.7373 (0.0021)	-12.6	-	1,078	-	-

Note: I&P represents Ingerson and Pearson, F&G represents Fontes and Garnier, and H&P represents Han and Plummer methods. Only two springs were suitable for radiocarbon analyses. The ^{13}C of soil CO_2 was assumed to range from -20‰ to -12‰ (Quade, Cerling, & Bowman, 1989). The ^{13}C of DIC in the bedrock for Jail Spring was assumed to range from -4‰ to +2‰ (Pettersen et al., 2011; Prave, 1999). The ^{13}C of DIC in the bedrock for Lower Warm Springs A and B was assumed to range from -2‰ to +3‰ (Brand, Webster, Azmy, & Logan, 2007). We think that the shaded sample for Tule Spring (March 2016, high pmC and high $\delta^{13}\text{C}$) is a result of the historic flooding event that occurred in Death Valley in October 2015. Alluvial recharge from the flood mixed with basin brines resulting in modern water at the spring. Rainfall would likely have a $\delta^{13}\text{C}$ of approximately -8.0‰ and the $\delta^{13}\text{C}$ of surface runoff generated from rain would be modified by interactions with carbonate rocks and soil. Thus, surface runoff could potentially have a $\delta^{13}\text{C}$ between -4‰ to +3‰ and +0.1‰ is not out of the range of possibilities. The residence times of Tule Spring were calculated using the Ingerson and Pearson approach assuming that the $\delta^{13}\text{C}$ of soil CO_2 was -15.0‰, a ^{14}C of soil = 100 pmC, and $\delta^{13}\text{C}$ of carbonate rock was 0‰.

*King and Bredehoeft (1999) provide uncorrected radiocarbon residence times for two springs not sampled in this study: Anvil Spring (UTM: 11S, 492,357 mE, 3,975,186 mN) and Dripping Spring (UTM: 11S, 496,410 mE, 4,019,396 mN). Anvil Spring emerges in Butte Valley which is geographically close to Lower Warm Spring A and Lower Spring B and had an uncorrected radiocarbon residence time of $2,710 \pm 40$ years. Dripping Spring is located near Wildrose Peak and had an uncorrected radiocarbon residence time of $6,430 \pm 40$ years. Comparisons with uncorrected radiocarbon data should be done cautiously.

specific, yet robust for the range of residence times (from 65 to 1,800 years).

Uncertainty in the calculation of $^{36}\text{Cl}/\text{Cl}_{\text{adj}}$

Adjusted $^{36}\text{Cl}/\text{Cl}$ ratios ($^{36}\text{Cl}/\text{Cl}_{\text{adj}}$) were calculated for the following springs: Warm Sulfur Spring and Upper Emigrant Spring ($^{36}\text{Cl}/\text{Cl}$ was not measured for these springs), Post Office Spring (the measured $^{36}\text{Cl}/\text{Cl}$ for this spring was heavily diluted by mixing with basin evaporites), and Thorndike Spring, Main Hanaupah Spring # 2, Main Hanaupah Spring # 2, South Hanaupah Spring # 3, and Uppermost Spring (all of these springs show evidence for ^{36}Cl dilution likely due to mixing with hydrothermal deposits). Therefore, $^{36}\text{Cl}/\text{Cl}_{\text{adj}}$ was estimated for these springs using: $^{36}\text{Cl}/\text{Cl}_{\text{adj}} = (2,796.3 \times ^3\text{H}) + 202.6$; $r^2 = 0.997$. The data are plotted in Figure S2. This is a linear regression. The uncertainty in the regression (σ_r) was calculated from:

$$\sigma_r = \sqrt{\frac{\sigma_{yy} - (m^2 - \sigma_{xx})^2}{N - 2}} \quad (6)$$

where,

$$\sigma_{yy} = \sum (y - y_{\text{avg}})^2 \quad (7)$$

$$\sigma_{xx} = \sum (x - x_{\text{avg}})^2 \quad (8)$$

$$m = \text{slope of linear regression} = \frac{\sigma_{xy}}{\sigma_{xx}} \quad (9)$$

$$\sigma_{xy} = \sum (x - x_{\text{avg}})(y - y_{\text{avg}}) \quad (10)$$

The estimated σ_r for this regression was 90.4. The uncertainty in the estimated y -value (σ_y) was calculated from (Montgomery & Runger, 2011):

$$\sigma_y = \sqrt{1 + \frac{1}{N} + \frac{(x - x_{\text{avg}})^2}{\sigma_{xx}}} \quad (11)$$

The estimated uncertainties for the calculated $^{36}\text{Cl}/\text{Cl}_{\text{adj}}$ values are shown in Table 6.

Uncertainty in the calculation of residence times from the $^{36}\text{Cl}/\text{Cl}$ chronometer

The $^{36}\text{Cl}/\text{Cl}$ chronometer is given by: $RT = 234,160 \times (^{36}\text{Cl}/\text{Cl})^{-0.878}$, $r^2 = 0.96$; where RT equals residence time in years. This equation is a power-law trendline (Figure S3) taking the form:

$$y = Bx^A \quad (12)$$

Thus, there is uncertainty in the ^{14}C residence times, ^3H residence times, and in the adjusted $^{36}\text{Cl}/\text{Cl}$ values shown above. These uncertainties surround the power-law trendline. We chose to linearize the equation and estimate the uncertainty as shown

TABLE 6 $^{36}\text{Cl}/\text{Cl}$ data and residence times estimated using the $^{36}\text{Cl}/\text{Cl}$ chronometer

Spring ID	Spring name	$^{36}\text{Cl}/\text{Cl} (\times 10^{-15})$	$^{36}\text{Cl}/\text{Cl}_{\text{unc}} (\pm ^{36}\text{Cl}/\text{Cl})$	Cl^-/Br^-	Residence time (years)	σ_{rt} (years) ^e
PAN 1	Jail Spring	5,487	98	137	122 ^d	39
PAN 2	Thorndike Spring	4,298 (3978) ^a	74 (108) ^b	1488 ^c	152 (162) ^d	47 (50)
PAN 3	Unnamed Panamint Spring E	608.5	14	155	843	273
PAN 4	Unnamed Panamint Spring F	600.1	19	171	853	277
PAN 5	Wheel Spring	258.5	8	225	1,787	614
PAN 6	High Noon Spring	327.4	12	126	1,452	504
PAN 7	Apron Spring	581.4 (2497) ^a	14 (98) ^b	261	877 (244) ^d	285 (75)
PAN 8	Main Hanaupah Spring # 2	2,885 (6970) ^a	55 (144) ^b	148	217 (99) ^d	67 (32)
PAN 9	Main Hanaupah Spring # 1	1,496 (2999) ^a	33 (100) ^b	1390 ^c	383 (208) ^d	117 (64)
PAN 11	South Hanaupah Spring # 3	1,479 (3167) ^a	33 (102) ^b	1340 ^c	387 (198) ^d	118 (61)
PAN 12	Wilson Spring	1,285	30	150	150	134
PAN 13	Lower Warm Spring A	271.2	9	199	1713 ^d	610
PAN 14	Lower Warm Spring B	271.2	8	235	1713 ^d	610
PAN 15	Uppermost Spring	877.3 (1461) ^a	22 (95) ^b	2580 ^c	611 (391)	192 (120)
PAN 16	Limekiln Spring	1,620	36	111	357	109
PAN 17	Unnamed Panamint Spring C	1,715	37	180	339	104
PAN 19	Warm Sulfur Spring	510 ^a	95 ^b	728	984	324
PAN 20	Post Office Spring	2020 ^a	96 ^b	688	294	90
IES-019	Tule Spring (March 2016)	26.94	1.2	3,012	–	–
IES-019	Tule Spring (March 2017)	23.18	1.5	7,129	–	–
IES-045	Upper Emigrant Spring	259 ^a	96 ^b	7760 ^c	1,784	639
IES-047	Poplar Spring	493.2	12	141	1,014	335

Note: The $^{36}\text{Cl}/\text{Cl}$ chronometer is given by: $RT = 234,160 \times (^{36}\text{Cl}/\text{Cl})^{-0.878}$, $r^2 = 0.96$; where RT equals residence time in years (see Figures S3).

^a $^{36}\text{Cl}/\text{Cl}$ was not measured for Warm Sulfur Spring and Upper Emigrant Spring. The $^{36}\text{Cl}/\text{Cl}$ for Post Office Spring was heavily diluted by mixing with basin brines and the ^{36}Cl for Thorndike Spring, Main Hanaupah Spring # 2, Main Hanaupah Spring # 2, South Hanaupah Spring # 3, and Uppermost Spring show evidence for dilution due to mixing with hydrothermal deposits. Therefore, an adjusted $^{36}\text{Cl}/\text{Cl}$ was estimated for these springs using the relationship between $^{36}\text{Cl}/\text{Cl}$ and ^3H ($^{36}\text{Cl}/\text{Cl}_{\text{adj}} = (2,796.3 \times ^3\text{H}) + 202.6$, $r^2 = 0.997$; see Figure S2). The estimated $^{36}\text{Cl}/\text{Cl}$ for Thorndike Spring, Main Hanaupah Spring # 2, Main Hanaupah Spring # 2, South Hanaupah Spring # 3, and Uppermost Spring is shown in italics.

