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 Thordike 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
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, & Eguiarte, 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 Owlshead 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) 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 (\(^3\text{H}\)) and radiocarbon (\(^{14}\text{C}\)) age-dating, and an isotopic chronometer built using chlorine-36 (\(^{36}\text{Cl}\)). 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...
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).
Telescope Peak (UTM: 11S, 491,979 mE, 4,002,807 mN) is the highest point in the Panamint Range (3,366 msl [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 msl, 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 Paleozoic 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 throughout 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; Chichanski, 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 snowmelt 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 msl) 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 msl) in Death Valley range from 30 to 46 °C and average winter temperature range from 4 to 10 °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.
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 μm 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 (C2H5OH) 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 Polarp 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 tophographic map showing spring location (7.5' Topo), drainage where spring emerges, rock units, rock type, contact type, and reference (Ref.)

<table>
<thead>
<tr>
<th>Spring ID</th>
<th>Spring name</th>
<th>Spring description</th>
<th>7.5' Topo</th>
<th>Drainage</th>
<th>Rock units</th>
<th>Rock type</th>
<th>Contact type</th>
<th>Ref.</th>
</tr>
</thead>
<tbody>
<tr>
<td>PAN 1</td>
<td>Jail Spring</td>
<td>Small spring</td>
<td>T.P.</td>
<td>Jail Canyon</td>
<td>Zn/Zj</td>
<td>LS</td>
<td>Depos.</td>
<td>ALLM</td>
</tr>
<tr>
<td>PAN 2</td>
<td>Thornside Spring</td>
<td>Small spring</td>
<td>T.P.</td>
<td>Mahogany Flat</td>
<td>Zj</td>
<td>LS</td>
<td>None</td>
<td>ALLM</td>
</tr>
<tr>
<td>PAN 3</td>
<td>Unnamed Panamint Spring A*</td>
<td>Small spring</td>
<td>B.</td>
<td>Pleasant Canyon</td>
<td>Zk</td>
<td>Qtz/LS</td>
<td>Depos.</td>
<td>ALLM</td>
</tr>
<tr>
<td>PAN 4</td>
<td>Unnamed Panamint Spring F*</td>
<td>Small spring</td>
<td>B.</td>
<td>Pleasant Cayon</td>
<td>Xmi</td>
<td>Ga</td>
<td>None</td>
<td>ALLM</td>
</tr>
<tr>
<td>PAN 5</td>
<td>Wheel Spring*</td>
<td>Seep</td>
<td>W.P. 0</td>
<td>Trail Canyon</td>
<td>Cc</td>
<td>LS</td>
<td>None</td>
<td>HM</td>
</tr>
<tr>
<td>PAN 6</td>
<td>High Noon Spring*</td>
<td>Small spring</td>
<td>W.P. 0</td>
<td>Trail Canyon</td>
<td>Kis/Cn</td>
<td>G/LS</td>
<td>Fault</td>
<td>HM</td>
</tr>
<tr>
<td>PAN 7</td>
<td>Apron Spring*</td>
<td>Small spring</td>
<td>W.P. 1</td>
<td>Trail Canyon</td>
<td>Cn</td>
<td>LS</td>
<td>Fault</td>
<td>HM</td>
</tr>
<tr>
<td>PAN 8</td>
<td>Main Hanaupah Spring # 2*</td>
<td>Moderate spring</td>
<td>T.P. 1</td>
<td>Hanaupah Canyon</td>
<td>Zn (RM)</td>
<td>DS</td>
<td>None</td>
<td>ALLM</td>
</tr>
<tr>
<td>PAN 9</td>
<td>Main Hanaupah Spring # 1*</td>
