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# Groundwater Hydrology of The Henry's Fork Springs


## Abstract

*The Henry's Fork springs supply water important to agriculture, recreation, and fish and wildlife, but no information exists about the recharge processes for these springs. In the absence of regional well data, a methodology combining physical data and environmental isotope investigations was developed to assess spring recharge areas, flow paths, and residence times. The stable isotopes of oxygen and hydrogen were used to constrain recharge elevation and location, and chlorofluorocarbons 11 and 12 were used to estimate residence times. Recharge area size was calculated from regional water budgets, and depth of water circulation was estimated from aquifer mass-balance considerations. Statistical analysis of temporal variability in spring discharge indicated that winter baseflows from the springs were a function of annual precipitation during the three previous years. Recharge for the larger springs occurs at elevations of 2,200 to 2,400 m, maximum depth of circulation is about 500 m, and recharge areas for the springs range from 15 to 300 km$^2$. Ratios of oxygen and hydrogen isotopes indicated negligible water-rock isotopic exchange under geothermal conditions nor evaporation during the recharge process. However, both high heat-flux values and evidence of mantle helium in spring waters point to connection via upward diffusion between spring flow paths and the underlying mantle. Chlorofluorocarbon (CFC) concentrations in Big Springs, Lucky Dog, Warm River, and Snow Creek indicated recharge years that ranged from pre-1900 to 1994.*



**Key words:** hydrology, springs, stable isotopes, chlorofluorocarbons, groundwater, residence time, Henry's Fork of the Snake River, recharge.


## Introduction

Groundwater resources in the western United States are increasingly being used to meet growing demands for water. Characterization of the physical and geochemical properties of a groundwater flow system requires accurate assessment of available water resources. In many areas however, limited data about hydraulic conductivity and head, e.g., when no wells exist in the recharge area, require that an alternative approach be taken. In this study, I used the variability in the discharge of groundwater at the Henry's Fork springs and environmental isotope investigations to provide a way to understand regional scale subsurface hydrologic processes such as recharge area, residence time, and flow paths.

The headwaters of the Henry's Fork of the Snake River originate as spring discharge on the eastern margin of Island Park Caldera west of Yellowstone National Park. Water from these springs provides 75 percent of the total baseflow in the Henry's Fork at Island Park and 42 percent of the total discharge in the river at Ashton (Whitehead 1978). Local agriculture, recreational interests, and fish and wildlife depend on the water in the Henry's Fork. Irrigated agriculture relies on the storage of winter baseflows from the Henry's Fork to provide late


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summer irrigation water, and much of the economy in the upper Henry's Fork basin is tied to recreation-related income. The health of the Henry's Fork blue-ribbon trout fishery depends on high winter baseflows (Van Kirk and Griffin 1997, Benjamin and Van Kirk 1999), and warmer winter water temperatures in the headwater springs of the Henry's Fork provide a greater chance of juvenile trout overwinter survival (Griffith *et al.* 1997). Although discharge from the springs maintains water quantity and quality in the river above Ashton, little is known about the aquifer(s) that feed those springs.

The Madison and Pitchstone plateaus are the common recharge areas for springs on the Island Park Caldera and some thermal systems in Yellowstone National Park. Yellowstone's Upper Geyser Basin, which contains the largest concentration of thermal features in the world, is located only 21.6 km east of the Island Park Caldera (Fig. 1). If a hydrologic connection exists between the Henry's Fork springs and Yellowstone thermal basins, any change in the recharge portion of the system may affect flow and circulation in the thermal discharge part of the system. Changes in land-use practices in recharge areas could result in unanticipated changes in both Yellowstone thermal springs and Henry's Fork springs. For example, clear-cutting on the Madison and Pitchstone plateaus during the 1960s and 1970s might result in changes in regional hydrology that have not yet

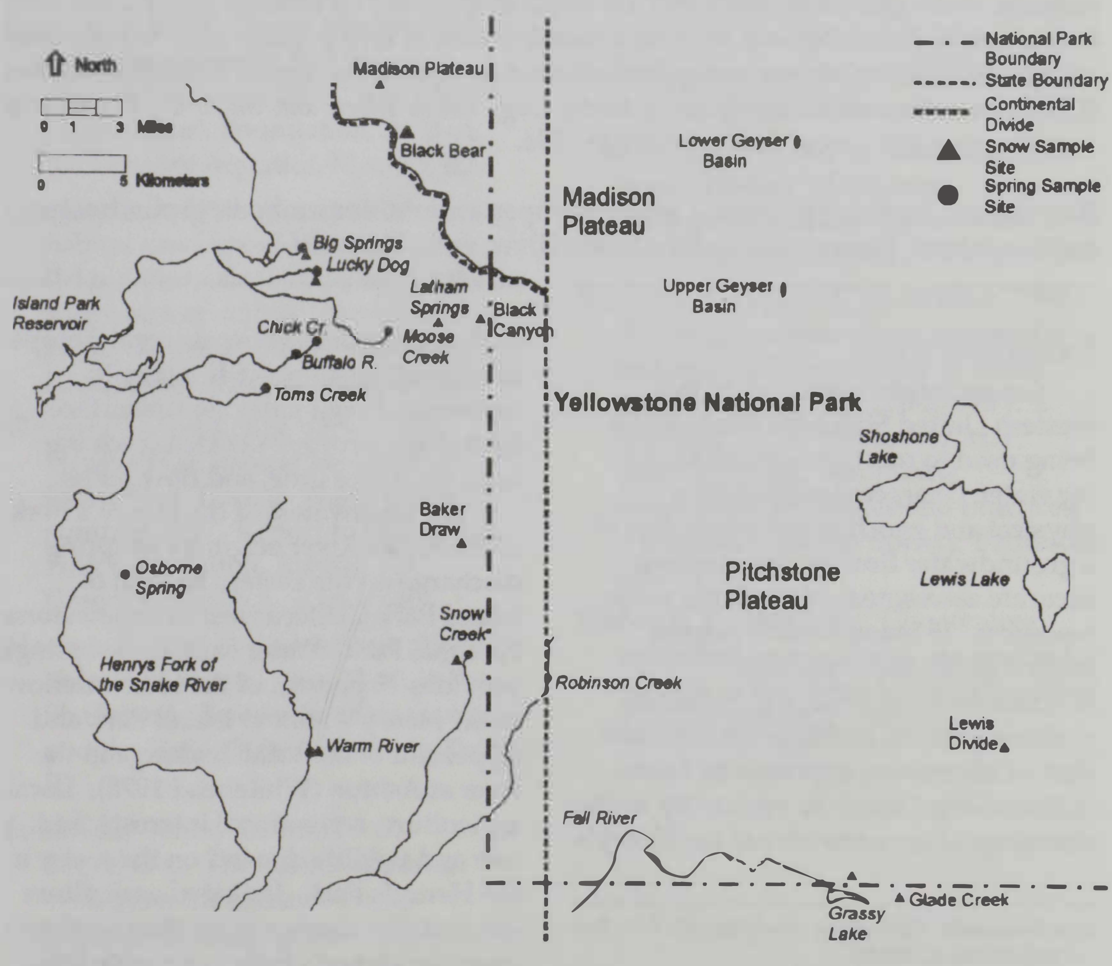



**Figure 1.** *Locations of springs and stable-isotope sampling sites, Henry's Fork region.*

been realized. Further protection for land within the recharge areas may be necessary.

The question of recharge to the Henry's Fork springs can be examined using three different timescales (Manga 1999): 1) the timing of peak discharge at the spring, reflecting the length of time that it takes the snowmelt pressure pulse to be propagated through the aquifer; 2) the length of time taken for changes in recharge to be reflected in changes in discharge, e.g., how long it takes drought patterns to be seen in spring discharge; and 3) the residence time of water in the aquifer, i.e., the transport time of the volume of water in the aquifer. I used statistical analysis of variability in spring discharge and precipitation to examine the second time scale and CFC analysis to determine the residence time of water in the aquifer. Additionally, I used stable isotope data to identify the recharge areas for the Henry's Fork springs. Watershed mass balance and heat flux calculations were used to estimate recharge area size and flow paths.

## Study Area

### Geologic Setting

Big Springs, Lucky Dog, Chick Creek, Buffalo River, Toms Creek, Snow Creek, and Warm River springs (the Henry's Fork springs) are a series of large springs that feed the Henry's Fork of the Snake River, providing about one half of the streamflow in the upper Henry's Fork watershed. Discharges that were measured at these springs are listed in Table 1. The Henry's Fork of the Snake River runs from north to south across the Island Park basin (Fig. 1), descending from the caldera rim to flow through irrigated agricultural lands until it joins the South Fork of the Snake River near Rexburg, Idaho (Van Kirk and Benjamin this issue). Island Park is located at the northeastern end of the Snake River Plain, providing a topographic and geologic transition from the basalts of the Snake River Plain to the Yellowstone volcanic field.

