Detecting Groundwater Discharge in the Clark Fork River near Stone Container Using Spectral Alpha Decay Detection for Dissolved Radon in Surface Water Samples Abstract

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Abstract

Radon-222 ($^{222}\text{Rn}$) was measured along 8.7 kilometers of the Clark Fork River, between Harper’s Bridge and Frenchtown, MT. Twelve water samples were taken along the stretch. Samples 1 through 4 and 10 through 12 were collected on a 1 km interval, samples 5 through 9 were taken on a 500 meter interval. Samples were analyzed for dissolved $^{222}\text{Rn}$ using a RAD7 spectral alpha decay detector. Instream $^{222}\text{Rn}$ was modeled to quantify groundwater discharge to the river. Literature on the Missoula Valley aquifer was analyzed, revealing an alluvial aquifer system to the east consisting of interbedded gravel, sand, silt, and clay. To the west, bedrock rises steeply from underneath the river to crop out at the surface. Analysis of the samples reveals that there are measurable quantities of $^{222}\text{Rn}$ through the entire stretch sampled, starting at 395 mBq/L near Harper’s bridge, with peaks of 950 mBq/L at 2 km and 632 mBq/L at 6.5 km. Lowest concentrations were 395 mBq/L at the start of sampling, 355 mBq/L at 4.3 km, and 336 mBq/L at 8.7 km. Modeling results averaged to 5.5×10^5 m$^3$/day of groundwater entering the river, with a standard deviation of 1.2×10^5 m$^3$/day, occurring in areas of high $^{222}\text{Rn}$ activity. This work identifies and quantifies the spatial distribution of groundwater discharge at the west end of the Missoula Valley postulated by previous works.
Introduction

Intermontane basins in the Northern Rocky Mountains often contain aquifer systems within their unconsolidated alluvium fill (USGS groundwater atlas). These alluvial aquifer systems are highly connected with adjacent rivers and streams. Estimating the location and volume of groundwater discharge to surface waters is important in constraining how groundwater and surface water interact in these areas. Understanding the connection between these systems is important for predicting how changes in surface water flow impacts aquifers, the effects of groundwater pumping on surface waters, and how contaminated water could be transferred between the two systems (Blomquist, 1991).

The Clark Fork River (CFR) is a gravel bed river in west-central Montana that runs through an urban environment. The CFR runs through many intermontane basins and integrates water from a variety of flow paths. Groundwater flow paths in mountain blocks are complex, and are dependent on topography, subsurface hydraulic conductivity, and structural configuration (Welch et al, 2012). The water table elevation loosely mirrors surface topography, increasing relief leads to more complex, longer flow paths (Töth, 1962 and 1963). While groundwater discharge tends to be concentrated at topographic low points in large valleys, subsurface configuration also plays a major roll in the distribution of discharge (Freeze and Witherspoon, 1967). Because subsurface configuration cannot be easily observed, other methods are required to determine where flow paths lead to discharge areas.

A common method of distinguishing flow paths is the use of natural-occurring and artificial tracers. Some tracers include radon-222 (^{222}\text{Rn}), terrigenous helium, chlorofluorocarbons, sulfur hexafluoride, stable oxygen isotopes, temperature, electrical conductivity, and major ion chemistry (Sklash et al, 1976; Rademacher et al, 2001; Becker et al, 2004; Gardner et al, 2011; McCallum et al, 2012). Groundwater tends to be enriched in dissolved ions, gathering ions with increasing residence time and becoming more conductive, increasing dissolved ion content and conductivity where it is present in surface waters.
Groundwater is also insulated from diurnal temperature shifts by the thermal inertia of the ground; thus, the temperature of groundwater remains constant near the mean annual temperature. Areas of the stream bed where groundwater is entering are influenced by the stable groundwater temperature and show less of a diurnal temperature change (Becker et al., 2004). Stable oxygen isotopes can be used to separate the rainfall component of streamflow, as the stable oxygen isotope ratios in rainfall can be easily distinguished from that of groundwater (Sklash et al., 1976). Sulfur hexafluoride and chlorofluorocarbons are anthropogenic gases with known historical concentrations within the last 70 years. The presence and age of groundwater can be constrained using these tracers (Rademacher et al., 2001; Smerdon et al., 2012). Terrigenous helium builds up in groundwater over time due to the radioactive decay of heavy elements and can be used to constrain groundwater age and discharge quantities in rivers (Gardner et al., 2011).