^bThe uncertainties in the estimated $^{36}\text{Cl}/\text{Cl}$ are shown in italics while the analytical uncertainty is shown using standard font. The methods used to calculate the uncertainty of the regression (σ_r) between ^3H and $^{36}\text{Cl}/\text{Cl}$ and the uncertainties in the predicted $^{36}\text{Cl}/\text{Cl}$ ratios (σ_y) are presented in the following section.

^cThe bromide concentrations in these springs were non-detects at the 0.01 mg L⁻¹ level, so we used ½ the detection limit in the calculation of the Cl^-/Br^- ratios for these springs.

^dThese springs were used in the creation of the $^{36}\text{Cl}/\text{Cl}$ chronometer and are reported here for comparison with their ^3H or ^{14}C residence times. The residence times in italics are calculated from the $^{36}\text{Cl}/\text{Cl}$ corrected for the effects of mixing with basin brines, evaporites in the spring runs, and/or hydrothermal deposits.

^eThe methods used to calculate the uncertainty of the $^{36}\text{Cl}/\text{Cl}$ chronometer residence times (σ_{rt}) are presented in the following section.

below. While this is likely an underestimate of the true uncertainty, it has the same magnitude as the uncertainty calculated for a linear geochemical chronometer presented in Frisbee et al. (2013). The equation can be linearized according to:

$$\log(RT) = A \log x + \log B \quad (13)$$

This results in:

$$\log(RT) = -0.878[\log(^{36}\text{Cl}/\text{Cl})] + \log(234,160) \quad (14)$$

The uncertainty in the estimated y -value (σ_y), in this case residence time, can then be calculated using Equations ((6)–(10). Uncertainties in residence times are shown in Table 6.

3.4 | Environmental isotopes used to infer groundwater flowpaths

3.4.1 | Strontium isotope analyses

Strontium isotopic ($^{87}\text{Sr}/^{86}\text{Sr}$) analyses were performed on all spring samples to identify the rock units that host flowpaths in the Panamint Range (Blum, Erel, & Brown, 1994; Clow, Mast, Bullen, & Turk, 1997; Hogan & Blum, 2003). Water samples were collected in 125 mL Nalgene HDPE bottles. Whole rock samples were collected from geologic formations near spring emergences including: Noon-day Formation, Kingston Peak Formation, Johnnie Formation, and Bird Spring Formation. Strontium was leached from these rock samples by crushing the rocks to 5–10 mm and leaching the crushed

rock in 1-L closed cells of deionized water (DI; initial pH of 7) for 2 months at ambient room temperatures (21°C). Spring water samples and rock-leachate waters were analyzed with a Nu Plasma HR multi collector inductively coupled-plasma mass-spectrometer at the University of Illinois Urbana-Champaign isotope geochemistry laboratory. Sr^{2+} concentrations in spring water samples were reported in the geochemistry analyses from the Analytical Chemistry Laboratory at the New Mexico Bureau of Geology and Mineral Resources (see section 6.2). The $^{87}\text{Sr}/^{86}\text{Sr}$ of our rock-leachate samples are representative of more easily weathered strontium-bearing minerals (Bailey, McArthur, Prince, & Thirlwall, 2000; Frisbee et al., 2017). DI blanks (bottles filled with DI water for 2 months) did not yield strontium indicating that strontium was not leached from the bottle or present in the DI. Our $^{87}\text{Sr}/^{86}\text{Sr}$ dataset (Table 7) was supplemented using (1) rock-leachate $^{87}\text{Sr}/^{86}\text{Sr}$ data from Warix, Rademacher, Meyers, and Frisbee (2020); Table 8), (2) whole rock $^{87}\text{Sr}/^{86}\text{Sr}$ data from Wasserburg et al. (1964); Table 8), (3) $^{87}\text{Sr}/^{86}\text{Sr}$ data from Panamint Range springs reported in King and Bredehoeft (1999); Table 9), and (4) whole rock $^{87}\text{Sr}/^{86}\text{Sr}$ leachate (using 0.2 M nitric acid) reported in Paces et al. (2007); Table 10).

4 | RESULTS

4.1 | Field chemistry, general chemistry, and geochemical analyses

There is a general evolution from calcium-bicarbonate dominated groundwaters (quadrant I) at higher elevations to calcium-sulfate groundwaters (quadrant IV) at low elevations (Figure 5). All high-elevation and some mid-elevation springs (Jail Spring, Thorndike Spring, Main Hanaupah Spring # 1, Main Hanaupah Spring # 2, South Hanaupah Spring, Uppermost Spring, and Upper Emigrant Spring) are characterized as calcium-bicarbonate waters (quadrant I of Figure 5). The remainder of the mid-elevation and all low-elevation springs are calcium-sulfate waters plotting in quadrant IV of Figure 5. The two warm springs also plot within quadrant IV. Basin springs are sodium-chloride waters (quadrant II of Figure 5).

Spring waters fall along the trendline describing dedolomitization given by: $\text{CaMg}(\text{CO}_3)_2(\text{s}) + \text{CaSO}_4 \cdot 2\text{H}_2\text{O}(\text{s}) + \text{H}^+ \rightleftharpoons \text{CaCO}_3(\text{s}) + \text{Ca}^{2+} + \text{Mg}^{2+} + \text{SO}_4^{2-} + \text{HCO}_3^- + 2\text{H}_2\text{O}$ (Figure 6; Back et al., 1983; Bischoff, Juliá, Shanks, & Rosenbauer, 1994; Eberts & George, 2000; Fisher & Mullican III, 1997). Tule Spring (IES-019) is not plotted on Figure 6 since its chemistry is the result of mixing between basin brines (proposed by Li, Lowenstein, & Blackburn, 1997) and subsurface flow from Hanaupah Canyon (discussed in section 8.3).

4.2 | ^3H and $^{36}\text{Cl}/\text{Cl}$ results

^3H concentrations are generally greater than 1.5 TU at high elevations and less than 0.5 TU at low elevations (Table 4, Figure 7). A total of

10 springs were tritium dead (<0.1 TU; Table 4, Figure 7b). $^{36}\text{Cl}/\text{Cl}$ ratios conform to the same trend with elevation (Table 6, Figure 7c). For context, Davis et al. (2003) report a pre-anthropogenic (pre-bomb) $^{36}\text{Cl}/\text{Cl}$ ratio of 300–500 ($\times 10^{-15}$) for the region. All but six springs have $^{36}\text{Cl}/\text{Cl}$ ratios higher than 500 ($\times 10^{-15}$). Two separate trends emerge when ^3H is plotted as a function of $^{36}\text{Cl}/\text{Cl}$ (Figure 8). One trend represents Thorndike Spring, Uppermost Spring, Hanaupah Canyon springs, and Trail Canyon springs. The other trend represents the remainder of the mountain-block springs. Basin springs (Tule Spring, Post Office Spring, and Warm Sulfur Spring) and warm springs (Lower Warm Spring A and Lower Warm Spring B) show scatter in Figure 8 but generally have ^3H concentrations less than 1.0 TU and $^{36}\text{Cl}/\text{Cl}$ ratios less than 900×10^{-15} .

4.3 | Estimates of spring residence times

Mean residence times of Panamint Range springs range from 65 ± 30 years to $1,829 \pm 613$ years (Tables 4–6). In general, the highest elevation springs have the shortest residence times (Table 4): Jail Spring (124 ± 15 years), Thorndike Spring (216 ± 24 years), and Main Hanaupah Spring # 2 (65 ± 30 years). In comparison, the longest residence times are found in low-elevation springs (Tables 5 and 6): Lower Warm Spring A ($1,409 \pm 613$ years), Lower Warm Spring B ($1,829 \pm 613$ years), Wheel Spring ($1,787 \pm 641$ years), and Upper Emigrant Spring ($1,784 \pm 639$ years). Basin springs are not always the oldest springs, for example, Post Office Spring has a mean residence time of 294 ± 90 years while Warm Sulfur Spring has a mean residence time of 984 ± 324 years. Tule Spring has a highly variable mean residence time ranging from <150 years to $1,369 \pm 400$ years.