The Henry's Fork springs lie on the eastern edge of the Island Park basin in the western part of the Yellowstone Plateau volcanic field (Fig. 1). The Yellowstone field evolved through three cycles of rhyolitic volcanism, and Island Park basin, an integral part of the field, is a product of all three cycles over a time span of 2 million years. The southwestern rim of Island Park was formed in the first cycle of eruption 2 million years ago with the collapse of a caldera that extended east into the Yellowstone Plateau. The northwestern part of Island Park is enclosed by a rim from the second-cycle caldera that occurred 1.3 million years ago and was centered in the northern part of Island Park. The eastern side of Island Park (where the Henry's Fork springs are located) is bordered by pre- and post-caldera lava flows from the third volcanic cycle, which vented on the

**Table 1.** *Temperatures, recharge areas, discharge rates, depths of circulation, geothermal heat, and mean heat flux of Henry's Fork springs.*

| Spring | Temperature (°C) | Recharge Area ($km^2$) | Discharge ($m^3/s$) | Depth of Circulation (m) | Heat (MW) | Mean Heat Flux ($mW/m^2$) |
|---|---|---|---|---|---|---|
| Lucky Dog | 12.6 | 31.6 | 0.8 | 326.2 | 25.50 | 800 |
| Big Springs | 12.5 | 214.5 | 5.4 | 251.6 | 169.70 | 800 |
| Buffalo River | 11.3 | 45.3 | 1.1 | | 29.00 | 600 |
| Chick Creek | 11.0 | 24.2 | 0.4 | | 10.00 | 400 |
| Snow Creek | 5.5 | 15.9 | 0.4 | 46.3 | 0.80 | 50 |
| Warm River | 12.0 | 284.5 | 5.6 | 517.6 | 164.20 | 600 |

eastern side of Yellowstone National Park about 630,000 years ago (Christiansen 1982). The Henry's Fork springs emerge from the base of these rhyolitic lava flows, forming tributaries that flow west into the Henry's Fork of the Snake River.

Figure 2 shows the geology of the region immediately surrounding the Henry's Fork springs. Big Springs, Lucky Dog, Moose Creek, and Snow Creek springs each are located where the westernmost edge of the Buffalo Lake rhyolite flow meets the older Lava Creek Tuff; younger surficial alluvium also occurs in the immediate vicinity of Big Springs and Lucky Dog. Lava Creek Tuff, extending over much of the Island

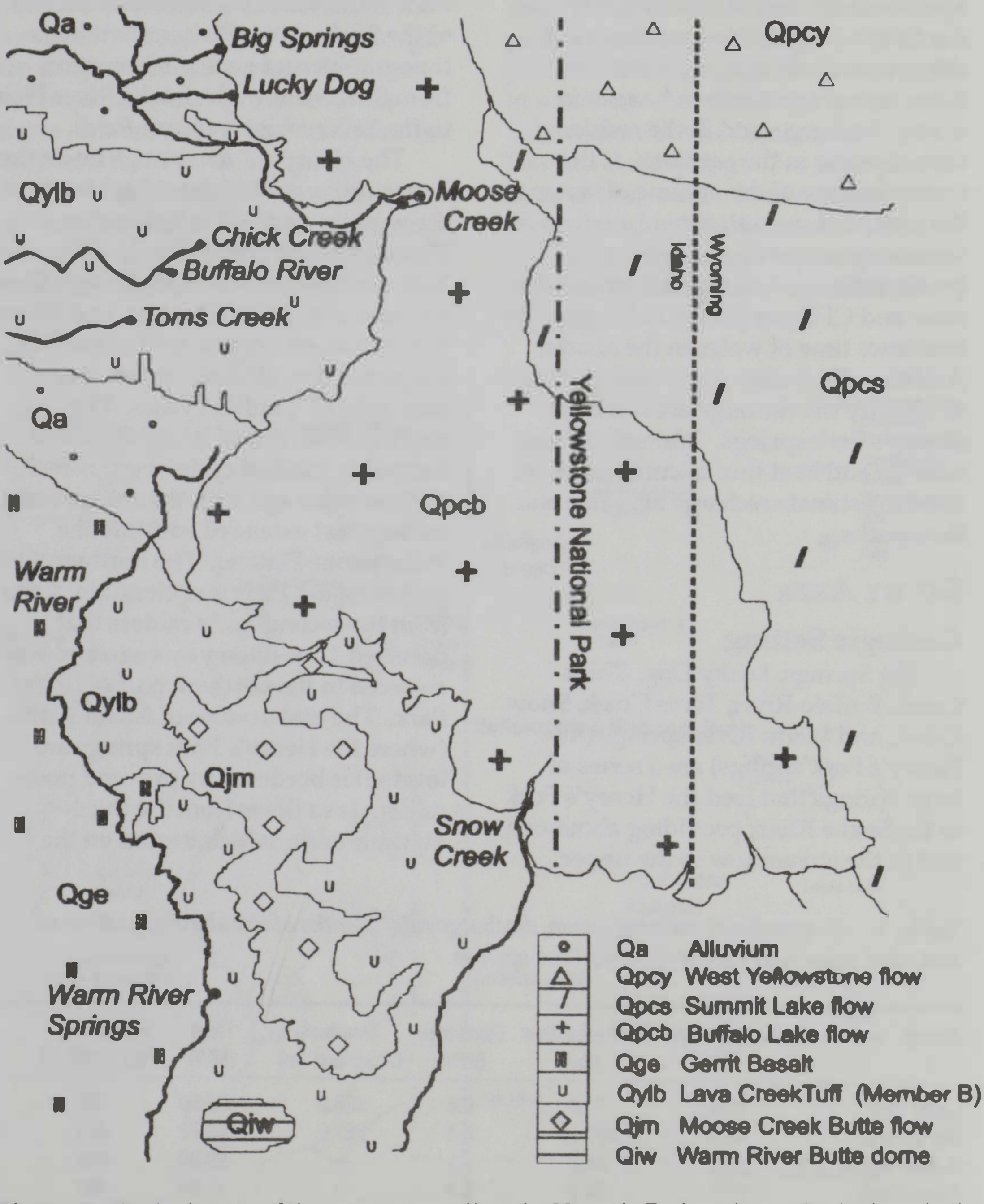



**Figure 2.** *Geologic map of the area surrounding the Henry's Fork springs. Geologic units in the legend are arranged in ascending chronological age. Adapted from Christiansen (1982).*

Park Caldera, was emplaced during the third volcanic cycle; potassium-argon dating establishes its age as 630,000 years (Christiansen 1982). At that time, the Yellowstone Caldera was formed by the collapse of the roof of the magma chamber along two ring-fracture zones. After the formation of the Yellowstone Caldera, rhyolite flows, which included Buffalo Lake flow, spilled over the Yellowstone caldera rim into the Island Park basin, forming the eastern edge of the Island Park caldera. The youngest post-caldera lava flows (150,000-70,000 yrs old) are collectively called the Plateau Rhyolite and comprise the flows that form the Madison, Pitchstone, Central, and Solfatara plateaus. The Buffalo Lake Tuff is one of the Central Plateau members of this group (Christiansen and Blank 1972). On the Madison Plateau, the Buffalo Lake Tuff is overlain by the younger Summit Lake and West Yellowstone flows. The flows in this group are large and thick with maximum diameters of 30 km and thicknesses of more than 300 m.

Chick Creek, Buffalo River and Toms Creek springs are located several kilometers to the west of the edge of the Buffalo Lake flow and emerge from Lava Creek Tuff. Warm River springs are located in a more geologically complex area (Richmond 1973). Warm River, above and below the springs, lies at the contact between Lava Creek Tuff and the Gerrit Basalt, a post-Lava Creek basalt flow that forms the floor of Island Park. Warm River springs emerge from Lava Creek Tuff, 2 km to the west of the Moose Creek Butte flow, a precaldera flow whose vent is now buried below the Madison Plateau. The geologic settings of Big Springs, Lucky Dog, Moose Creek, and Snow Creek springs described above provide insights into the mechanisms of spring emergence. It is likely that recharge waters permeate and flow down-gradient through the Central Plateau rhyolites and emerge as springs on contact with the older and less permeable Lava Creek Tuff. I speculate that a similar mechanism occurs at Buffalo River, Chick Creek, and Toms Creek springs but that local stratigraphy determines where the springs appear. It is likely that recharge for the large springs at Warm River occurs through Moose Creek Butte, Lava Creek Tuff, and Buffalo Lake flow rhyolites.

## Hydrologic Setting

Previous hydrologic studies in the upper Henry's Fork basin were limited and focused on surface-water investigations and water resources assessments. Whitehead (1978) provided the most comprehensive survey of surface water supply and water quality in the region. The U.S. Geological Survey (USGS) has maintained stream-gauging stations below Henry's Lake and at Island Park since 1920 and 1933, respectively, and more recently has installed stream gauges above Island Park Reservoir and at Big Springs and Lucky Dog springs. Benjamin (1997) provided an assessment of Island Park Reservoir operations and its effects on the unregulated hydrologic regime of the Henry's Fork at Island Park.