<td>Moderate spring</td>
<td>T.P. 0</td>
<td>Hanaupah Canyon</td>
<td>Zk/Zn</td>
<td>MS/LS/C</td>
<td>Fault</td>
<td>ALLM</td>
</tr>
<tr>
<td>PAN 10</td>
<td>South Hanaupah Spring # 3*</td>
<td>Moderate spring</td>
<td>T.P. 0</td>
<td>Hanaupah Canyon</td>
<td>Zj</td>
<td>LS</td>
<td>Fault Block</td>
<td>ALLM</td>
</tr>
<tr>
<td>PAN 11</td>
<td>Wilson Spring*</td>
<td>Small spring</td>
<td>P. 1</td>
<td>Johnson Canyon</td>
<td>Zj</td>
<td>LS</td>
<td>Fault</td>
<td>ALLM</td>
</tr>
<tr>
<td>PAN 12</td>
<td>Lower Warm Spring A*</td>
<td>Small spring</td>
<td>A.S. 1</td>
<td>Warm Spring Canyon</td>
<td>PIPb/Jis</td>
<td>BVT</td>
<td>Noncon.</td>
<td>WSS</td>
</tr>
<tr>
<td>PAN 13</td>
<td>Lower Warm Spring B*</td>
<td>Moderate spring</td>
<td>A.S. 1</td>
<td>Warm Spring Canyon</td>
<td>PIPb/Jis</td>
<td>BVT</td>
<td>Noncon.</td>
<td>WSS</td>
</tr>
<tr>
<td>PAN 14</td>
<td>Uppermost Spring*</td>
<td>Seep</td>
<td>W.P. 0</td>
<td>Death Valley Canyon</td>
<td>Zj/Zs</td>
<td>LS/Qtz</td>
<td>None/fault</td>
<td>HM</td>
</tr>
<tr>
<td>PAN 15</td>
<td>Limekiln Spring</td>
<td>Small spring</td>
<td>B. 1</td>
<td>Surprise Canyon</td>
<td>Zj</td>
<td>LS</td>
<td>None</td>
<td>ALLM</td>
</tr>
<tr>
<td>PAN 16</td>
<td>Unnamed Panamint Spring C*</td>
<td>Small spring</td>
<td>B. 1</td>
<td>Surprise Canyon</td>
<td>Zs</td>
<td>Qtz</td>
<td>Fault</td>
<td>ALLM</td>
</tr>
<tr>
<td>PAN 17</td>
<td>Warm Sulfur Spring</td>
<td>Small spring</td>
<td>B. 1</td>
<td>Panamint Valley</td>
<td>Qay</td>
<td>MS/SS</td>
<td>Depos.</td>
<td>ALLM</td>
</tr>
<tr>
<td>PAN 18</td>
<td>Post Office Spring</td>
<td>Seep</td>
<td>B. 1</td>
<td>Panamint Valley</td>
<td>Qay</td>
<td>MS/SS</td>
<td>Depos.</td>
<td>ALLM</td>
</tr>
<tr>
<td>IES-019</td>
<td>Tule Spring (March 2016)</td>
<td>Seep</td>
<td>H.C. 1</td>
<td>Death Valley</td>
<td>Qay</td>
<td>SP/Carb</td>
<td>Depos.</td>
<td>HM</td>
</tr>
<tr>
<td>IES-019</td>
<td>Tule Spring (March 2017)</td>
<td>Seep</td>
<td>H.C. 1</td>
<td>Death Valley</td>
<td>Qay</td>
<td>SP/Carb</td>
<td>Depos.</td>
<td>HM</td>
</tr>
<tr>
<td>IES-045</td>
<td>Upper Emigrant Spring</td>
<td>Small spring</td>
<td>E.C. 1</td>
<td>Emigrant Canyon</td>
<td>NF/Kis</td>
<td>SS/G</td>
<td>Fault</td>
<td>HM</td>
</tr>
<tr>
<td>IES-047</td>
<td>Poplar Spring*</td>
<td>Seep</td>
<td>E.P. 1</td>
<td>Wildrose Canyon</td>
<td>Zk brecc</td>
<td>MS/C</td>
<td>?</td>
<td>ALLM</td>
</tr>
</tbody>
</table>

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 P.E. = 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 (breccia = 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; μS cm⁻¹ corrected to 25°C) was calculated by the meter (SpC = EC * 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 (TDS = EC × 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⁺, Mg²⁺, Sr²⁺, and Ca²⁺) were measured using a PerkinElmer Optima 5,300 DV ICP-OES according to EPA 200.7. Anions (Br⁻, Cl⁻, F⁻, NO₂⁻, NO₃⁻, PO₄³⁻, and SO₄²⁻) 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, t½ = 12.32 years) was analyzed on all spring samples in the Panamint Range to quantify young groundwater mean residence times (< ~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),

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**TABLE 2** Elevation and field chemistry of Panamint Range springs