4.4 | $^{87}\text{Sr}/^{86}\text{Sr}$ Results

The $^{87}\text{Sr}/^{86}\text{Sr}$ of rock leachate samples are plotted along the y-axis of Figure 9 and in some cases, are bracketed where more than one sample was analyzed. Springs are plotted according to their $1/\text{Sr}^{2+}$ value. Springs which plot near a rock leachate sample are assumed to be flowing through rocks similar that of the rock leachate sample. Springs in Hanaupah Canyon (Main Hanaupah Spring # 1, Main Hanaupah Spring # 2, and South Hanaupah Spring # 3), Wilson Spring, springs in Warm Springs Canyon (Lower Warm Spring A and Lower Warm Spring B), Tule Spring, and Poplar Spring have $^{87}\text{Sr}/^{86}\text{Sr}$ ranging from 0.71000 to 0.72400. These springs appear to be associated with flow through the Kingston Peak Formation based on the similarity between their $^{87}\text{Sr}/^{86}\text{Sr}$ and that of the Kingston Peak Formation (Figure 9). One of the highest elevation springs (Thorndike Spring), all springs in Trail Canyon (Wheel Spring, High Noon Spring, and Apron Spring), Upper Emigrant Spring, and two basin springs (Warm Sulfur Spring and Post Office Spring) have $^{87}\text{Sr}/^{86}\text{Sr}$ ranging from 0.71600 to 0.73000. These springs appear to be associated with flow through the Johnnie Formation and the Noonday Dolomite (Figure 9). Two springs in Surprise Canyon (Limekiln Spring and Unnamed Panamint Spring C)

TABLE 7 $^{87}\text{Sr}/^{86}\text{Sr}$ data for springs analyzed in this study

Spring ID	Spring name	$^{87}\text{Sr}/^{86}\text{Sr}$	Uncertainty ($\pm$)	Sr^{2+} (mg L $^{-1}$)
PAN 1	Jail Spring	0.72875	0.00005	0.173
PAN 2	Thorndike Spring	0.72168	0.00005	0.279
PAN 3	Unnamed Panamint Spring E	0.73236	0.00005	0.811
PAN 4	Unnamed Panamint Spring F	0.73164	0.00005	0.800
PAN 5	Wheel Spring	0.72323	0.00005	0.561
PAN 6	High Noon Spring	0.72146	0.00005	3.00
PAN 7	Apron Spring	0.72479	0.00005	1.44
PAN 8	Main Hanaupah Spring # 2	0.71889	0.00005	0.108
PAN 9	Main Hanaupah Spring # 1	0.71579	0.00005	0.219
PAN 11	South Hanaupah Spring # 3	0.71674	0.00005	0.517
PAN 12	Wilson Spring	0.71418	0.00005	1.06
PAN 13	Lower Warm Spring A	0.71245	0.00005	0.810
PAN 14	Lower Warm Spring B	0.71249	0.00005	0.806
PAN 15	Uppermost Spring	0.73342	0.00005	1.85
PAN 16	Limekiln Spring	0.73373	0.00005	0.460
PAN 17	Unnamed Panamint Spring C	0.73311	0.00005	0.451
PAN 19	Warm Sulfur Spring	0.72393	0.00005	0.862
PAN 20	Post Office Spring	0.72373	0.00005	8.78
IES-019	Tule Spring (March 2016)	0.71701	0.00005	2.41
IES-019	Tule Spring (March 2017)	0.71705	0.00005	2.23
IES-045	Upper Emigrant Spring	0.72198	0.00005	0.89
IES-047	Poplar Spring	0.71645	0.00005	1.68

TABLE 8 $^{87}\text{Sr}/^{86}\text{Sr}$ data from rock leachates (RL) from this study and Warix et al. (2020) along with whole rock $^{87}\text{Sr}/^{86}\text{Sr}$ data from Wasserburg, Albee, and Lanphere (1964)

Sample ID	Rock unit	$^{87}\text{Sr}/^{86}\text{Sr}$	Uncertainty ($\pm$)	Data source
RL-BSF	Bird Spring Formation	0.710308	0.00005	This study
RL-KPF1	Kingston Peak Formation	0.716589	0.00005	This study
RL-KPF2	Kingston Peak Formation	0.713698	0.00005	This study
RL-JF	Johnnie Formation	0.722231	0.00005	This study
RL-ND	Noonday Dolomite	0.722918	0.00005	This study
CP-349	Hornblende diorite dike	0.7427	–	Wasserburg et al.
CP-263	Diorite sill	0.7415	–	Wasserburg et al.
CP-155a	Diorite dike	0.7273	–	Wasserburg et al.
CP-155b	Diorite dike	0.7298	–	Wasserburg et al.
CP-156	Hornblende diorite sill	0.7223	–	Wasserburg et al.
CP-151	Beck Spring Dolomite	0.7101	–	Wasserburg et al.
RL-SQ	Stirling Quartzite	0.711767	0.00005	Warix et al.
RL-ZQ	Zabriskie Quartzite	0.722366	0.00005	Warix et al.

and two springs in Pleasant Canyon (Unnamed Panamint Spring E and Unnamed Panamint Spring F) have $^{87}\text{Sr}/^{86}\text{Sr}$ ranging from 0.7200 to 0.74500. These springs appear to be associated with flow through the World Beater Complex or older rock units that crosscut the World Beater Complex. Two high-elevation springs (Jail Spring and Uppermost Spring) have anomalously high $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (0.72875 and 0.73342, respectively) based on the rocks present in their emergence (Figure 9).

5 | DISCUSSION OF RESULTS

5.1 | Identification of rock units hosting groundwater flowpaths

King and Bredehoeft (1999) sampled 10 Panamint Mountain springs and reported Sr^{+2} concentrations between 0.1811 and 1.2358 ppm (Table 9). King and Bredehoeft (1999) also measured ^{87}Sr

TABLE 9 The strontium solute concentrations (Sr^{2+}) and strontium isotopic data ($^{87}\text{Sr}/^{86}\text{Sr}$) for 10 Panamint Range springs that were sampled by King and Bredehoeft (1999) are compiled here for comparison

Spring name	UTM 11S mE	UTM 11S mN	Elev. (mrsI)	Sr^{2+} (ppm)	$\delta^{87}\text{Sr}$ (‰)	$^{87}\text{Sr}/^{86}\text{Sr}$
Anvil Spring (NPS # 123)	492,357.0	3,975,186.0	1,296.3	0.1811	16.15	0.7207
Burns # 1 Spring (NPS # 53)	483,135.8	4,027,730.1	1,596.8	0.8944	4.50	0.7124
Dripping Spring (NPS # 63)	496,409.5	401,395.5	1,157.9	1.0364	26.68	0.7281
Hummingbird Spring (NPS # 76)	490,288.3	4,008,308.5	2,194.6	0.3010	28.00	0.7291
Johnnie Shoshone Spring (NPS # 74)	493,610.9	4,011,201.6	2,194.6	0.3365 0.3291	21.50 21.60	0.7245 0.7245
Limekiln Spring* (NPS # NA)	485,700.0	3,996,422.0	1,236.9	0.4559	33.91	0.7333
Surprise Canyon Crk Spring (NPS # 248)	484,500.0	39,966,239.0	833.0	0.4540	34.03	0.7333
Thorndike Spring* (NPS # 75)	493,285.3	4,009,722.9	2,395.7	0.1297	17.30	0.7215
Upper Emigrant Spring* (NPS # 45)	482,544.0	4,030,966.0	1,257.9	1.1121	17.36	0.7215
Wildrose Spring* (NPS # 72)	482,584.0	4,013,281.0	1,197.9	1.2358	10.64	0.7168

Note: All data, except for the $^{87}\text{Sr}/^{86}\text{Sr}$ values, were published in King and Bredehoeft (1999). We calculated $^{87}\text{Sr}/^{86}\text{Sr}$ based on the following equation: $[(\delta^{87}\text{Sr}/1,000) + 1] \times ^{87}\text{Sr}/^{86}\text{Sr}_{\text{std}}$, where $^{87}\text{Sr}/^{86}\text{Sr}_{\text{std}}$ equals the $^{87}\text{Sr}/^{86}\text{Sr}$ standard for modern sea water (0.70920; Paces, Peterman, Futa, Oliver, & Marshall, 2007)). The springs with an asterisk denote springs which were also sampled in this study. Johnnie Shoshone Spring was visited during this study, but it was dry and obviously could not be sampled.