Two distinct hydrologic regimes exist for tributary streams in the Henry's Fork basin. On the west side of the basin, streams that emerge from the Centennial Mountains are fed by snowmelt and have low winter base flows and high spring peak flows. On the east side, streams that originate from springs at the base of the Yellowstone lava flow have high baseflows and relatively small peaks typical of groundwater-dominated flow systems. The unregulated upper Henry's Fork of the Snake River itself shows the same groundwater-dominated characteristics (Fig. 3). Water recharged during the peak nuclear-weapons testing period (1953-1962) is still being flushed from the aquifer as assessed by a chlorine-36/stable chlorine ratio of $2110 * 10^{15}$ and a

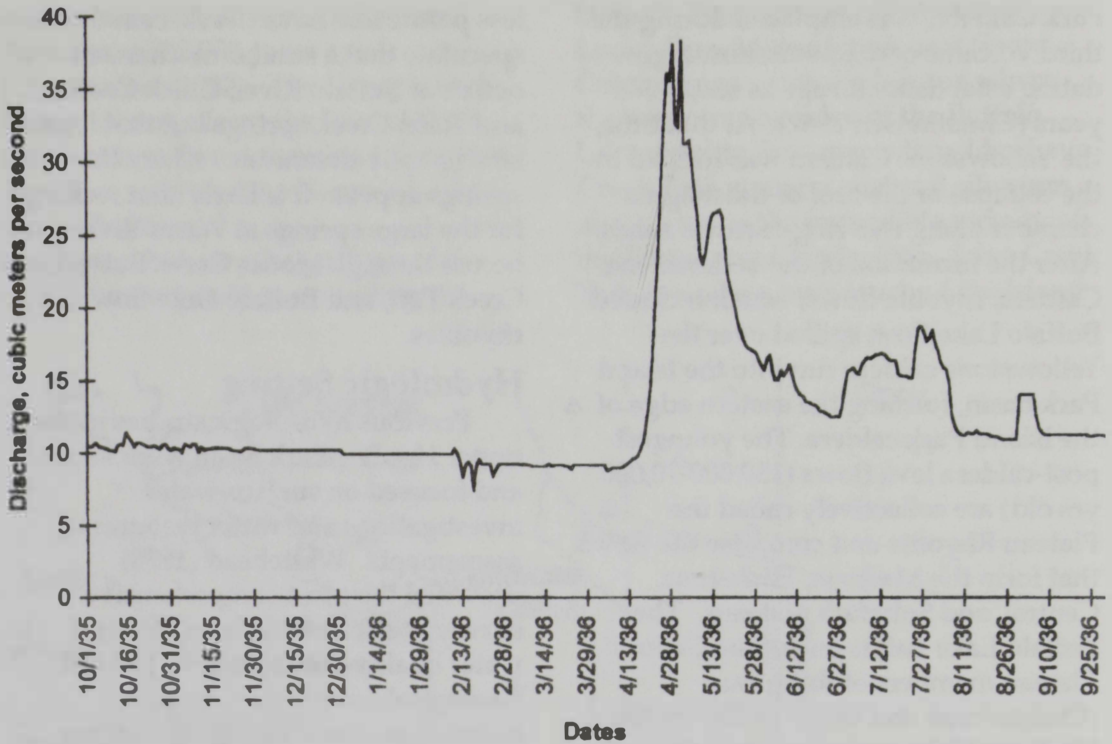



**Figure 3.** *Hydrograph of the Henry's Fork at Island Park, water year 1936.*

tritium value of 40.6 tritium units (L. D. Cecil, U.S. Geological Survey, personal communication).

# Methods

## Stable Isotopes

The stable isotope composition of water has been widely used to identify groundwater source areas, better understand recharge processes, and provide insight into aquifer residence times (Sorey *et al.* 1978, Ingraham *et al.* 1991, Ellins 1992, Smith *et al.* 1992, Sveinbjörnsdóttir and Johnsen 1992, Winograd *et al.* 1998). Studies that combine stable isotope analysis with physical data are most relevant to current research on the Henry's Fork springs. Recent examples include investigations of spring systems in California's Lassen Highlands (Rose *et al.* 1996) and Oregon's eastern Cascades (Manga 1996, 1997). Spatial variability in the stable isotope ratios in precipitation often is reflected in groundwater, indicating the source area of groundwater (Ingraham and Taylor 1991, Friedman *et al.* 1992). Similarly, temporal variability in precipitation and groundwater can indicate season of recharge. Additionally, the relationship between $\delta\,^{18}O$ and $\delta D$ ($\delta^2H$) can point to processes that occur prior to and during recharge (Buttle and Sami 1990). $\delta$ (delta) is defined by the equation

$$d = \left( \frac{R_x}{R_{std}} - 1 \right) * 1000 \qquad (1)$$

where $R_x = \frac{^{18}O}{^{16}O} \text{ or } \frac{^{2}H}{^{1}H}$

in the sample, and $R_{std}$ = the same ratio in the reference standard. The Vienna Standard Mean Ocean Water (VSMOW) reference standard is used for both oxygen and hydrogen isotopes in water.

The ratios of the stable isotopes of oxygen ($^{16}O$ and $^{18}O$) and hydrogen ($^2H$ and H) in atmospheric moisture vary systematically as a function of temperature because of isotopic fractionation processes during evaporation from the ocean and subsequent condensation (Clark and Fritz 1997). The stable isotope ratios of water vapor in an air mass reflect the origin of the air mass, and the ratios in the precipitation that evolves from the air mass reflect both the origin of the air mass and the conditions under which condensation occurs. As an air mass travels away from the ocean, precipitation that is enriched in the heavier isotopes leaves the air mass first, enriching the remaining water vapor in lighter isotopes. Subsequent precipitation has increasingly lighter stable isotope compositions. This depletion process has been called the "continental effect" and generally results in lighter stable isotope ratios further inland. Gradients in deuterium composition of precipitation in California and Nevada range from 20‰ per 100 km close to the coast to 2‰ per 100 km further inland (Friedman and Norton 1970, Ingraham and Taylor 1991, Williams and Rodoni 1997). Furthermore, a strong linear correlation exists between mean annual isotopic composition of precipitation and mean annual surface temperature. This relationship corresponds to a 1‰ decrease in mean annual $\delta^{18}O$ with a 1.1 to 1.7 °C decrease in mean annual temperature (Dansgaard 1964); $\delta D$ varies with temperature similarly to $\delta^{18}O$. As a result, precipitation that occurs at higher latitudes has a lighter stable isotope composition than that which occurs closer to the equator. This temperature effect is also seen as a result of altitude; cooler temperatures at higher elevations result in $\delta^{18}O$ depletion that varies between -0.15 and -0.50‰ per 100-m rise in altitude (Clark and Fritz 1997). Similarly, strong seasonal variability in stable isotopic composition of precipitation occurs as a result of seasonal differences in temperature. This variability is particularly pronounced in continental locations where seasonal temperature differences are extreme.

The relationship between $\delta D$ and $\delta^{18}O$ in precipitation conforms closely to an empirical relationship known as the Global Meteoric Water Line (GMWL) developed by Craig (1961) and expressed by the equation

$$\delta^{2}H = 8\,\delta^{18}O + 10‰ \qquad (2)$$

This relationship was developed as an average of $\delta^{18}O$ and $\delta D$ values collected in many different climatic and geographic regions. The slope of the evaporation line in the $\delta^{18}O$ - $\delta D$ space is related to the ratio of the kinetic fractionation factors for hydrogen and oxygen isotopes, a function of humidity and temperature. In arid regions for example, greater isotopic fractionation of $^{18}O$ than deuterium with evaporation during rainfall results in disproportional enrichment of $^{18}O$ relative to deuterium and a lower slope for the Local Meteoric Water Line (LMWL).

Variability in stable isotopic ratios in precipitation is transferred to groundwaters, providing information about recharge location and seasonality. For example, Ingraham *et al.* (1991) and Sveinbjörnsdóttir and Johnsen (1992) used variability of stable isotope values in local precipitation to identify sources of recharge to spring systems, and Winograd *et al.* (1997) used seasonal variability in precipitation to identify the relative contributions of summer and winter precipitation to groundwater recharge. However, during recharge snowmelt, evaporation and geothermal heating may modify the meteoric input signal. Evaporation of water after precipitation, but prior to infiltration, may cause heavy isotope enrichments in groundwaters. Differential fractionation of $\delta^{18}O$ and $\delta D$ during evaporation of water from the unsaturated zone or infiltrating runoff water leads to the slope of the $\delta^{18}O$ and $\delta D$ relationship in groundwaters being much lower than that of the local precipitation. Geothermal heating also may alter the $\delta^{18}O$ and $\delta D$ relationship because as meteoric water is circulated to greater depths and heated, the oxygen isotopes exchange with oxygen in the rocks (Truesdell *et al.* 1977, Sorey *et al.* 1978, Fournier 1989). Several studies (Friedman *et al.* 1991, Mast *et al.* 1995) have found that snowpacks show an evolution toward isotopically heavier values over time. During snowmelt, as meltwater percolates through the pack, an isotopic equilibrium between water and ice is established that alters the stable isotopic ratios of the meltwater.

I collected water for oxygen-18 and deuterium analysis to identify recharge areas and processes at sites shown in Figure 1 and listed with geographic descriptors in Table 2. Three types of waters were collected for analysis: water from each spring, late March snow cores for analysis of winter precipitation, and summer precipitation samples. All water samples were submitted to the Isotope Fractionation Project at the U.S. Geologic Survey laboratory in Reston, Virginia, for analysis.

Each of the medium to large discharge springs in the region, i.e., Big Springs, Lucky Dog, Chick Creek, Buffalo River, Warm River, and Snow Creek, were sampled in July 1998 and 1999 and September 1999. Fifty ml of spring water were collected in glass containers with polyseal caps where the springs emerged from the ground.