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<thead>
<tr>
<th>Spring ID</th>
<th>Spring name</th>
<th>Date sampled</th>
<th>UTM 11S mE</th>
<th>UTM 11S mN</th>
<th>Elev. (mrl)</th>
<th>Temp. (°C)</th>
<th>pH</th>
<th>SpC (μS cm⁻¹)</th>
<th>TDS (ppm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>PAN 1</td>
<td>Jail Spring</td>
<td>May 24, 2017</td>
<td>491,216</td>
<td>4,005,046</td>
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<td>Thorndike Spring</td>
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<td>PAN 7</td>
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<td>4,004,384</td>
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<td>8.34</td>
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<tr>
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<td>Lower Warm Spring A</td>
<td>May 30, 2017</td>
<td>506,301</td>
<td>3,980,186</td>
<td>754.7</td>
<td>34.4</td>
<td>7.64</td>
<td>654.0</td>
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<tr>
<td>PAN 13</td>
<td>Lower Warm Spring B</td>
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<td>506,108</td>
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<td>Uppermost Spring</td>
<td>May 31, 2017</td>
<td>496,410</td>
<td>4,012,068</td>
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<td>16.5</td>
<td>8.45</td>
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<td>PAN 16</td>
<td>Limekiln Spring</td>
<td>June 1, 2017</td>
<td>486,446</td>
<td>3,996,617</td>
<td>1,223.1</td>
<td>19.4</td>
<td>8.13</td>
<td>765.0</td>
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<tr>
<td>PAN 17</td>
<td>Unnamed Panamint Spring C</td>
<td>June 1, 2017</td>
<td>486,503</td>
<td>3,996,454</td>
<td>1,206.4</td>
<td>16.5</td>
<td>7.99</td>
<td>747.0</td>
<td>475.0</td>
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<tr>
<td>PAN 19</td>
<td>Warm Sulfur Spring</td>
<td>June 1, 2017</td>
<td>480,753</td>
<td>3,997,248</td>
<td>317.6</td>
<td>32.0</td>
<td>7.86</td>
<td>3,791</td>
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<tr>
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<td>479,772</td>
<td>3,998,537</td>
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<td>7.72</td>
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<td>5,798</td>
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<tr>
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<td>Tule Spring (March 2016)</td>
<td>March 17, 2016</td>
<td>510,652</td>
<td>4,010,962</td>
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<td>26.8</td>
<td>7.54</td>
<td>5,127</td>
<td>3,335</td>
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<tr>
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<td>Tule Spring (March 2017)</td>
<td>March 14, 2017</td>
<td>510,652</td>
<td>4,010,962</td>
<td>−77.4</td>
<td>27.4</td>
<td>7.35</td>
<td>3,181</td>
<td>2074</td>
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<tr>
<td>IES-045</td>
<td>Upper Emigrant Spring</td>
<td>May 19, 2016</td>
<td>482,674</td>
<td>4,031,167</td>
<td>1,230.7</td>
<td>19.8</td>
<td>7.13</td>
<td>931.0</td>
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</tr>
<tr>
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<td>Poplar Spring</td>
<td>March 13, 2017</td>
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<td>4,013,420</td>
<td>1,224.8</td>
<td>17.7</td>
<td>7.40</td>
<td>1,070</td>
<td>695.5</td>
</tr>
</tbody>
</table>

**Note:** “Elev” is the elevation of the spring emergence, “SpC” is specific electrical conductivity corrected to 25°C and “TDS” is total dissolved solids.
was used to estimate mean residence times of springs (Table 4). Residence times were calculated using TracerLPM only for springs having $^3$H 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 $^3$H 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 $^3$H 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:

$$R_j = \left( \text{random}_0^{0.20} \right) \times P_{av}$$

(1)

where; $R_i$ is the estimated recharge (cm) for year $i$, the average annual precipitation ($P_{av}$, 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:

$$R_j = \sum_{i=0}^{j} R_i$$

(2)

where; $R_j$ is the cumulative (backward in time) annual recharge (cm; recharge is accumulated from 2017 backward in time to year $j$).
\[ \frac{T_i}{C_3^i} = T_i e^{-\lambda t} \]  
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 \( \lambda \) is the decay constant for \( ^3H \) (0.05626 year\(^{-1}\); see Rose, 1993). The calculated decay-corrected \( ^3H \) for each year \( i \) is then scaled by the calculated annual recharge according to:

### Table 4

<table>
<thead>
<tr>
<th>Spring ID</th>
<th>Spring Name</th>
<th>( ^3H ) (TU)</th>
<th>( ^3H_{unc} ) (TU)</th>
<th>Residence Time (yrs)</th>
<th>( \sigma_{rt} ) (yrs)</th>
<th>DP</th>
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</thead>
<tbody>
<tr>
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<tr>
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<td>0.09</td>
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<td>-</td>
<td>-</td>
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<tr>
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<td>0.09</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>PAN 15</td>
<td>Uppermost Spring</td>
<td>0.45</td>
<td>0.09</td>
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<td>-</td>
<td>-</td>
</tr>
<tr>
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<td>Limekiln Spring</td>
<td>0.53</td>
<td>0.09</td>
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<td>0.55</td>
<td>0.09</td>
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<td>-</td>
<td>-</td>
</tr>
<tr>
<td>PAN 19</td>
<td>Warm Sulfur Spring</td>
<td>0.11</td>
<td>0.09</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>PAN 20</td>
<td>Post Office Spring</td>
<td>0.65</td>
<td>0.09</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
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<td>Tule Spring (March 2016)</td>
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<td>0.09</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>IES-019</td>
<td>Tule Spring (March 2017)</td>
<td>0</td>
<td>0.09</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>IES-045</td>
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<td>-</td>
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</tr>
<tr>
<td>IES-047</td>
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<td>0.15</td>
<td>0.09</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
</tbody>
</table>

Note: Please note that residence times were not calculated for springs with \( ^3H \) concentrations less than 1.0 TU. \( ^3H_{unc} \) is the reported analytical uncertainty. \( \sigma_{rt} \) 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 \( ^3H > 1.0 \) have residence times less than 400 years), and (3) the lower and upper boundaries for the DP parameter = 0.01 and 0.99, respectively.
where: $T_{Ri}$ is the recharge-scaled annual $^3$H (TU·cm) for year $i$. The recharge-scaled annual $^3$H is then accumulated backward in time from 2017 to year $j$ according to:

$$T_j = \sum_{2017}^{j} R_i T_i^j$$

The accumulated recharge-scaled $^3$H (TU·cm) through year $j$ is then divided by the accumulated recharge (cm) through year $j$ to provide a recharge-weighted $^3$H 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 $^3$H 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 $^3$H in the aquifer. We compare the results of the TracerLPM model with this mixing model to assess the uncertainty in $^3$H 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 $\delta^{13}$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 Garner (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 $\delta^{13}$C mixing processes. In comparison, the Fontes and Garner (F&G) and Han & Plummer (H&P) models both combine $\delta^{13}$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 $\delta^{13}$C in all the carbonate facies present in the Panamint Range (it allows us to use broad endmembers in the evolution of $\delta^{13}$C 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 $\delta^{13}$Cterm by ±1‰ since it is considered the most uncertain term in the correction model.

### 3.3.3 Chlorine-36 analyses and creation of the $^{36}$Cl/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 ($^{36}$Cl/Cl) were analyzed by the Purdue Rare Isotope Measurement Laboratory (PRIME Lab) using an accelerator mass spectrometer with a relative measurement uncertainty between ±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 $^3$H concentrations (<0.7 TU). Therefore, a $^{36}$Cl/Cl chronometer was created by fitting a power-law trendline to the relationship between mean residence time (years) and measured $^{36}$Cl/Cl ($\times 10^{-15}$) of seven springs that had acceptable $^3$H and $^{14}$C residence times (Figures S2–S4). The $^{36}$Cl/Cl chronometer is given by:

$$Residence\;time\; \left(\text{years}\right) = 234.160 \times \left(\frac{^{36}\text{Cl}}{^{35}\text{Cl}}\right)^{-0.878}; \quad r^2 = 0.96$$

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 $^{36}$Cl residence times for Jail Spring, Thordndike Spring, Apron Spring, Main Hanaupah Spring # 1, Main Hanaupah Spring # 2, and South Hanaupah Spring # 3 (Table 4).