TABLE 10 The $^{87}\text{Sr}/^{86}\text{Sr}$ data for only the geologic units found in the Panamint Range spring are compiled here from Paces et al. (2007). Map symbols shown in parentheses indicate the symbology used by Workman et al. (2016)

Geologic map symbol	Stratigraphic rock unit	Epoch	Min. $^{87}\text{Sr}/^{86}\text{Sr}$	Max. $^{87}\text{Sr}/^{86}\text{Sr}$	Mean $^{87}\text{Sr}/^{86}\text{Sr}$
PMb (PIPb)	Bird Spring Formation—Undivided	Late Mississippian Early Permian	0.7073	0.7084	0.7079
Cn	Nopah Formation	Late Cambrian	0.7091	0.7092	0.7091
Cb	Bonanza King Formation	Middle and Late Cambrian	0.7088	0.7092	0.7091
Cc	Carrarra Formation	Early and Middle Cambrian	0.7085	0.7092	0.7087
CZw (Czw)	Wood Canyon Formation	Neoproterozoic and Early Cambrian	0.7079	0.7088	0.7084
Zs	Stirling Quartzite	Neoproterozoic	0.7055	0.7085	NR
Zj	Johnnie Formation	Neoproterozoic	0.7055	0.7085	NR
Zn	Noonday Dolomite	Neoproterozoic	0.7055	0.7085	NR
ZYb (Zb)	Beck Spring Dolomite	Neoproterozoic and Mesoproterozoic	0.7055	0.7085	NR
ZYc (Zc)	Crystal Spring Formation	Neoproterozoic and Mesoproterozoic	0.7055	0.7085	NR

concentration in these spring samples, but not ^{86}Sr . They correctly state that $^{87}\text{Sr}/^{86}\text{Sr}$ ratios are suitable for determining the source water for springs; however, the lack of ^{86}Sr concentrations limits interpretation of their data to broad correlations (note, we calculate the $^{87}\text{Sr}/^{86}\text{Sr}$ for their samples for easier comparison in Table 9). For example, higher Sr^{+2} is consistent with granitic rocks in bedrock highlands and the highest concentration of Sr^{+2} is from springs at or near salt pans.

The $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of Panamint Range springs in this study range from 0.71245 to 0.73373 (Table 7). The Sr^{+2} concentrations range from 0.108 to 8.78 ppm, which encompasses the range of concentrations found by King and Bredehoeft (1999). The combination of $^{87}\text{Sr}/^{86}\text{Sr}$ ratios from spring water and $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of rock leachate from this study, Warix et al. (2020) and Wasserburg et al. (1964); Table 8) allows us to improve the interpretation of the geologically complex spring flow and geochemistry from recharge zone to spring discharge.

FIGURE 5 Piper diagram showing solute trends for the Panamint Range springs. High-elevation springs (2,000–2,500 msl) are shown using blue symbols, mid elevation springs (1,000–2,000 msl) are shown using dark green symbols, low-elevation springs (500–1,000 msl) are shown using light green symbols, basin springs are shown using orange symbols, and warm springs are shown using orange-red symbols (this same color scheme is used throughout the paper unless noted otherwise). Quadrants are defined as follows: I = calcium-bicarbonate water, II = sodium-chloride waters, III = sodium bicarbonate waters, and IV = calcium-sulfate waters

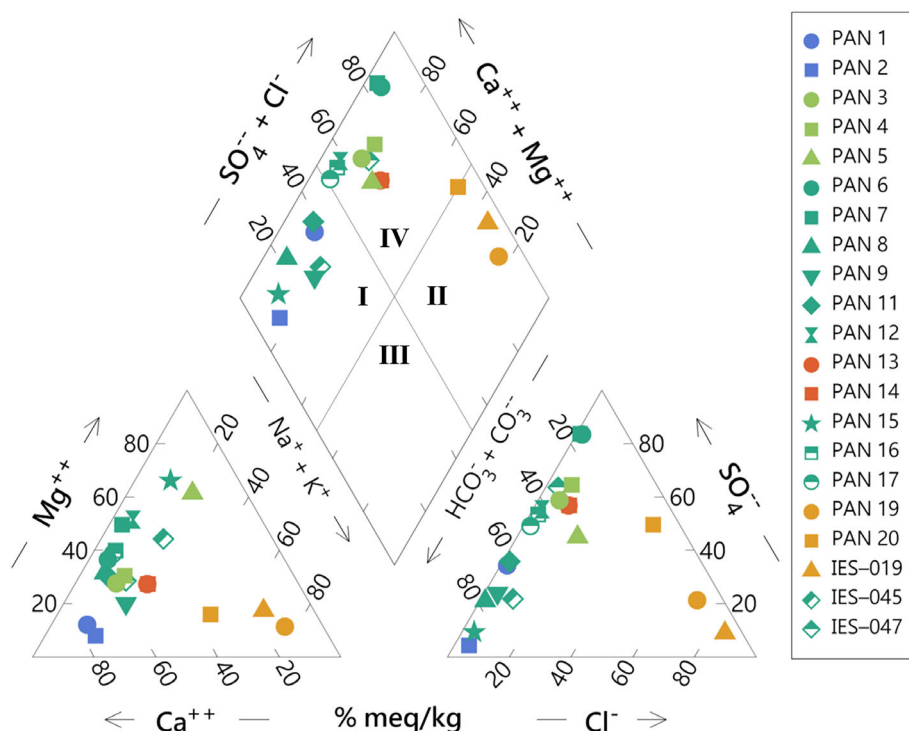
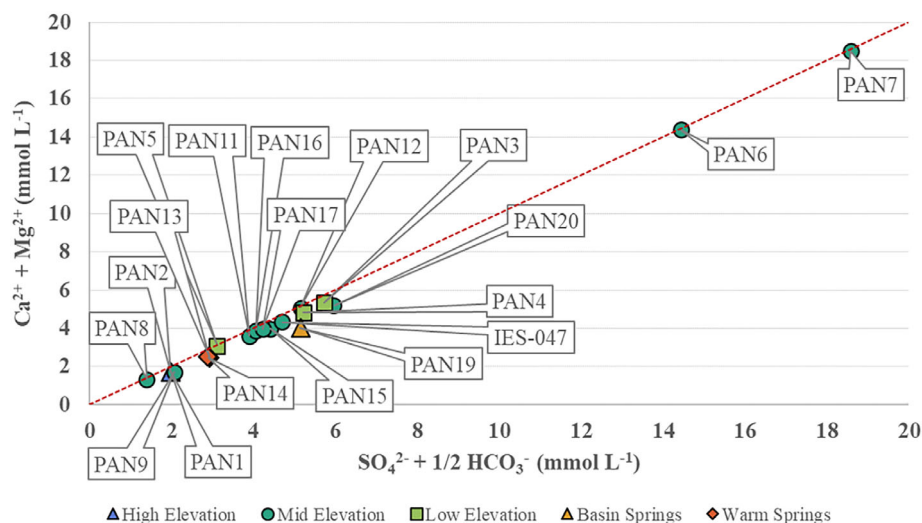


FIGURE 6 The red dotted line represents the expected 1:1 relationship for dedolomitization. IES-019 is not shown since it is a mixture of basin brines from Badwater Basin with MFR from Hanaupah Canyon. In general, high-elevation springs plot closer to the origin of the axes and low-elevation springs plot further from the origin; thus, illustrating the relationship between geochemical maturity and extent of dedolomitization. Although High Noon Spring (PAN 6) and Apron Spring (PAN 7) fall on this trendline, their sulfate concentrations are likely impacted by hydrothermal deposits. This inference is consistent with King and Bredehoeft (1999)



The Proterozoic Johnnie Formation ($^{87}\text{Sr}/^{86}\text{Sr} = 0.72223$), Noonday Dolomite ($^{87}\text{Sr}/^{86}\text{Sr} = 0.72291$) and Kingston Peak Formation ($^{87}\text{Sr}/^{86}\text{Sr} = 0.71659$) crop out along the Panamint Range crest (Figures 3 and 9; Albee et al., 1981). Jail Spring (PAN 1) and Thorndike Spring (PAN 2) are the simplest flow systems. Thorndike Spring is a high-elevation spring that emerges within the upper Johnnie Formation (Figure 3; Albee et al., 1981) with a residence time of 110 years (Table 4). The $^{87}\text{Sr}/^{86}\text{Sr}$ ratio for Thorndike Spring is nearly identical to our Johnnie Formation leachate (Figure 9), which indicates that the $^{87}\text{Sr}/^{86}\text{Sr}$ is derived from the Johnnie Formation. Jail Spring emerges at the contact between the lower member of the Johnnie Formation and the underlying Redlands Member of the Noonday Dolomite (Albee et al., 1981) and has a residence time of 86 years (Table 4). The

$^{87}\text{Sr}/^{86}\text{Sr}$ ratio is higher than either the Johnnie Formation or Noonday Dolomite leachates (Figure 9). We offer two hypotheses to explain the higher $^{87}\text{Sr}/^{86}\text{Sr}$ ratio: (1) unmapped igneous intrusions or (2) higher $^{87}\text{Sr}/^{86}\text{Sr}$ ratio in the lower member of the Johnnie Formation. Our preferred explanation is that the lower member of the Johnnie Formation has a higher $^{87}\text{Sr}/^{86}\text{Sr}$ ratio, but we cannot eliminate the possibility of unmapped igneous dikes.