Bulk snow cores were collected at established Snotel or Snocourse sites at Black Bear and Madison Plateau, Montana, Latham Springs, Idaho, and Lewis Divide, Grassy Lake, and Glade Creek, Wyoming. Snow cores also were collected at Black Canyon, upper Baker Draw, Big Springs, Lucky Dog, Warm River, and Snow Creek in March 1999. Snow cores were collected by snowmobile with assistance from the Natural Resources Conservation Service (NRCS) and U.S. Bureau of Reclamation (BOR) at each of the sites described above between 20 March and 30 March 1999 and 2000. All snow cores were collected using a Mt. Rose snow corer provided by the NRCS and the BOR. Cores were then placed in large plastic bags and sealed with duct tape. As the cores melted, the melt water was mixed and transferred to glass containers with polyseal caps.

Summer precipitation was collected between July and September 1999 at Lucky Dog, Warm River, upper Baker Draw, and Latham Springs. No attempt was made to establish the stable isotope values of individual storm events; rather, precipitation was collected for the whole summer and an average value was established. Precipitation collection devices consisted of 1.9-liter glass jars with funnels sealed onto the top and a 0.5-cm layer of mineral oil at the bottom of the jar to prevent evaporation. The jars were partially buried in the ground at each site. Three stakes were hammered into the ground and taped to the funnels to stabilize the equipment. Water was extracted from below the mineral oil layer using a syringe, transferred into 50-ml glass containers, and submitted to USGS for analysis.

Hydrogen isotope ratio analyses were performed using a hydrogen gas equilibration technique at 30 °C (Coplen *et al.* 1991) and a $CO_2$ equilibration technique at 25 °C (Epstein and Mayeda 1953), followed by analysis on an isotope mass spectrometer.

All available precipitation, groundwater, and surface-water stable isotope data from eastern Idaho and western Yellowstone National Park were obtained (Rightmire and Lewis 1987, Wood and Low 1988, Thordsen *et al.* 1992, Rye and Truesdell 1993, Bartholomay and Edwards 1994, Mann and Low 1994, Ott *et al.* 1994, Bartholomay *et al.* 1995, 1996, Knobel *et al.* 1999). Oxygen-18 and deuterium data were plotted to produce a local meteoric water line. Stable isotope data

**Table 2.** *$\delta^{18}O$ and $\delta D$ values of Henry's Fork spring water, snow cores, and summer precipitation at prescribed elevations and geographic coordinates. Values are reported in per mill relative to Vienna Standard Mean Ocean Water (VSMOW). Two sigma uncertainty of oxygen isotopic results = 0.2 ‰, and of hydrogen isotopic results = 2.0 ‰.*

| | Water Samples | | | Snow Core Samples | | | Summer Precipi tation | | |
|---|---|---|---|---|---|---|---|---|---|
| Site | Date | $\delta D$ | $\delta^{18}O$ | Date | $\delta D$ | $\delta^{18}O$ | Date | $\delta D$ | $\delta^{18}O$ |
| Big Springs | JUL 99 | -135.66 | -18.23 | MAR 99 | -141.43 | -18.94 | | | |
| 1951 m | JUL 99 | -134.62 | -18.29 | MAR 00 | -143.80 | -19.15 | | | |
| 44.50N; 111.25W | SEP 99 | -135.80 | -18.28 | | | | | | |
| Lucky Dog | JUL 98 | -134.31 | -18.33 | MAR 99 | -137.60 | -18.45 | 1999 | -79.74 | -11.57 |
| 1951 m | SEP 99 | -134.95 | -18.31 | | | | | | |
| 44.48N; 111.25W | JUL 98 | -135.50 | -18.24 | | | | | | |
| Warm River | JUL 98 | -133.95 | -18.00 | MAR 99 | -129.54 | -17.46 | 1999 | -55.00 | -7.58 |
| 1780 m | JUL 99 | -133.14 | -17.94 | MAR 00 | -150.40 | -19.85 | | | |
| 44.21N; 111.25W | SEP 99 | -133.70 | -17.96 | | | | | | |
| Snow Creek | JUL 98 | -130.64 | -17.63 | MAR 99 | -129.61 | -17.40 | | | |
| 2134 m | JUL 99 | -131.01 | -17.83 | MAR 00 | -150.50 | -20.17 | | | |
| 44.26N; 111.12W | SEP 99 | -133.20 | -17.82 | | | | | | |
| Chick Creek | JUL 98 | -136.07 | -18.11 | | | | | | |
| 1926 m | JUL 99 | -134.10 | -18.16 | | | | | | |
| 44.44N; 111.26W | | | | | | | | | |
| Buffalo River | JUL 98 | -135.32 | -18.18 | | | | | | |
| 1926 m | JUL 99 | -132.94 | -18.11 | | | | | | |
| 44.43N; 111.27W | SEP 99 | -134.80 | -18.17 | | | | | | |
| Latham Springs | JUL 98 | -134.55 | -17.93 | MAR 99 | -137.56 | -18.59 | 1999 | -75.69 | -11.14 |
| 2325 m | | | | MAR 00 | -145.30 | -19.60 | | | |
| 44.45N; 111.15W | | | | | | | | | |
| Baker Draw | | | | MAR 99 | -134.22 | -17.99 | 1999 | -70.25 | -10.10 |
| 2386 m | | | | MAR 00 | -143.00 | | | | |
| 44.34N; 111.12W | | | | | | | | | |
| Black Bear | | | | MAR 99 | -137.58 | -18.69 | | | |
| 2438 m | | | | | | | | | |
| 44.50N; 111.13W | | | | | | | | | |
| Madison Plateau | | | | MAR 99 | -141.48 | -18.99 | | | |
| 2316 m | | | | | | | | | |
| 44.58N; 111.12W | | | | | | | | | |
| Black Canyon | | | | MAR 99 | -138.49 | -18.80 | | | |
| 2463 m | | | | | | | | | |
| 44.46N; 111.10W | | | | | | | | | |
| Glade Creek | | | | MAR 99 | -151.00 | -20.30 | | | |
| 2137 m | | | | | | | | | |
| 44.10N; 110.83W | | | | | | | | | |
| Grassy Lake | | | | MAR 99 | -142.32 | -19.10 | | | |
| 2214 m | | | | | | | | | |
| 44.13N; 110.80W | | | | | | | | | |
| Lewis Lake Divide | | | | MAR 99 | -140.94 | -18.98 | | | |
| 2393 m | | | | | | | | | |
| 44.20N; 110.67W | | | | | | | | | |
| Moose Creek | JUL 98 | -133.88 | -18.23 | | | | | | |
| 2073 m | | | | | | | | | |
| 44.45N; 111.45W | | | | | | | | | |
| Osborne Springs | JUL 98 | -130.96 | -17.79 | | | | | | |
| 1768 m | | | | | | | | | |
| 44.37N; 111.41W | | | | | | | | | |

from the Henry's Fork springs were plotted next to these data for comparison.

## Chlorofluorocarbon Analysis

The mean residence time of water discharged at a spring refers to the transport time of the volume of water in the aquifer, which is dependent on the hydrologic characteristics of the aquifer (Manga 1999). Chlorofluorocarbons are transient tracers that are increasingly being used to determine residence times (Cook *et al.* 1995, Szabo *et al.* 1996, Cook and Solomon 1997). Chlorofluorocarbons are entirely anthropogenic in origin, were first manufactured in the 1930s, and are detectable in groundwater recharged after 1940 (Busenberg and Plummer 1992). Measurements of atmospheric concentrations have been made since 1978 and show few spatial variations globally; thus, a reliable input function exists. The concentration of CFCs in groundwater is a function of the atmospheric partial pressure of the CFCs at the time of recharge, when meteoric water was isolated from the atmosphere, and the CFC solubility, which is a function of temperature and alinity (Warner and Weiss 1985). Two chlorofluorocarbons, CFC-11 and CFC-12, were used in this study.

Samples of water for chlorofluorocarbon analysis were collected in copper tubes at Big Springs, Lucky Dog, Warm River, and Snow Creek by pulling water through the tubes with a hand pump and a vacuum flask. The apparatus was flushed at each site with water from the spring to prevent site-to-site contamination. Samples were kept away from any sources of CFCs such as refrigerators and air conditioning units during transport and storage; chlorofluorocarbon groundwater dating is highly sensitive to contamination from local sources of CFCs or equipment. Samples of water from each site were sent to D. K. Solomon at the University of Utah for analysis.

Recharge year for each spring was calculated from the concentrations of CFC-11 and CFC-12 at the site. The expected CFC concentration in water discharged at the spring in a given year can be estimated from the equation

$$C_{out}(t) = \int_0^t C_{in}(\tau) g(t-\tau) d\tau, \quad (3)$$

where $C_{out}(t)$ is the expected CFC concentration in the spring water, $C_{in}(\tau)$ is the known CFC concentration in water entering the aquifer in year $\tau$, and $g(t-\tau)$ is a probability density (weight) function describing the fraction of spring discharge at time *t* resulting from water that entered the aquifer at some prior time $\tau$ (Maloszewski and Zuber 1982).

The mean time, $T_{age}$, required for water to travel through the aquifer is given by the mean of the distribution $g(t)$, that is,

$$T_{age} = \int_0^t \tau \, g(\tau) d\tau. \quad (4)$$

We can consider the weight function $g(t)$ to be characterized by this travel time parameter $T_{age}$. Thus, for known input concentrations $C_{in}(\tau)$ a given weight function $g(t)$ (parameterized by $T_{age}$), and a fixed output year *t*, equation (3) describes the expected output concentration $C_{out}(t)$ in year *t* as a function of $T_{age}$. An observed output concentration in a given year can then be compared to the expected output in that year calculated by equation (3) to yield an estimate of $T_{age}$. The mean recharge year is given by $t - T_{age}$.