The robustness of the $^{36}$Cl/Cl chronometer was assessed using Cl$^-$/Br$^-$ ratios. The selection of a Cl$^-$/Br$^-$ representative of ground-water "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 $^{36}$Cl/Cl chronometer is site
TABLE 5  Carbon isotope data for selected springs (δ¹³C and δ¹⁴C data and residence times for springs)

<table>
<thead>
<tr>
<th>Spring ID</th>
<th>Spring name</th>
<th>pmC (uncertainty)</th>
<th>Measured δ¹³C (%)</th>
<th>Calculated δ¹³C (%)</th>
<th>I &amp; P R.T. (years)</th>
<th>F &amp; G R.T. (years)</th>
<th>H &amp; P R.T. (years)*</th>
</tr>
</thead>
<tbody>
<tr>
<td>PAN 1</td>
<td>Jail Spring</td>
<td>0.7654 (0.0026)</td>
<td>−9.1</td>
<td>−9.09</td>
<td>Modern (&lt;150 years)</td>
<td>Modern (&lt;150 years)</td>
<td>Modern (&lt;150 years)</td>
</tr>
<tr>
<td>PAN 13</td>
<td>Lower Warm Spring A</td>
<td>0.3408 (0.0013)</td>
<td>−4.3</td>
<td>−4.30</td>
<td>1,409</td>
<td>1,421</td>
<td>1,412</td>
</tr>
<tr>
<td>PAN 14</td>
<td>Lower Warm Spring B</td>
<td>0.3028 (0.0012)</td>
<td>−4.8</td>
<td>−4.79</td>
<td>1,829</td>
<td>1,842</td>
<td>1,832</td>
</tr>
<tr>
<td>IES-019</td>
<td>Tule Spring (March 2016)</td>
<td>0.7763 (0.0028)</td>
<td>+0.1</td>
<td>–</td>
<td>Modern (&lt;150 years)</td>
<td>–</td>
<td>–</td>
</tr>
<tr>
<td>IES-019</td>
<td>Tule Spring (December 2016)</td>
<td>0.5367 (0.0017)</td>
<td>−9.5</td>
<td>–</td>
<td>1,369</td>
<td>–</td>
<td>–</td>
</tr>
<tr>
<td>IES-019</td>
<td>Tule Spring (March 2017)</td>
<td>0.7373 (0.0021)</td>
<td>−12.6</td>
<td>–</td>
<td>1,078</td>
<td>–</td>
<td>–</td>
</tr>
</tbody>
</table>

Note: I&P represents Ingeron 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 ¹³C of soil CO₂ was assumed to range from −20‰ to −12‰ (Quade, Cerling, & Bowman, 1989). The ¹³C of DIC in the bedrock for Jail Spring was assumed to range from −4% to +2% (Peterson et al., 2011; Prave, 1999). The ¹³C 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 δ¹³C is not out of the range of possibilities. The residence times of Tule Spring were calculated using the Ingeron and Pearson approach assuming that the δ¹³C of soil CO₂ was −15.0‰, δ¹⁴C of soil = 100 pmC, and δ¹³C 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, 496,410 mE, 4,019,396 mN) and Dripping Spring (UTM: 11S, 492,357 mE, 3,975,186 mN). Anvil Spring emerges in Butte Valley which is geographically close to Lower Warm Spring A and Lower Warm Spring B and had an uncorrected radiocarbon residence time of 2,710 ± 40 years. Dripping Spring is located near Wildrose Peak and had an uncorrected radiocarbon residence time of 6,430 ± 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 ³⁶Cl/Cladj Adjusted ³⁶Cl/Cl ratios (³⁶Cl/Cladj) were calculated for the following springs: Warm Sulfur Spring and Upper Emigrant Spring (³⁶Cl/Cl was not measured for these springs), Post Office Spring (the measured ³⁶Cl/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 ³⁶Cl dilution likely due to mixing with hydrothermal deposits). Therefore, ³⁶Cl/Cladj was estimated for these springs using: ³⁶Cl/Cladj = (2,796.3 × δ³H) + 202.6; r² = 0.997. The data are plotted in Figure S2. This is a linear regression. The uncertainty in the regression (σr) was calculated from:

\[ \sigma_r = \sqrt{\frac{\sigma_{yy} - (m^2 - \sigma_{xx})^2}{N-2}} \]  

where,

\[ \sigma_{yy} = \sum (y - \bar{y})^2 \]  
\[ \sigma_{xx} = \sum (x - \bar{x})^2 \]  

\[ m = \text{slope of linear regression} = \frac{\sigma_{xy}}{\sigma_{xx}} \]  

\[ \sigma_y = \sqrt{1 + \frac{1}{N} + \frac{(x - \bar{x})^2}{\sigma_{xx}}} \]  

The estimated uncertainties for the calculated ³⁶Cl/Cladj values are shown in Table 6.