Springs emerging on the western side of the Panamint Range tend to have the highest $^{87}\text{Sr}/^{86}\text{Sr}$ ratios. Unnamed Panamint Spring E (PAN 3) and Unnamed Panamint Spring F (PAN 4) emerge in the Kingston Peak Formation ($^{87}\text{Sr}/^{86}\text{Sr} = 0.713\text{--}0.716$) upstream of the South Park Canyon low-angle, normal fault with $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of 0.73236 and 0.73164, respectively. These springs are about 6 km

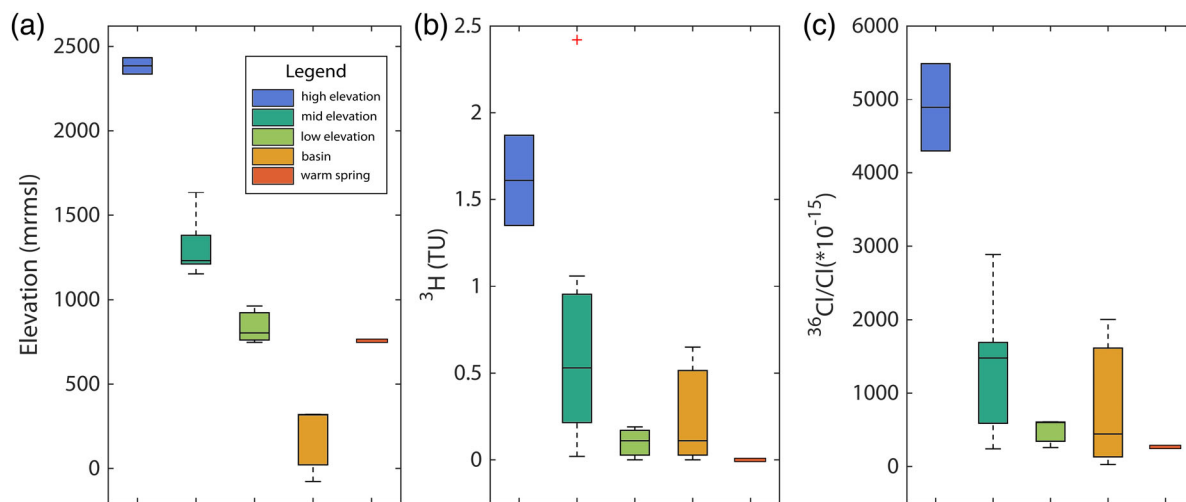


FIGURE 7 (a) Springs binned according to elevation. (b) ^3H concentrations of springs binned according to elevation, and (c) $^{36}\text{Cl}/\text{Cl}$ ratios binned according to elevation

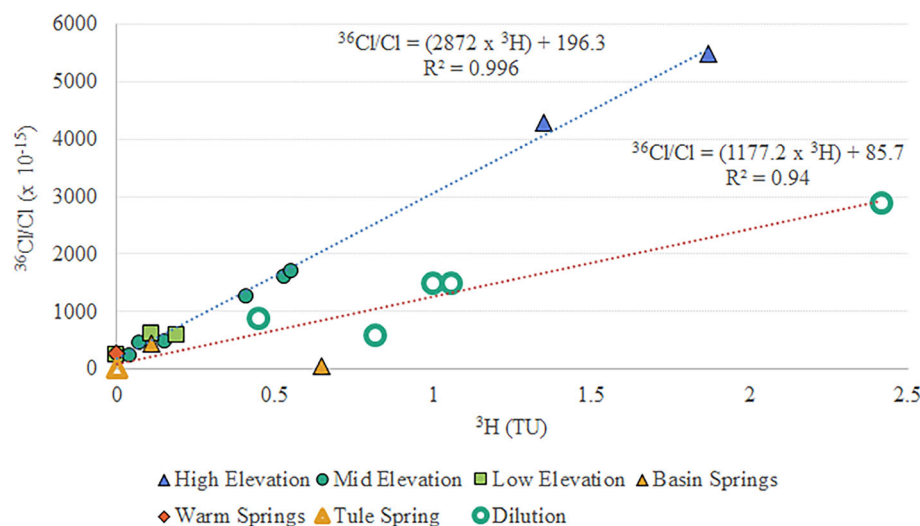


FIGURE 8 Two trends are present in the data: (1) Thorndike Spring, Uppermost Spring, Hanaupah Canyon springs, and Trail Canyon springs, and (2) all other springs emerging in the mountain block

downstream from the igneous intrusions sampled by Wasserburg et al. (1964) with $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of 0.7427 and 0.7415 (Table 8; Figure 9). Our interpretation is that the $^{87}\text{Sr}/^{86}\text{Sr}$ of these springs is the result of water–rock interaction with the igneous intrusions and the Kingston Peak Formation. The geologic conditions are similar at Limekiln Spring (PAN 16) and Unnamed Panamint Spring C (PAN 17) and we interpret the resulting $^{87}\text{Sr}/^{86}\text{Sr}$ ratios as due to a similar water–rock interaction system.

A geochemical continuum is present in Hanaupah Canyon springs (Figure 9). Main Hanaupah Spring # 2 (PAN 8) is the highest elevation sampled spring in the headwaters of Hanaupah Canyon and has the lowest Sr^{+2} concentration (highest $1/\text{Sr}^{+2}$). The $^{87}\text{Sr}/^{86}\text{Sr}$ ratio suggests contributions from both the Johnnie Formation and Kingston Peak Formation, both of which are in the headwater area. The Sr^{+2} concentration increases at the lower elevation South Hanaupah Spring # 3 (PAN 11) about 1 km down canyon. Although South Hanaupah Spring #3 emerges in the Noonday Dolomite, the

$^{87}\text{Sr}/^{86}\text{Sr}$ ratio is consistent with water–rock interaction with the Kingston Peak Formation. At the base of the Hanaupah Canyon alluvial fan, 14 km down canyon from Hanaupah Spring #3 is Tule Spring (IES-019). Tule Spring emerges at the contact between the Quaternary alluvium and the salt pan (Hunt & Mabey, 1966) and has a $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of 0.71401 and Sr^{+2} concentration of 2.23–2.41 mg L^{-1} . The higher Sr^{+2} concentration suggests dissolution of salt pan minerals; however, the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio is similar to the Hanaupah Canyon springs (Figure 9). We interpret the relatively consistent $^{87}\text{Sr}/^{86}\text{Sr}$ ratio down Hanaupah Canyon to Tule Spring as an indication that the $^{87}\text{Sr}/^{86}\text{Sr}$ is set by water–rock interaction with the Johnnie and Kingston Peak Formations. This $^{87}\text{Sr}/^{86}\text{Sr}$ ratio persists from the upper Hanaupah Canyon springs to Tule Spring at the toe of the fan. These data and inferred flowpath connectivity are consistent with the 162–310-year residence times for springs in Hanaupah Canyon (Table 6) and the >1,429-year residence time for Tule Spring (Table 5).