The output concentrations predicted by equation (3) depend substantially on the choice of $g(t)$ which reflects the physics of the particular aquifer flow model assumed. Two commonly used flow models are the piston model and the well-mixed (exponential) model (Manga in press). The piston model assumes that water entering the aquifer during a given year does not mix with water entering during other recharge years. In this model, all water entering the aquifer requires the same amount of time to travel through the aquifer. This fixed time is the mean recharge time $T_{age}$, and the only contribution to discharge in year $t$ is from water that entered the aquifer at time $t - T_{age}$. In this case, the weight function is given by

$$g(t) = \delta(t - T_{age}),$$

where $\delta(t)$ is the Dirac delta distribution, and equation (3) yields

$$C_{out}(t) = C_{in}(t - T_{age}) \qquad (5)$$

For a fixed output year $t$, the functional dependence of output concentration on the parameter $T_{age}$ is immediately apparent. The right-hand side of equation (5) is simply the input concentration in the recharge year $t - T_{age}$. Graphs of $C_{out}$ (1999) as a function of recharge year for CFC-11 and CFC-12 calculated from equation (5) and known input concentrations are shown in Figure 4. A recharge temperature of 5 °C based on noble gas concentrations and recharge elevation of 2133.6 m were assumed for computing the recharge year (D. K. Solomon, University of Utah, personal communication).

The well-mixed or exponential flow model assumes that aquifer travel times are exponentially distributed with mean $T_{age}$. Thus, spring discharge at time $t$ is influenced by water that has entered the aquifer during all years prior to $t$, with the degree of influence decreasing exponentially with time. In this model, the weight function is given by

$$g(t) = \frac{1}{T_{age}} \exp\left(-\frac{t}{T_{age}}\right) \qquad (6)$$

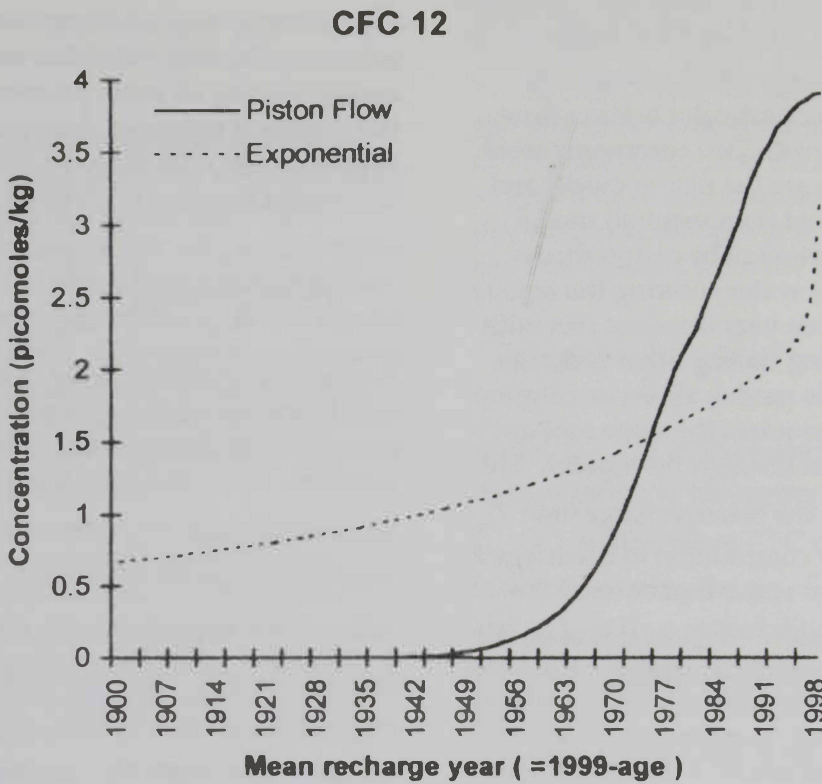
Explicit computation of the right-hand side of equation (3) is generally not possible with the exponential weight function. However, for a fixed $t$ and known values of $C_{in}(\tau)$ the integral in equation (3) can be computed numerically to generate $C_{out}(t)$ as a function of $T_{age}$. Expected 1999 output concentrations as a function of mean recharge year 1999 - $T_{age}$ calculated in this way are shown for CFC-11 and CFC-12 in Figure 4.

## Recharge Area

The size of the recharge area of each spring system was estimated using established values for spring discharge, local precipitation, and evapotranspiration in the following equation for a water budget in a watershed

$$P = ET + R + G + \Delta S \qquad (7)$$

where $P$ is precipitation, $ET$ is evapotranspiration, $R$ is surface runoff, $G$ is groundwater recharge, and $\Delta S$ is the change in storage (Manga 1997). Change in storage for the short time period under consideration here would make a relatively small contribution to the total water budget and this term was therefore omitted. I assumed that each spring system was fed by groundwater; thus, $G$ was estimated by the measured discharge in the springs. Surface runoff in the study area is negligible with no evidence of perennial streams or



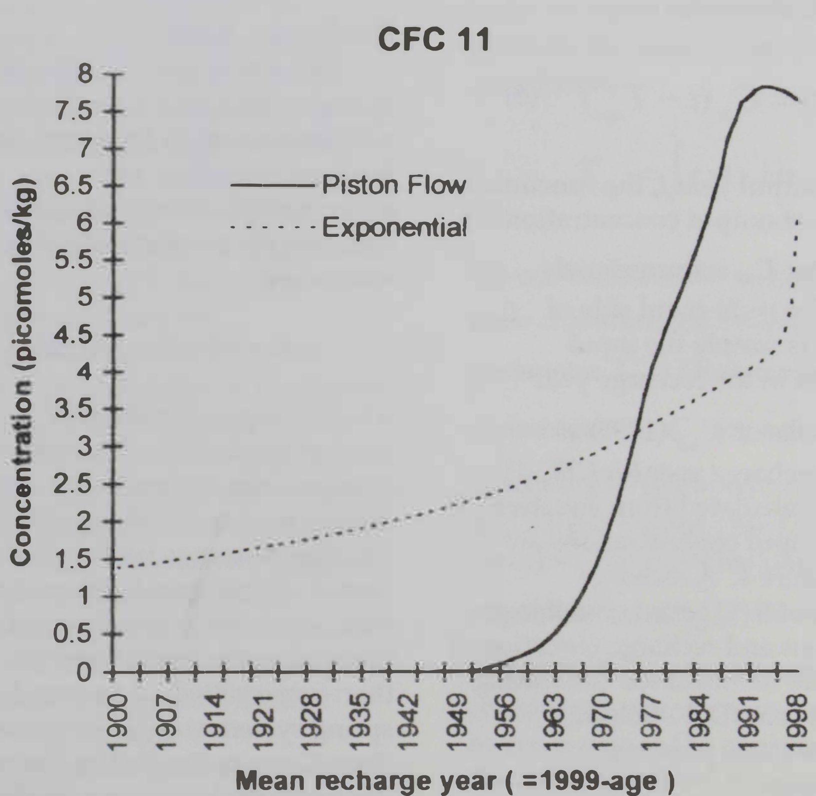




**Figure 4.** *Expected concentrations of CFCs 12 and 11 discharged in spring water as a function of mean recharge year derived using both exponential and piston-flow models.*

**Table 3.** *Mean annual precipitation rates for Snotel sites in the Henry's Fork region (1961 to 1991).*

| Site | Elevation (m) | Precipitation (mm) |
|---|---|---|
| Black Bear, MT | 2423 | 1567 |
| Madison Plateau, MT | 2362 | 1084 |
| Whiskey Creek, MT | 2072 | 929 |
| Big Springs, ID | 1981 | 779 |
| Ashton, ID | 1584 | 523 |
| Island Park, ID | 1917 | 767 |
| White Elephant, ID | 2350 | 1219 |
| Lewis Lake Divide, WY | 2392 | 1447 |
| Grassy Lake, WY | 2214 | 1422 |

region obtained from the NRCS and Dirks and Martner (1982) (Table 3, Fig. 5). Potential evapotranspiration values were obtained from Martner (1986) and Wyoming Water Development Commission and University of Wyoming (1990). Given mean annual precipitation and evapotranspiration rates, an estimate of the area needed to supply enough water for each spring system's discharge was calculated from equation (7). I used a 1:100,000 scale topographic map of the region and a 50-cm by 35-cm transparent grid with 1-cm squares to estimate recharge areas for each of the spring systems based on the areas calculated above and regional surface topography.