Uncertainty in the calculation of residence times from the ³⁶Cl/Cl chronometer

The ³⁶Cl/Cl chronometer is given by: RT = 234,160 × (³⁶Cl/Cl)⁻⁰·⁸⁷⁷⁸, r² = 0·96; where RT equals residence time in years. This equation is a power-law trendline (Figure S3) taking the form:

\[ y = bx^a \]  

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

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

This results in:

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

The uncertainty in the estimated y-value ($\sigma_y$), 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 ($^{87}$Sr/$^{86}$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: Noonday 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 Brediehoeft (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-calcium-sulfate waters plotting in quadrant IV of Figure 5. The two 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 ± 613 years), Lower Warm Spring B (1,829 ± 613 years), Wheel Spring (1,787 ± 641 years), and Upper Emigrant Spring (1,784 ± 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 ± 400 years.

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

\(^{3}\text{H}\) 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 (\(10^{-15}\)) for the region. All but six springs have \(^{36}\text{Cl}/\text{Cl}\) ratios higher than 500 (\(10^{-15}\)). Two separate trends emerge when \(^{3}\text{H}\) 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 \(^{3}\text{H}\) concentrations less than 1.0 TU and \(^{36}\text{Cl}/\text{Cl}\) ratios less than 900 × 10\(^{-15}\).
and two springs in Pleasant Canyon (Unnamed Panamint Spring E and Unnamed Panamint Spring F) have $^{87}$Sr/$^{86}$Sr ranging from 0.7200 to 0.7450. 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 $^{87}$Sr/$^{86}$Sr ratios (0.72875 and 0.73342, respectively) based on the rocks present in their emergence (Figure 9).

### 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
concentration in these spring samples, but not 86Sr. They correctly state that 87Sr/86Sr ratios are suitable for determining the source water for springs; however, the lack of 86Sr concentrations limits interpretation of their data to broad correlations (note, we calculate the 87Sr/86Sr for their samples for easier comparison in Table 9). For example, higher Sr2+ is consistent with granitic rocks in bedrock highlands and the highest concentration of Sr2+ is from springs at or near salt pans.

The 87Sr/86Sr ratios of Panamint Range springs in this study range from 0.71245 to 0.73373 (Table 7). The Sr2+ concentrations range from 0.108 to 8.78 ppm, which encompasses the range of concentrations found by King and Bredehoeft (1999). The combination of 87Sr/86Sr ratios from spring water and 87Sr/86Sr 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.

### Table 9
The strontium solute concentrations (Sr2+) and strontium isotopic data (87Sr/86Sr) for 10 Panamint Range springs that were sampled by King and Bredehoeft (1999) are compiled here for comparison.