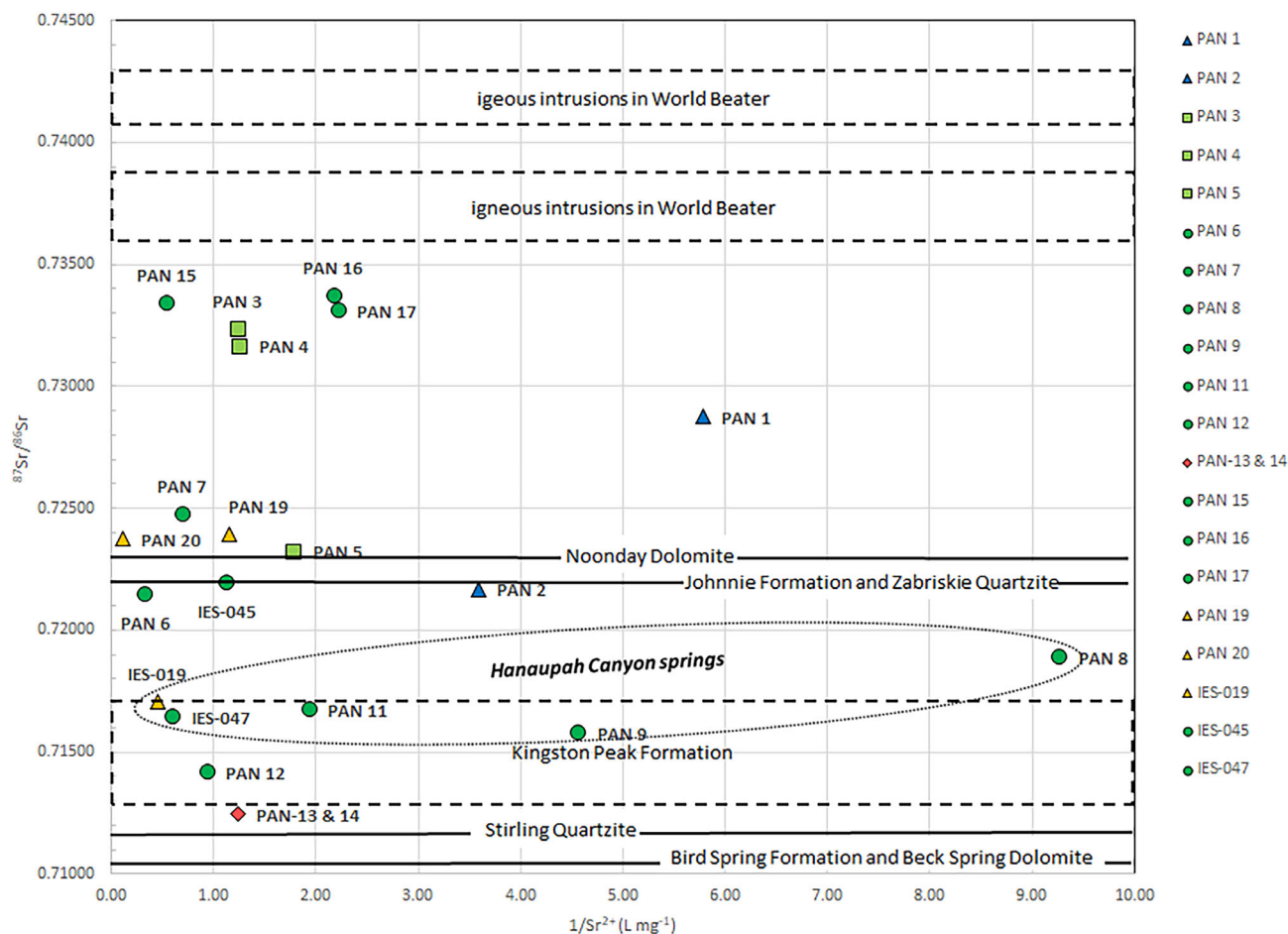


FIGURE 9 Relationships between the $^{87}\text{Sr}/^{86}\text{Sr}$ of springs and rock leachate samples in the Panamint Range. The $^{87}\text{Sr}/^{86}\text{Sr}$ of rock leachate samples are shown as bold lines or bracketed ranges. The Hanaupah Canyon to Tule Spring flow path is shown as a dotted oval

Two $^{87}\text{Sr}/^{86}\text{Sr}$ ratios were measured for Tule Spring: 0.71701 in March 2016 and 0.71705 in March 2017. Tule Spring, as mentioned previously, emerges at the base of the Hanaupah Canyon alluvial fan at the transition from fan to basin sediments. Gleason et al. (2019) developed a conceptual model for Tule Spring where it is supported by a mixture of two groundwater sources: (1) MFR from Hanaupah Canyon whereby surface water flowing from Hanaupah Canyon recharges at the mountain-front on the Hanaupah Canyon alluvial fan and then flows through the alluvium toward the basin, and (2) basin brines associated with Badwater Basin mixed with alluvial recharge from the Amargosa River when it floods. The slightly lower $^{87}\text{Sr}/^{86}\text{Sr}$ ratio measured in March 2016 may reflect the effects of mixing with recent flood waters since rainfall typically has low $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (<0.71000 ; see: Sherman, Blum, Dvonch, Gratz, & Landis, 2015), although $^{87}\text{Sr}/^{86}\text{Sr}$ data for rainfall are sparse.

Many of the other springs clearly show that water–rock interaction is the source of the $^{87}\text{Sr}/^{86}\text{Sr}$ ratios. The springs of Trail Canyon (Apron Spring [PAN 7], High Noon Spring [PAN 6], and Wheel Spring [PAN 5]) on the eastern side of the Panamint Range have $^{87}\text{Sr}/^{86}\text{Sr}$ ratios that bracket the $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of the Johnnie Formation and Noonday Dolomite (Figure 9) found in the headwaters. Similarly,

Poplar Spring (IES-047) emerges in an area of brecciated Kingston Peak Formation and the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio is consistent with that source rock. Wilson Spring (PAN 12), in Johnson Canyon of the eastern Panamint Mountains, has an $^{87}\text{Sr}/^{86}\text{Sr}$ ratio consistent with groundwater derived from the Beck Spring Dolomite, Johnnie Formation and Noonday Dolomite bedrock (Figure 9) found in the headwaters.

The Lower Warm Springs (PAN 13 and PAN 14) in the southern Panamint Mountains have the lowest $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (0.71245 and 0.71249, respectively) in this study. These springs emerge along the inferred trace of the Butte Valley Thrust Fault in the limestone and dolomite of Bird Spring Formation (Wrucke et al., 2007). The 34°C temperature (Table 2) and 1,624-year residence time (Table 5) indicate deeply circulating water. We interpret the $^{87}\text{Sr}/^{86}\text{Sr}$ ratios as reflective of water–rock interaction between the Proterozoic bedrock (i.e., Kingston Peak Formation) and Bird Spring Formation. This interpretation is tentative considering the local geology consists of Mesozoic plutonic rocks along with the extensively faulted Bird Spring and older bedrock units.

Interpretation of the $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of several springs is tentative. Warm Sulfur Spring (PAN 19) and Post Office Spring (PAN 20) emerge on the Panamint Valley floor along the Panamint Valley fault

zone. Post Office Spring has relatively high Sr^{2+} , which we interpret as evidence of water interaction with evaporite deposits in the valley floor deposits. Gleason et al. (2019) inferred that Post Office Spring was a mixture of two groundwater sources: (1) MFR on the Pleasant Canyon alluvial fan and (2) mountain-system recharge which is brought to the surface at a fault at the toe of the fan. This spring has a short residence time of 294 years which seems to support the inferred groundwater mixture. It also has a $^{87}\text{Sr}/^{86}\text{Sr}$ of 0.72373 which may reflect flow or mixing of flowpaths through older geologic units. In comparison, Warm Sulfur Spring emerges at a mountain-front fault and Gleason et al. (2019) inferred that it was primarily supported by mountain-system recharge since it is not associated with an alluvial fan. This spring has a residence time of 1,006 years (Table 5) and likewise, seems to support the inference of Gleason et al. (2019). The 32°C temperature (Table 2) of Warm Sulfur Spring suggests a complex, deep circulation of water circulation, possibly along the fault zone. This sort of water-rock interaction at the spring site and flow along the fault zone makes interpreting the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio (0.72393) difficult at best.

Upper Emigrant Spring (IES-045) emerges along the fault contact between Neogene conglomerates of the Nova Formation and the Skidoo Granite (Hunt & Mabey, 1966), yet the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio (0.72198) is consistent with the Proterozoic Johnnie Formation, Noonday Dolomite and Zabriskie Quartzite. We interpret this $^{87}\text{Sr}/^{86}\text{Sr}$ ratio as reflecting water interaction with Proterozoic clasts within the Nova Formation. Hunt and Mabey (1966) noted that the Nova Formation clasts consist of these Proterozoic rocks.

We have no clear explanation of the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio and Sr^{2+} concentration of Uppermost Spring (PAN 15) in Death Valley Canyon. The $^{87}\text{Sr}/^{86}\text{Sr}$ ratio (0.73342) is similar to groundwater that interacts with igneous intrusions into the World Beater Complex; however, these rocks are not mapped within the drainage basin. The drainage basin is underlain by Johnnie Formation, Zabriskie Quartzite and Stirling Quartzite (Hunt & Mabey, 1966). Water interaction with these bedrock units should produce a significantly lower $^{87}\text{Sr}/^{86}\text{Sr}$ ratio.

5.2 | Geochemical processes responsible for the geochemistry of springs

All springs in the study area, except for Tule Spring, fall along the geochemical trendline describing dedolomitization (Figure 6) indicating that groundwater flowpaths are hosted in dolomitic rocks. In principle, progressive increases in both $\text{Mg}^{2+}/\text{Ca}^{2+}$ ratios and SO_4^{2-} concentrations of spring waters (due to the dissolution of dolomite and gypsum with concurrent precipitation of calcite) should be observed as residence time increases (Figure 6). Jail Spring (PAN 1), Thorndike Spring (PAN 2), Main Hanaupah Spring # 1 (PAN 9), Main Hanaupah Spring # 2 (PAN 8), South Hanaupah Spring # 3 (PAN 11), Uppermost Spring (PAN 15), and Upper Emigrant Spring (IES-045) are calcium-bicarbonate waters (quadrant I of Figure 5). These springs tend to have shorter residence times, with the exception of Uppermost Spring (Table 6), and plot at the lower concentration end of Figure 6. The

remaining mountain-block springs are calcium-sulfate waters (quadrant IV of Figure 5), tend to have longer residence times (Table 6), and plot at the upper concentration end of Figure 6. Dedolomitization is supported by the $^{87}\text{Sr}/^{86}\text{Sr}$ data which indicate that the Kingston Peak Formation, Johnnie Formation, and Noonday Dolomite (Figure 9) are the primary formations hosting groundwater flowpaths. Each formation contains dolomitic units (Hunt & Mabey, 1966).