## Heat Flux

The total heat, $H$, discharged by each of the Henry's Fork springs was calculated using the equation

$$H = \rho C q \Delta\theta \qquad (8)$$

in which $\rho$ and $C$ are the density and heat capacity of water, $q$ is the discharge from the spring, and $\Delta\theta$ is the change in temperature of the groundwater between recharge and discharge (Manga 1998, in press). It is assumed that groundwater movement in the aquifer advects all the heat horizontally. The

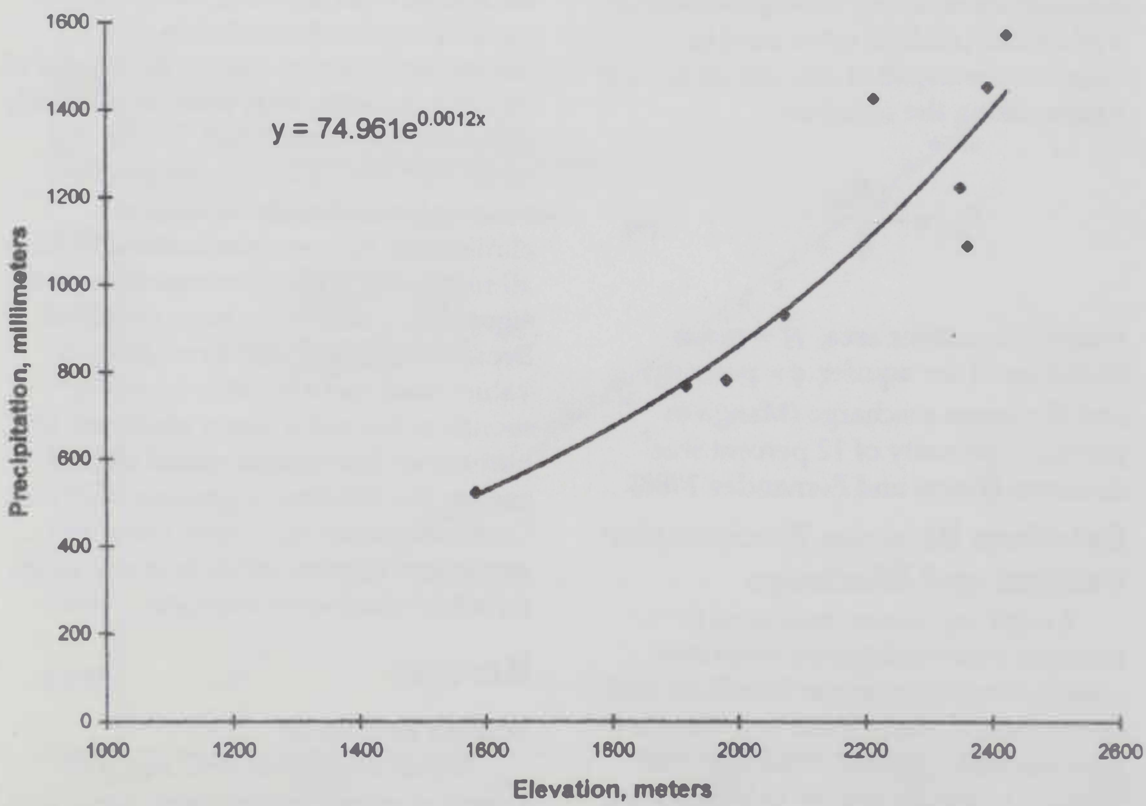



**Figure 5.** *Mean annual precipitation as a function of altitude in western Yellowstone National Park and eastern Island Park, Idaho.*

heat capacity of water, $q$ is the discharge from the spring, and $\Delta\theta$ is the change in temperature of the groundwater between recharge and discharge (Manga 1998, in press). It is assumed that groundwater movement in the aquifer advects all the heat horizontally. The mean heat flux entering the base of the aquifer is equal to the total heat flux divided by the surface area of the aquifer.

Mean temperature of recharge (at 2,195 to 2,377 m) was estimated to be 5 °C based on recharge temperatures calculated from noble-gas analysis (D. K. Solomon, personal communication). Spring water temperatures measured with data-loggers continually between 1997 and 1998 were essentially constant at all of the springs (Table 1).

## Depth of Circulation

The residence times of water, $T_{age}$, in the aquifers feeding the Henry's Fork springs (derived from CFC concentrations using the exponential or well-mixed models) were used to calculate the depth of circulation in each aquifer using the equation

$$T_{age} = \frac{A\overline{H}\phi}{\overline{q}} \qquad (9)$$

where $A$ = surface area, $\underline{H}$ = mean thickness of the aquifer, $\phi$ = porosity, and $\overline{q}$ = mean discharge (Manga in press). A porosity of 12 percent was assumed (Wood and Fernandez 1988).

## Relations Between Precipitation Patterns and Discharge

Linear regression was u ed to examine relationships between total annual discharge, winter baseflow, and spring runoff and annual and seasonal precipitation. Annual discharge was defined as annual inflow to Island Park Reservoir, winter baseflow was considered to be inflow to Island Park Reservoir for the months October through March, and spring runoff was considered to be inflow to Island Park Reservoir during the months of April, May and June (Benjamin 1997). Independent variables were annual precipitation, winter precipitation, and summer precipitation (obtained from the Idaho State Climate Center, Moscow). I considered winter precipitation as that falling during the months of October through April and considered summer precipitation as that falling during the months of July, August, and September. For each combination of dependent and independent variables, e.g., winter baseflow as a function of winter precipitation, multiple linear regression was used to analyze the functional dependence of the discharge variable on values of the precipitation variable for the current and all previous water years. Because dependence of discharge on precipitation in previous years generally decreased with time lag, independent variables were eliminated in chronological order (oldest first) until all regression coefficients were significantly different from zero ($t$-test, $P < 0$). All significant multiple regressions were recorded, and the chronological elimination process was continued until all remaining regression equations with significant coefficients were obtained. Because discharge and precipitation values were not available for every month in the water years analyzed, the number of data points varied slightly among the different regression analyses. Each individual regression used the maximum number of these water years for which data were available.

# Results

## Stable Isotopes

Trends evident in $\delta^{18}O$ and $\delta D$ values of spring water, snow cores, and summer precipitation in the Henry's Fork region (Table 2) included: 1) little springs and the same as spring water at Snow Creek; 4) stable isotope values of all spring waters and snow cores were significantly lighter than of summer precipitation; 5) stable isotope values of summer precipitation were heaviest at Warm River and became progressively lighter to the north; and 6) the lightest stable isotope values of snow cores were those on the eastern side of the Continental Divide.

Stable isotope values of precipitation in the Henry's Fork region plotted close to the GMWL and were similar to other values in eastern Idaho (Fig. 6). Relative enrichment in heavy isotopes of summer precipitation compared to winter precipitation was evident; all samples with $\delta^{18}O$ greater than -15 ‰ were collected during summer. Water from the Henry's Fork springs also plotted close to the GMWL (Fig. 7). The similarity between snowpack and Henry's Fork spring water isotopic compositions indicates that winter precipitation was the dominant source of recharge to the springs.

## Chlorofluorocarbons

Measured concentrations of CFCs from each of the springs (Table 4) related to expected concentrations (Fig. 4) yielded recharge years ranging from pre-1900 to 1994 (Table 4). The piston-flow and exponential models derived quite different recharge years (Table 4). Given the large size of the aquifers, the stable-isotope values, and results of my analysis of streamflow and precipitation (see below), the exponential model provided the most accurate representation of recharge processes. Warm River discharged water older than 100 years through a large regional aquifer system in complex geologic strata. Lucky Dog and Big Springs discharged more modern water (circa 1950 and 1960, respectively) from a regional aquifer through the Buffalo Lake flow. Snow Creek discharged