<table>
<thead>
<tr>
<th>Spring name</th>
<th>UTM 115 mE</th>
<th>UTM 115 mN</th>
<th>Elev. (mrsl)</th>
<th>Sr2+ (ppm)</th>
<th>δ87Sr (‰)</th>
<th>87Sr/86Sr</th>
</tr>
</thead>
<tbody>
<tr>
<td>Anvil Spring (NPS # 123)</td>
<td>492,357.0</td>
<td>3,975,186.0</td>
<td>1,296.3</td>
<td>0.1811</td>
<td>16.15</td>
<td>0.7207</td>
</tr>
<tr>
<td>Burns # 1 Spring (NPS # 53)</td>
<td>483,135.8</td>
<td>4,027,730.1</td>
<td>1,596.8</td>
<td>0.8944</td>
<td>4.50</td>
<td>0.7124</td>
</tr>
<tr>
<td>Dripping Spring (NPS # 63)</td>
<td>496,409.5</td>
<td>401,395.5</td>
<td>1,157.9</td>
<td>1.0364</td>
<td>26.68</td>
<td>0.7281</td>
</tr>
<tr>
<td>Hummingbird Spring (NPS # 76)</td>
<td>490,288.3</td>
<td>4,008,308.5</td>
<td>2,194.6</td>
<td>0.3010</td>
<td>28.00</td>
<td>0.7291</td>
</tr>
<tr>
<td>Johnnie Shoshone Spring (NPS # 74)</td>
<td>493,610.9</td>
<td>4,011,201.6</td>
<td>2,194.6</td>
<td>0.3365</td>
<td>21.50</td>
<td>0.7245</td>
</tr>
<tr>
<td>Limekiln Spring* (NPS # NA)</td>
<td>485,700.0</td>
<td>3,996,422.0</td>
<td>1,236.9</td>
<td>0.4559</td>
<td>33.91</td>
<td>0.7333</td>
</tr>
<tr>
<td>Surprise Canyon Crk Spring (NPS # 248)</td>
<td>484,500.0</td>
<td>39,966,239.0</td>
<td>833.0</td>
<td>0.4540</td>
<td>34.03</td>
<td>0.7333</td>
</tr>
<tr>
<td>Thorndike Spring* (NPS # 75)</td>
<td>493,285.3</td>
<td>4,009,722.9</td>
<td>2,395.7</td>
<td>0.1297</td>
<td>17.30</td>
<td>0.7215</td>
</tr>
<tr>
<td>Upper Emigrant Spring* (NPS # 45)</td>
<td>482,544.0</td>
<td>4,030,966.0</td>
<td>1,257.9</td>
<td>1.1121</td>
<td>17.36</td>
<td>0.7215</td>
</tr>
<tr>
<td>Wildrose Spring* (NPS # 72)</td>
<td>482,584.0</td>
<td>4,013,281.0</td>
<td>1,197.9</td>
<td>1.2358</td>
<td>10.64</td>
<td>0.7168</td>
</tr>
</tbody>
</table>

Note: All data, except for the 87Sr/86Sr values, were published in King and Bredehoeft (1999). We calculated 87Sr/86Sr based on the following equation: \[ (\delta^{87}\text{Sr}/1,000) + 1 \times 87\text{Sr}/86\text{Sr}_{\text{std}} \] where 87Sr/86Srstd equals the 87Sr/86Sr 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 87Sr/86Sr 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).

<table>
<thead>
<tr>
<th>Geologic map symbol</th>
<th>Stratigraphic rock unit</th>
<th>Epoch</th>
<th>Min. 87Sr/86Sr</th>
<th>Max. 87Sr/86Sr</th>
<th>Mean 87Sr/86Sr</th>
</tr>
</thead>
<tbody>
<tr>
<td>PMb (PnPb)</td>
<td>Bird Spring Formation—Undivided</td>
<td>Late Mississippian Early Permian</td>
<td>0.7073</td>
<td>0.7084</td>
<td>0.7079</td>
</tr>
<tr>
<td>Cn</td>
<td>Nopah Formation</td>
<td>Late Cambrian</td>
<td>0.7091</td>
<td>0.7092</td>
<td>0.7091</td>
</tr>
<tr>
<td>Cb</td>
<td>Bonanza King Formation</td>
<td>Middle and Late Cambrian</td>
<td>0.7088</td>
<td>0.7092</td>
<td>0.7091</td>
</tr>
<tr>
<td>Cc</td>
<td>Carrara Formation</td>
<td>Early and Middle Cambrian</td>
<td>0.7085</td>
<td>0.7092</td>
<td>0.7087</td>
</tr>
<tr>
<td>CZw (Czw)</td>
<td>Wood Canyon Formation</td>
<td>Neoproterozoic and Early Cambrian</td>
<td>0.7079</td>
<td>0.7088</td>
<td>0.7084</td>
</tr>
<tr>
<td>Zs</td>
<td>Stirling Quartzite</td>
<td>Neoproterozoic</td>
<td>0.7055</td>
<td>0.7085</td>
<td>NR</td>
</tr>
<tr>
<td>Zj</td>
<td>Johnnie Formation</td>
<td>Neoproterozoic</td>
<td>0.7055</td>
<td>0.7085</td>
<td>NR</td>
</tr>
<tr>
<td>Zn</td>
<td>Noonday Dolomite</td>
<td>Neoproterozoic</td>
<td>0.7055</td>
<td>0.7085</td>
<td>NR</td>
</tr>
<tr>
<td>ZYb (Zb)</td>
<td>Beck Spring Dolomite</td>
<td>Neoproterozoic and Mesoproterozoic</td>
<td>0.7055</td>
<td>0.7085</td>
<td>NR</td>
</tr>
<tr>
<td>ZYc (Zc)</td>
<td>Crystal Spring Formation</td>
<td>Neoproterozoic and Mesoproterozoic</td>
<td>0.7055</td>
<td>0.7085</td>
<td>NR</td>
</tr>
</tbody>
</table>
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 ($^{87}\text{Sr}/^{86}\text{Sr} = 0.713–0.716$) 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
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/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).
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 $21^\circ$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