Springs in Trail Canyon, Hanaupah Canyon, Uppermost Spring, and Thorndike Spring appear to show the effects of mixing with hydrothermal deposits. Thorndike Spring, Main Hanaupah Spring # 1, South Hanaupah Spring # 3, and Uppermost Spring have Cl^-/Br^- greater than 1,340 (Table 6) implying that there is an additional source chloride to the springs. Apron Spring, Main Hanaupah Spring # 2, Main Hanaupah Spring # 1, South Hanaupah Spring # 3, and Uppermost Spring have low $^{36}\text{Cl}/\text{Cl}$ (lower than predicted by their ^3H concentration; Figure 8) consistent with $^{36}\text{Cl}/\text{Cl}$ dilution. Although High Noon Spring and Apron Spring both plot on the dedolomitization trendline, their sulfate concentrations are extremely high (1,180 and 1,510 mg L^{-1} , respectively) compared with other springs. King and Bredehoeft (1999) inferred that high sulfate concentrations of their high-elevation springs were affected by hydrothermal deposits since small relic mining operations were also found in close proximity to those springs. Mineralization occurs along detachment faults (Reynolds & Lister, 1987) when hot, reducing fluids flowing in the lower plate come into contact with metal-rich, warm, oxidizing fluids of the upper plate along brecciated rocks associated with the fault (Long, 2004; Michalski, Reynolds, Niles, Sharp, & Christensen, 2007). One explanation for these low $^{36}\text{Cl}/\text{Cl}$ ratios is that groundwater in Trail Canyon, Hanaupah Canyon, and along the crest near Uppermost Spring and Thorndike Spring has exchanged with metal-chloride complexes found in hydrothermal deposits (Helgeson, 1970; Murphy, 1929). Old chloride associated with these metal-chloride deposits would tend to dilute (decrease) the $^{36}\text{Cl}/\text{Cl}$ of water interacting with it. However, the dissolution rates of these metal complexes are low and would likely require long residence times.

There is anecdotal evidence that the Reese River Process was used to extract ore from rock in mining operations located in the Panamint Range (Murphy, 1929) and this may provide another explanation for the excess salt. During the Reese River Process, crushed ore-bearing rock is roasted in salt (likely transported from Searles Valley; Murphy, 1929) and combined with copper sulfate to form copper chloride. The copper chloride, in turn, combines with silver to form silver chloride, and the silver chloride combines with mercury to form an amalgam of silver-mercury (Oberbillig, 1967). Tons of chloride were likely transported into the mining towns of the Panamint Range (Oberbillig, 1967). Phillips, Smith, Bentley, Elmore, and Gove (1983) report $^{36}\text{Cl}/\text{Cl}$ ratios of $55 (\times 10^{-15}) \pm 2.75 (\times 10^{-15})$ for halite taken from the upper 14.2 m of Searles Lake. Thus, it's plausible that residual chloride contamination around the spring runs near these ghost towns may explain the elevated Cl^-/Br^- and the observed ^{36}Cl dilution, especially in the lower reaches of Hanaupah Canyon and Trail Canyon where mining relics are present today. However, it's unlikely that tons of salt were transported to high-elevation springs. A third

explanation is that salts from the surrounding playas (including Searles Lake) are transported into the high elevations of the Panamint Range by wind; however, aeolian salt deposition would likely affect all high-elevation springs (unless there are topographic features which enhance deposition near one spring and not another). Regardless, dry deposition data are not available to test this explanation at this time.

5.3 | Residence times of springs

Mean residence times range from 65 ± 30 years (Main Hanaupah Spring # 2) to $1,829 \pm 613$ years (Lower Warm Spring B). A total of 11 springs have mean residence times less than 500 years (Tables 4–6), three mid-elevation springs have mean residence times ranging from 843 ± 277 to 984 ± 324 years, and the remainder of springs ($n = 7$) have mean residence times greater than 1,000 years (Table 6). In general, the shortest residence times are found at the highest elevations, and the longest at low elevations.

Basin springs do not have the longest residence times. Post Office Spring (249 ± 90 years) and Warm Sulfur Spring (984 ± 324 years) emerge on the western side of the Panamint Range. Post Office Spring emerges at a fault along the base of the Pleasant Canyon alluvial fan. We infer that it has a short residence time due to mixing between potentially old groundwater supported by MSR brought to the surface by the fault and modern groundwater recharge occurring on the Pleasant Canyon alluvial fan (MFR). Warm Sulfur Spring, in comparison, emerges at a mountain-front fault and does not emerge near an alluvial fan. It is supported primarily by MSR, hence the longer mean residence time.

Tule Spring emerges at the base of the Hanaupah Canyon alluvial fan (bajada); it is tritium dead and its $^{36}\text{Cl}/\text{Cl}$ is heavily diluted due to mixing with basin brines. The residence time of this spring is variable (<150 years to $1,369 \pm 400$ years). The conceptual flow model for Tule Spring developed by Gleason et al. (2019) indicates that Tule Spring is supported by two different groundwater sources: (1) seasonal runoff from Hanaupah Canyon which recharges at the mountain front on the alluvial fan draining Hanaupah Canyon (MFR), and (2) basin brines which mix with alluvial recharge from periodic flooding and runoff in Badwater Basin from the Amargosa River. For example, the $\delta^{13}\text{C}$ was $+0.1\text{‰}$ and the pmC was 0.7763 in March 2016 after a record winter snowpack and a historic flooding event in October 2015. For context, Scotty's Castle in Death Valley National Park was damaged by the flooding event in October 2015 (see Wines, 2019). However, 1 year after the flooding event, the $\delta^{13}\text{C}$ decreased to -9.5‰ and the pmC decreased to 0.5367 in December 2016. In March 2017, the $\delta^{13}\text{C}$ had decreased to -12.6‰ and the pmC had increased to 0.7373. The radiocarbon residence times for Tule Spring are modern (<150 years) in March 1, 2016, 369 ± 400 years (December 2016), and $1,078 \pm 400$ years (March 2017). Tule Spring has the lowest $^{36}\text{Cl}/\text{Cl}$ (23.18×10^{-15}) of all Panamint Springs (Table 6). This provides partial support for the inference of Li et al. (1997) and Gleason et al. (2019) that Tule Spring discharges groundwater that has mixed with basin brines, since basin brines

commonly have very low $^{36}\text{Cl}/\text{Cl}$ ratios and tend to dilute the $^{36}\text{Cl}/\text{Cl}$ of fresh water when brines mix with fresh waters (Phillips, 2000). However, the $^{87}\text{Sr}/^{86}\text{Sr}$ data for Tule Spring closely matches the $^{87}\text{Sr}/^{86}\text{Sr}$ of Hanaupah Canyon springs. This supports our inference that MFR from Hanaupah Canyon mixes with basin brines in Badwater Basin to support flow to Tule Spring.

Thorndike Spring, Uppermost Spring, springs in Hanaupah Canyon (Main Hanaupah Spring # 2, Main Hanaupah Spring # 1, and South Hanaupah Spring # 3), and springs in Trail Canyon (Wheel Spring, High Noon Spring, and Apron Spring) have lower $^{36}\text{Cl}/\text{Cl}$ relative to the other springs (as predicted by their ^3H concentrations; Figure 8). To correct for the observed $^{36}\text{Cl}/\text{Cl}$ dilution observed in these springs, the relationship between $^{36}\text{Cl}/\text{Cl}$ and ^3H (for unaffected springs) was used to calculate an adjusted $^{36}\text{Cl}/\text{Cl}_{\text{adj}}$ given by: $^{36}\text{Cl}/\text{Cl}_{\text{adj}} = (2,796.3 \times ^3\text{H}) + 202.6$; $r^2 = 0.997$, $p < .01$ at the 95% confidence interval (Figure S2). Adjusted $^{36}\text{Cl}/\text{Cl}$ data are shown in italics in Table 6 along with their associated residence times. Since Thorndike Spring, Main Hanaupah Spring # 2, Main Hanaupah Spring # 1, South Hanaupah Spring # 3, and Uppermost Spring had reliable ^3H residence times (Table 6), their $^{36}\text{Cl}/\text{Cl}_{\text{adj}}$ values were then used to provide additional calibration points in the $^{36}\text{Cl}/\text{Cl}$ chronometer (Figures S3 and S4). In the absence of additional sources of chloride from evaporites, brines, and hydrothermal deposits, the relationship between $^{36}\text{Cl}/\text{Cl}$ and ^3H should be strongly linear as indicated by the fit (high r^2 and very low p -value) of the linear regression for unaffected springs. Thus, we consider the mean residence times calculated using the $^{36}\text{Cl}/\text{Cl}$ chronometer to be robust with the $^{36}\text{Cl}/\text{Cl}_{\text{adj}}$ values.