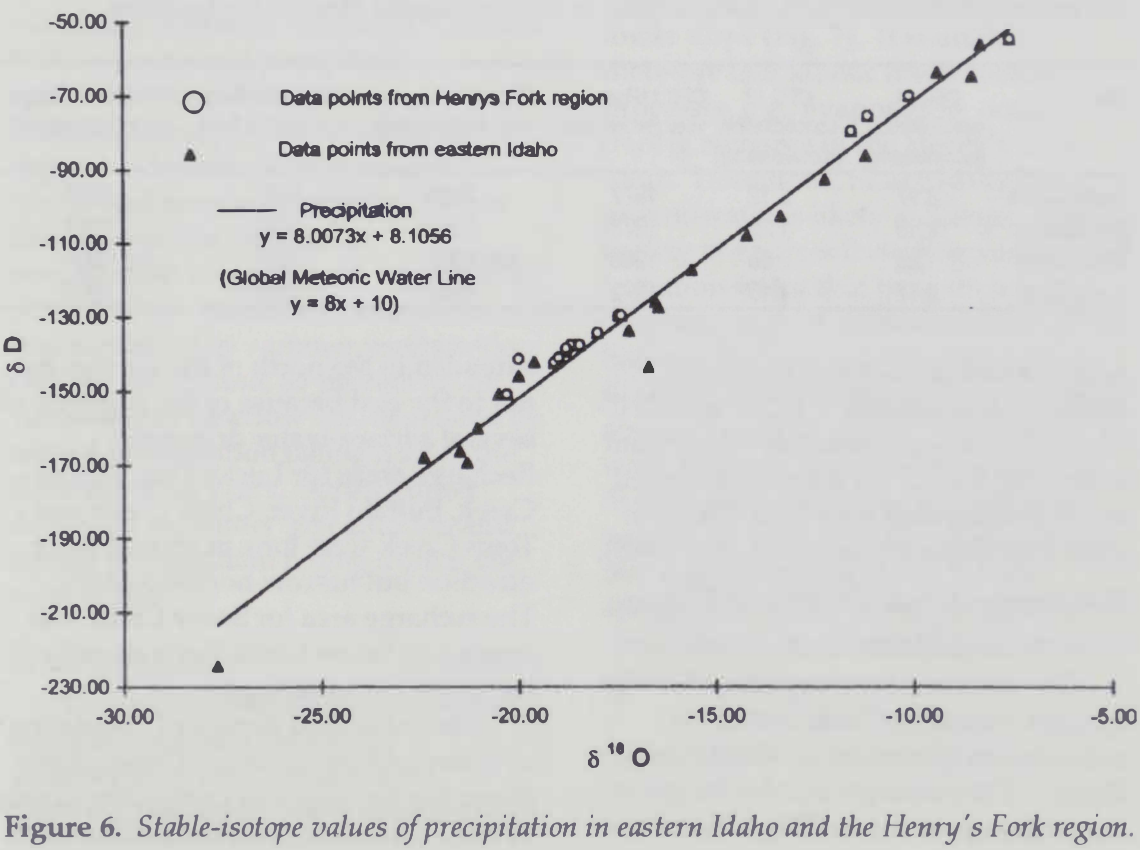



**Figure 6.** *Stable-isotope values of precipitation in eastern Idaho and the Henry's Fork region.*

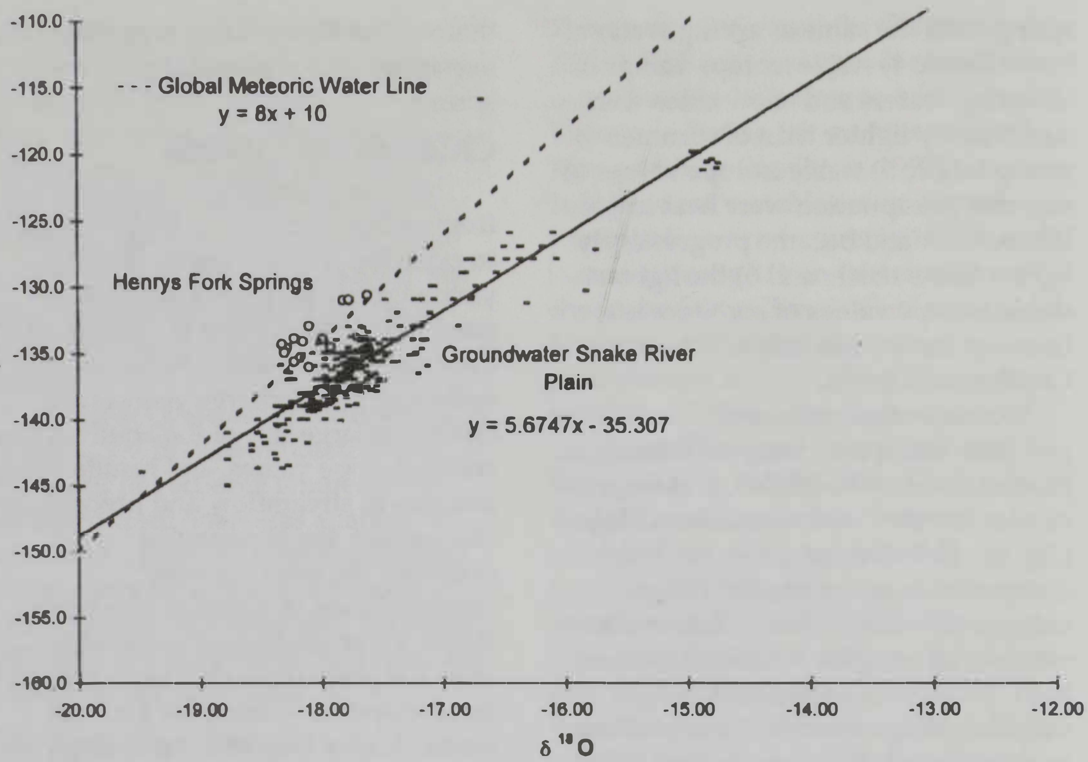



**Figure 7.** *Stable-isotope values of groundwater on the Snake River Plain and from the Henry's Fork springs.*

**Table 4.** *Concentrations of CFC 11 and 12 and recharge years, calculated using-piston flow and exponential models, for water emerging in 1999 at selected Henry's Fork springs.*

| Site | CFC 11 concentration (picomoles/kg) | CFC 12 concentration (picomoles/kg) | CFC 11Recharge year (piston flow) | CFC 11 Recharge year (exponential) | CFC 12 Recharge year (piston flow) | CFC 12 Recharge year (exponential) |
|---|---|---|---|---|---|---|
| Snow Creek | 3.97 | 2.12 | 1977 | 1990 | 1982 | 1997 |
| Big Springs | 2.55 | 1.29 | 1973 | 1959 | 1973 | 1964 |
| Warm River | 1.28 | 0.66 | 1968 | pre-1900 | 1967 | pre-1900 |
| Lucky Dog | 2.17 | 1.13 | 1972 | 1947 | 1972 | 1953 |

recent water (mid-1990s) through a shallow, local aquifer. CFC-11 and CFC-12 ages were approximately concordant indicating that CFCs were not retarded in the unsaturated zone by sorption or microbial degradation (Cook *et al.* 1995).

## Recharge Area, Depth of Circulation, and Heat Flux

The estimated recharge area for Big Springs was large (Table 1) and extended to the east of the Continental Divide. The recharge area for Warm River springs was even larger and extended to the north of the springs but not to the east because of the presence of several surface-water drainages. Recharge areas for Lucky Dog, Moose Creek, Buffalo River, Chick Creek and Toms Creek were long in an east-west direction but narrow north-to-south. The recharge area for Snow Creek was located on Snow Creek Butte directly to the north of the springs.

The calculated depths of circulation of water recharging Snow Creek, Warm River, Big Springs and Lucky Dog (Table 1) were consistent with the estimated thickness of the Buffalo Lake flow (300 m). Warm River, Big Springs and Lucky Dog are the result of a large, regional flow system, whereas Snow Creek is produced by a local, shallow flow system. Mean heat flux rates from the Henry's Fork springs (except Snow Creek) were relatively high, ranging from 400 to 800 mW/m$^2$ (Table 1).

## Precipitation and Streamflow Relationships

Annual and winter precipitation influenced discharge to a greater degree than did summer precipitation (Table 5). Discharge depended significantly on precipitation values as far back as three water-years previous; I found no significant relationships between discharge and precipitation falling more than four years previous. Annual discharge depended on annual, winter, and summer precipitation to some degree, with the strongest influence coming from annual precipitation during the current and two previous years. These precipitation variables accounted for 43 percent of the variability in annual discharge. Current water-year annual precipitation explained only 27 percent of variation in annual discharge. Winter baseflow depended to some degree on annual, winter and summer precipitation, but the dependence on summer precipitation was minimal. Less than 15 percent of the variability in baseflow was explained by summer precipitation. One the other hand, 55 percent of the variability in baseflow was explained by annual precipitation falling during the three previous years. Spring runoff depended on winter precipitation and annual precipitation falling during the current year.

# Discussion

The stable isotope values of spring water did not vary seasonally or annually between 1998 and 1999 implying that water being discharged at the springs was well mixed. Warm River Springs, Big Springs and Lucky Dog spring water all showed similar $\delta D$ and $\delta^{18}O$ values to those of snow cores taken at higher elevations on the Madison Plateau and significantly different values from those of snow cores taken at the same elevation as the springs. Therefore, recharge takes place on the plateau above the springs. The stable isotope values of summer precipitation were 7 to 10 ‰ ($\delta^{18}O$) and 60 to 75 ‰ ($\delta D$) heavier than winter values indicating that recharge was from snowmelt rather than summer rainfall events (Table 2). Regression analysis (Table 5) also indicated that summer precipitation does not play a significant role in spring discharge. Stable isotope values of snow cores at Snow Creek were similar to the water discharged at the springs indicating that recharge to those springs occurs quite locally.

Stable isotope values of water from the Henry's Fork springs plot on the GMWL, in contrast those of the Snake River Plain groundwaters, which have a lower slope (Fig. 7). It is unlikely therefore that kinetic fractionation processes, e.g., evaporation, occur during recharge in the Henry's Fork basin. In contrast, Snake River Plain groundwater is likely recharged from a variety of sources including surface and irrigation water that have undergone evaporation, and meteoric water, which has not. The water that emerges from the Henry's Fork springs does not show any shift in $\delta^{18}O$ values characteristic of water that has circulated to great depths, been heated, or reacted with oxygen in rocks. It is therefore unlikely that water from the Henry's Fork springs undergoes the same depth of circulation that water in Yellowstone thermal systems does.