**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.
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 $^{87}$Sr/$^{86}$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 $^{87}$Sr/$^{86}$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 Skidoob Granite (Hunt & Mabey, 1966), yet the $^{87}$Sr/$^{86}$Sr ratio (0.72198) is consistent with the Proterozoic Johnnie Formation, Noonday Dolomite and Zabriskie Quartzite. We interpret this $^{87}$Sr/$^{86}$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 $^{87}$Sr/$^{86}$Sr ratio and Sr$^{2+}$ concentration of Uppermost Spring (PAN 15) in Death Valley Canyon. The $^{87}$Sr/$^{86}$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 Sierraville Granite (Hunt & Mabey, 1966). We interpret this $^{87}$Sr/$^{86}$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.

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 Mg$^{2+}$/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 $^{87}$Sr/$^{86}$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 $^{36}$Cl/Cl (lower than predicted by their $^3$H concentration; Figure 8) consistent with $^{36}$Cl/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. We have no clear explanation of the $^{87}$Sr/$^{86}$Sr ratio and Sr$^{2+}$ concentration of Uppermost Spring (PAN 15) in Death Valley Canyon. The $^{87}$Sr/$^{86}$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 Sierraville Granite (Hunt & Mabey, 1966). Water interaction with these bedrock units should produce a significantly lower $^{87}$Sr/$^{86}$Sr ratio.
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 ± 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 36Cl/Cl is heavily diluted due to mixing with basin brines. The residence time of this spring is variable (<150 years to 1,369 ± 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 δ13C was +0.1% 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 δ13C decreased to −9.5% and the pmC decreased to 0.5367 in December 2016. In March 2017, the δ13C 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 ± 400 years (March 2017). Tule Spring has the lowest 36Cl/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 36Cl/Cl ratios and tend to dilute the 36Cl/Cl of fresh water when brines mix with fresh waters (Phillips, 2000). However, the 87Sr/86Sr data for Tule Spring closely matches the 87Sr/86Sr 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 36Cl/Cl relative to the other springs (as predicted by their 3H concentrations; Figure 8). To correct for the observed 36Cl/Cl dilution observed in these springs, the relationship between 36Cl/Cl and 3H (for unaffected springs) was used to calculate an adjusted 36Cl/Cladj given by: 36Cl/Cladj = (2,796.3 × 3H) + 202.6; r² = 0.997, p < .01 at the 95% confidence interval (Figure S2). Adjusted 36Cl/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 36Cl/Cladj values were then used to provide additional calibration points in the 36Cl/Cl chronometer (Figures S3 and S4). In the absence of additional sources of chloride from evaporites, brines, and hydrothermal deposits, the relationship between 36Cl/Cl and 3H should be strongly linear as indicated by the fit (high r² and very low p-value) of the linear regression for unaffected springs. Thus, we consider the mean residence times calculated using the 36Cl/Cl chronometer to be robust with the 36Cl/Cladj values.

The basin springs are extremely difficult to age-date. Tule Spring is tritium dead and has a 36Cl/Cl modified by mixing with basin brines. However, three radiocarbon correction methods provided similar residence times for Tule Spring (Table 5). Using the Ingeron and Pearson Jr. (1964) radiocarbon correction model, we calculated radiocarbon residence times ranging from modern (<150 years) to 1,369 ± 400 years (Table 5). A radiocarbon sample was not collected for Post Office Spring since it was sampled in the spring run, and the 36Cl/Cl sample was diluted by mixing with evaporites in its spring run. Warm Sulfur Spring was not sampled for 36Cl/Cl and its radiocarbon was suspect. Therefore, 36Cl/Cladj ratios were also calculated for these two springs. Using the 36Cl/Cladj ratios, we calculate 36Cl/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 87Sr/86Sr 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 Thorn dike 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 ± 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 hydrogeologic 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 Piaschyk, 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 (*Egaria 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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