The basin springs are extremely difficult to age-date. Tule Spring is tritium dead and has a $^{36}\text{Cl}/\text{Cl}$ modified by mixing with basin brines. However, three radiocarbon correction methods provided similar residence times for Tule Spring (Table 5). Using the Ingerson and Pearson Jr. (1964) radiocarbon correction model, we calculated radiocarbon residence times ranging from modern (<150 years) to $1,369 \pm 400$ years (Table 5). A radiocarbon sample was not collected for Post Office Spring since it was sampled in the spring run, and the $^{36}\text{Cl}/\text{Cl}$ sample was diluted by mixing with evaporites in its spring run. Warm Sulfur Spring was not sampled for $^{36}\text{Cl}/\text{Cl}$ and its radiocarbon was suspect. Therefore, $^{36}\text{Cl}/\text{Cl}_{\text{adj}}$ ratios were also calculated for these two springs. Using the $^{36}\text{Cl}/\text{Cl}_{\text{adj}}$ ratios, we calculate $^{36}\text{Cl}/\text{Cl}$ chronometer residence times for Warm Sulfur Spring and Post Office Spring of 984 ± 324 and 294 ± 90 years, respectively.

5.4 | Role of faults in hydrogeological processes in the Panamint Range

Faults affect flow and geochemical processes in the Panamint Range. Seventeen of twenty-one sampled springs in the Panamint Range emerge at faults; 13 springs emerge at low-angle normal faults or detachment faults (Figures 3 and S1). The $^{87}\text{Sr}/^{86}\text{Sr}$ data, geochemical data, and field observations all indicate that the Kingston Peak Formation, Johnnie Formation, and Noonday Dolomite are the primary formations hosting groundwater flowpaths. Detachment surfaces have

been mapped in the Noonday Dolomite and Johnnie Formation as well (Hunt & Mabey, 1966; Norton, 2011) and a detachment surface is well exposed in the lower reaches of Hanaupah Canyon. We infer that groundwater flowing along damage zones is brought to the surface where low-angle faults intersect the nearly vertical geologic units on the eastern side of the Panamint Range. In comparison, groundwater flows through fractured Paleoproterozoic bedrock and is brought to the surface where low-angle faults intersect Cenozoic breccias and fanglomerates on the western side of the Panamint Range. Mineralization associated with these faults likely influences the geochemical composition of springs emerging in Hanaupah Canyon and Trail Canyon.

6 | CONCLUSIONS

This study was conducted to address the following questions: (1) which rock units support groundwater flow to springs in the Panamint Range, (2) what are the geochemical kinetics of these aquifers, and (3) what are the residence times of springs in the Panamint Range? With respect to Question 1, $^{87}\text{Sr}/^{86}\text{Sr}$ data indicate that the Kingston Peak Formation, Noonday Dolomite, and Johnnie Formation are critical to recharge and groundwater flow. Springs on the western side of the Panamint Range are also supported by groundwater flow through fractured Paleoproterozoic metamorphic and igneous rocks that crop out broadly on the western flank. Groundwater may circulate down to the brecciated detachment surface along the mylonitic gneiss of the World Beater Complex, but it's unlikely that it circulates deep within the gneiss.

With respect to Question 2, all springs are supported to some extent by flow through dolomitic rocks. Dedolomitization is the dominant weathering reaction. However, interactions with metal-chloride complexes found in hydrothermal deposits in Trail Canyon, Hanaupah Canyon, and along the crest of the Panamint Range near Thorndike Spring and Uppermost Spring also impact the geochemical composition of those springs.

With respect to Question 3, spring residence times range from 65 ± 30 years (Main Hanaupah Spring # 2) to $1,829 \pm 613$ years (Lower Warm Spring B). A total of 11 springs have mean residence times less than 500 years. Seven springs have mean residence times greater than 1,000 years (Table 6). In general, the shortest residence times are found at the highest elevations, and the longest at low elevations. Basin springs show wide variability in mean residence times as a function of their hydrogeological connection to: (1) MFR occurring on alluvial fans or bajadas (Tule Spring and Post Office Spring), (2) groundwater discharge at mountain-front faults (Warm Sulfur Spring), and (3) basin brines (Tule Spring). The relatively short residence times documented here support a groundwater model where groundwater circulation depth is likely controlled by the depth of the detachment surface since many of these springs emerge at low-angle normal or detachment faults (Albee et al., 1981). For context, the springs of Ash Meadows National Wildlife Refuge, located east of Death Valley National Park, are likely supported by regional

groundwater flow through the DVRFS and these springs have residence times ranging from 5,000 to 25,000 years (Anderson, Nelson, Mayo, & Tingey, 2006). In comparison, a majority of springs ($n = 11$) of the Panamint Range have relatively short mean residence times less than 500 years. Thus, while the role of faults on groundwater flow and transport processes has received considerable attention in the hydrogeologic literature (for reviews, see Gudmundsson, 2000; Bense, Gleeson, Loveless, Bour, & Scibek, 2013), the role of detachment faults in groundwater flow has received comparatively less attention (exceptions are Piaschky, 2007; Morealli, 2010). Our data indicate that the low-angle faults in the Panamint Range are highly permeable. Low-angle faults are common in the southern Great Basin (Wright, 1976) and, therefore, likely indicate that there is limited hydrogeological connection between some mountain aquifers and the regional aquifer system.

A second consequence of the relatively short residence times (<500 years) of many Panamint Range springs, is that these springs are entirely dependent on local recharge (Gleason et al., 2019) and groundwater flowpaths are constrained to within the mountain block (i.e., there is no connection to the DVRFS). When these results are assessed in the context of the overarching conceptual model for spring vulnerability (Figure 2), these data suggest that the Panamint Range springs are extremely vulnerable to the effects of climate change (e.g., reduced recharge due to changes in the character, timing, and duration of snow and snowpack). Under increasing aridification, we expect that springs with the shortest residence times to respond the quickest to reductions in recharge. In fact, Pistol Spring, Johnnie Shoshone Spring, Unnamed Panamint Spring G, and Tarantula Spring were dry at the time of this sampling (Figure 3). These springs were flowing as recently as the late 1990s (King & Bredehoeft, 1999) and in 2004 and 2005 (Sada & Pohlmann, 2007). However, it's difficult to rule out the possibility that these four springs are ephemeral since the results presented here are from a single sampling campaign representing a snap-shot in time and since past sampling campaigns in the Panamint Range by other research groups have also been sporadic. It should be noted that aquifer response times are a function of aquifer diffusivity (which itself is a function of hydraulic conductivity, specific yield, and aquifer thickness) and the length of flowpath (Erksine & Papaioannou, 1997; Manga, 1999; Walker, Gilfedder, Dawes, & Rassam, 2015). Thus, in the absence of repeat sampling data and due to the lack of long-term meteorological data from the Panamint Range (Gleason et al., 2019), response times cannot be easily determined using spectral analysis techniques such as those presented in Scheihing, Moya, and Tröger (2017) and Scheihing (2018). While residence times are not equivalent to response times, they do, nonetheless, provide meaningful insight into the physical processes that control flow in the mountain block. In addition, there are predictable correlations between aquatic community structure and geochemical metrics and groundwater residence times. Thus, the Panamint Range springs may be the least hydrologically robust.

The hydrological community, and state and federal agencies do a good job monitoring streamflow. However, very few springs are equipped for long-term monitoring of flow or chemistry. One notable

exception is the work of Manning et al. (2012). In addition, very few springs have been surveyed ecologically. As a consequence, (1) we do not know what biodiversity is present at these springs, (2) we have no idea what biodiversity has been lost, and (3) it remains unknown which species and communities are most vulnerable. Unfortunately, this is especially true for desert springs where the terrain and remote location of these springs may limit accessibility. The Panamint Range is geologically complex and the location, discharge, and geochemical composition of its springs are closely coupled to the geology. Given the short mean residence times of many of these springs, changes in recharge may rapidly affect the flow rate, geochemical composition, and residence times of the springs with impacts to their associated aquatic ecosystems. Aquatic species are not the only species that will be impacted by spring desiccation. For example, the Panamint alligator lizard (*Elgaria panamintina*) is listed by the IUCN as a vulnerable species and is found in riparian habitat in the Panamint Range (Phillips et al., 1983; Pister, 1991). During this study, the Pacific Treefrog (*Pseudacris regilla*) was found in springs and spring runs in Hanaupah Canyon and Surprise Canyon. Furthermore, Wauer (1964) reported that the Panamint Range supports over 140 species of birds, of which, over 70 species nest there and many of the bird species nest in riparian habitat associated with springs. Thus, the springs of the Panamint Range support riparian habitat which, in turn, supports amphibians, reptiles, and birds. Future investigations should focus on how geochemical parameters, residence times, and isotopic concentrations in the spring systems vary (or not) under differing hydrologic conditions, including summer vs. winter and during extended periods of drought. This will further highlight the sensitivity of these springs and their ecosystems to hydrologic changes.

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DATA AVAILABILITY STATEMENT

The data that support the findings of this study are available from the corresponding author upon reasonable request.

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SUPPORTING INFORMATION

Additional supporting information may be found online in the Supporting Information section at the end of this article.

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