Precipitation in the Henry's Fork region became increasingly lighter isotopically from south to north in both winter 1998-1999 and summer 1999,

**Table 5.** *Summary of results of regression analysis of precipitation-discharge relationships on the Henry's Fork of the Snake River at Island Park. P-values are those indicating the significance of the regression (F-test).*

| $y$ ($10^6$ $m^3$) | $x_1$ (cm) | $x_2$ (cm) | $x_3$ (cm) | regression equation | $n$ | $r^2$ | $P$ |
|---|---|---|---|---|---|---|---|
| annual discharge | annual precipitation | | | $y = 3.15x_1 + 332$ | 53 | 0.27 | 7.35E-05 |
| annual discharge | annual precipitation | annual precipitation 1 year ago | | $y = 2.68x_1 + 2.22x_2 + 200$ | 52 | 0.38 | 7.01E-06 |
| annual discharge | annual precipitation | annual precipitation 1 year ago | annual precipitation 2 years ago | $y = 2.21x_1 + 1.93x_2 + 1.62x_3 + 135$ | 52 | 0.43 | 4.36E-06 |
| annual discharge | winter precipitation | | | $y = 3.33x_1 + 402$ | 54 | 0.24 | 1.64E-04 |
| annual discharge | winter precipitation | winter precipitation 1 year ago | | $y = 2.90x_1 + 2.28x_2 + 310$ | 54 | 0.34 | 2.13E-05 |
| annual discharge | summer precipitation | summer precipitation 1 year ago | summer precipitation 2 years ago | $y = 5.02x_1 + 4.89x_2 +5.84x_3 + 383$ | 54 | 0.22 | 5.34E-03 |
| winter baseflow | annual precipitation 1 year ago | | | $y = 1.46x_1 + 112$ | 54 | 0.42 | 1.20E-07 |
| winter baseflow | annual precipitation 1 year ago | annual precipitation 2 years ago | | $y = 1.26x_1 + 0.77x_2 + 68$ | 53 | 0.51 | 1.74E-08 |
| winter baseflow | annual precipitation 1 year ago | annual precipitation 2 years ago | annual precipitation 3 years ago | $y = 1.10x_1 + 0.66x_2 + 0.51x_3 + 51$ | 52 | 0.55 | 2.14E-08 |
| winter baseflow | winter precipitation 1 year ago | | | $y = 1.53x_1 + 144$ | 55 | 0.37 | 6.99E-07 |
| winter baseflow | winter precipitation 1 year ago | winter precipitation 2 years ago | | $y = 1.38x_1 + 0.82x_2 + 111$ | 55 | 0.47 | 6.36E-08 |
| winter baseflow | summer precipitation 1 year ago | | | $y = 1.88x_1 + 198$ | 55 | 0.08 | 3.87E-02 |
| winter baseflow | summer precipitation 1 year ago | summer precipitation 2 years ago | | $y = 1.87x_1 + 1.79x_2 + 177$ | 55 | 0.15 | 1.51E-02 |
| spring runoff | annual precipitation | | | $y = 2.27x_1 + 35$ | 54 | 0.39 | 3.66E-07 |
| spring runoff | winter precipitation | | | $y = 2.60x_1 + 76$ | 55 | 0.42 | 1.06E-07 |

similar to the latitudinal trend in stable isotopes of precipitation in Yellowstone National Park (Rye and Truesdell 1992). This likely resulted from Pacific storm trajectories that travel up the Snake River Plain from southwest to northeast and progressively become enriched in lighter isotopes. In contrast, snow cores collected in 2000 showed an opposite trend and were isotopically lighter than those collected in 1999 (Fig. 8) suggesting that the majority of storms originated in the Arctic and traveled from north to south through the Henry's Fork region. El Niño and La Niña likely affect stable-isotope values in Henry's Fork precipitation because in an El Niño year, e.g., the winter of 1998-1999, the majority of storms in the region come from the south, whereas northern storms predominate in a La Niña year (winter 1999-2000). Although trends of increasing depletion at higher altitudes have been observed in other regions (James *et al.* in press), I observed no such trend in this study.

Large regional flow systems were evident for all of the Henry's Fork springs except Snow Creek. Recharge areas for Lucky Dog, Chick Creek, Buffalo River springs and Toms Creek extend from west to east on the Madison Plateau above the springs; the close proximity of the springs to each other implies that their recharge areas are long in a west-to-east direction but narrow north-to-south. Recharge water travels through the Buffalo Lake rhyolite flow and emerges as a result of local controls in the Lava Creek Tuff. The recharge area for Big Springs is large, extending to the north to Reas Pass and to the east of the Continental Divide. The size of the recharge area of Big Springs necessitates that the groundwater divide does not follow surface topographic features. Big Springs emerges where the Buffalo Lake flow contacts the Lava Creek Tuff. The recharge area for Warm River springs generally lies to the north and east of the springs and it is likely that recharge occurs through Moose

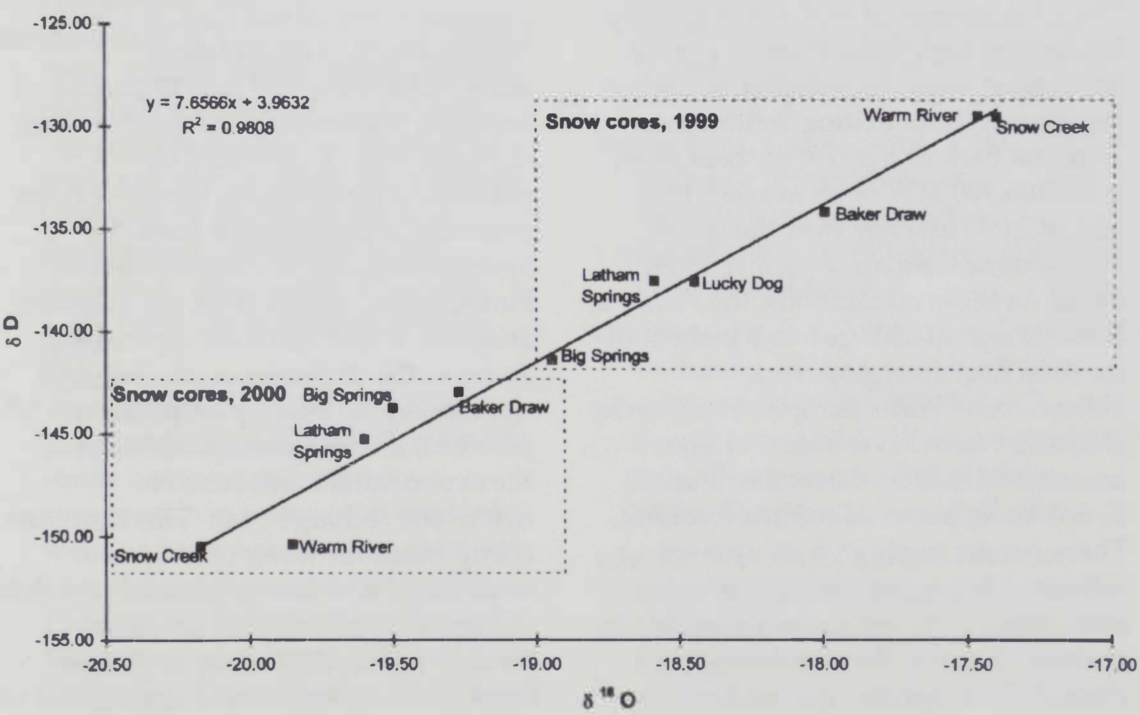



*Figure 8. Stable-isotope values of snow cores collected in the Henry's Fork region, March 1999 and 2000.*

Creek Butte, Lava Creek Tuff and Buffalo Lake flow rhyolites. Snow Creek has the smallest recharge area, which extends to the north of the springs on Snow Creek Butte. Calculated depths of circulation of water discharging at the springs (Table 1) correlate well with the known depth of the Buffalo Lake lava flow (300 m) and indicate an extensive regional flow system. Warm River recharge water appears to travel through several lava flows and has the greatest depth of circulation. Recharge age and depth of circulation indicate that a more local, shallower flow system is present at Snow Creek.

Large heat fluxes from all of the Henry's Fork springs except Snow Creek are consistent with their proximity to the Yellowstone "hot spot," but surprisingly large compared to other values in Island Park and on the eastern Snake River Plain. Heat-flux measurements range from 120 to 150 $mW/m^2$ on the northeastern margins of the Snake River Plain and 40 to 60 $mW/m^2$ at Island Park (Brott*et al.* 1981). The total amount of heat discharged by the Yellowstone hydrothermal system (calculated from chloride flux in each of the river systems exiting Yellowstone National Park as a proxy for heat flow) is 5,000-6,900 MW or an average heat flux of 2,000 $mW/m^2$ over the entire Yellowstone Caldera (Fournier 1989). Based on these calculations, the Henry's Fork springs discharge 6 to 8 percent of the total heat emerging from Yellowstone. Water samples from Lucky Dog and Warm River contain large amounts of mantle-derived helium (D. K. Solomon, personal communication). These results imply that an upward diffusive flux of mantle helium is interacting with water discharged by the springs. Regional flow systems of the Henry's Fork springs, except Snow Creek, apparently are interacting indirectly with the "hot spot" and are discharging much deeper (mantle) sources of heat and gases.

I used two different timescales to examine the question of recharge to the Henry' Fork springs. Chlorofluorocarbon analysis addressed the timescale of the residence time of water in the aquifer and statistical regression analysis of streamflow and precipitation addressed the timescale whereby changes in recharge are reflected in changes in discharge. Regression analysis showed that winter baseflows at Island Park were related to the three previous years' annual precipitation and that annual discharge was related to the two previous years' annual precipitation. This analysis provided a good characterization of the relationship between spring discharge and precipitation despite the fact that the analysis used streamflow and precipitation at Island Park rather than discharge from individual springs and associated precipitation at higher elevations. Spring flows comprise 75 percent of the streamflow at Island Park and precipitation increases on the plateaus above Island Park in a fairly linear pattern. The statistical relationships implied that 55 percent of baseflow discharge in the Henry's Fork at Island Park is explained by precipitation during the three previous years and that discharge from Big Springs probably is similarly driven. Furthermore, results from the statistical analyses, which show an exponential decay in the dependence of discharge upon previous years' precipitation (Table 5), are congruent with those of the exponential model used to determine recharge year. This timescale is important for water managers to understand and incorporate into models of Island Park Reservoir operation because it provides a way to predict response of spring-driven streamflow to changes in recharge.

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