\[^{222}\text{Rn}\] is produced in the Uranium-238 decay series from the decay of Radium-226, and has a half-life of 3.82 days (Torgerson, 1980). \[^{222}\text{Rn}\] enters groundwater through direct diffusion and decay of dissolved minerals (Torgersen, 1980). After about two weeks in the groundwater system, radon concentration reaches a secular equilibrium; the rate of \[^{222}\text{Rn}\] entering the groundwater equals the rate of decay (Ellins et al., 1990). Uranium, and therefore Radium, is present in most crustal rocks. The amount of \[^{222}\text{Rn}\] in groundwater is determined by aquifer mineralogy, uranium concentration, aquifer permeability, and aquifer transport characteristics in a given place (Ball et al., 1991). Granitic sediments from the Bitterroot range and Volcanic sediments in underlying tertiary rocks in the Missoula Valley are a probable source of \[^{222}\text{Rn}\] in Missoula Valley groundwater. Ward (1997) performed a survey of \[^{222}\text{Rn}\] levels in wells drilled in the Missoula Valley for health reasons and found that radon levels were highest in the Rattlesnake valley, near the airport, and near valley margins.

Numerical models of instream \[^{222}\text{Rn}\] have been used to locate and quantify groundwater discharge to rivers (Cook et al., 2003). \[^{222}\text{Rn}\] exists in high concentrations in phreatic zone water.
and low concentrations in vadose zone waters, where it is aerated and allowed to equilibrate with the atmosphere (Genereux and Hemmond, 1990). Concentrations of $^{222}$Rn in groundwater are often orders of magnitude higher than instream $^{222}$Rn concentrations; $^{222}$Rn concentrations decline rapidly downstream of groundwater discharge areas due to gas exchange with the atmosphere (Ellins et al, 1990). Instream $^{222}$Rn exchanges between surface waters and the hyporheic zone can introduce $^{222}$Rn into surface waters where there is not groundwater discharge and can cause significant errors in discharge estimates if not accounted for (Cook et al, 2006). Modeling of instream $^{222}$Rn has been successfully applied by multiple authors to basins with different geology, including the Missoula Valley (Gardner et al, 2011; Smerdon et al, 2012; Horne, 2017).

In this study, we will use synoptic surface water samples from the CFR to test for the $^{222}$Rn activity in the CFR on an 8.7 kilometer stretch below Harper’s bridge west of Missoula, Montana. We will use the numerical methods outlined by Cook et al (2006) to quantify the discharge of groundwater into the CFR in the study area. The purpose of this study is to better understand the spatial distribution and quantity of groundwater discharge. Groundwater discharge in this area has been suggested by previous research, but never successfully quantified (Smith, 1992).
The area selected for this study lies on the western edge of the Missoula Valley, 8.7 km downstream of the Harper’s Bridge fishing access site, past the former Smurfit-Stone Pulp Mill site. The CFR runs north-northwest through this area adjacent to a prominent bedrock rise to the west, with many exposed outcrops and cliff bands. East of the CFR there is agricultural and industrial land, including the former Smurfit-Stone pulp mill adjacent to the CFR for most of the run. Further east lies the populated portion of the Missoula Valley.

The CFR has been actively migrating through this area, relic river landforms are present to the east of the CFR on the valley floor. For much of the reach, an earthen berm constructed
from large boulders separates the CFR from the Smurfit-Stone mill site. Much of the mill site is within the 100-year floodplain of the CFR. To the west, the CFR is adjacent to steep bedrock outcrops, except where short sections of alluvium separate the CFR from the bedrock. Three tributaries, Deep Creek, Albert Creek, and Rock Creek enter the CFR along this reach (figure 1). None were flowing during the time of the study.

Figure 2. Aerial photograph of the study area overlain with geologic mapping from Lewis (1998).

The Missoula Valley is a geologically heterogeneous area which has been described in detail by many authors. Bedrock geology around western Missoula has been mapped and detailed by Lewis (1998) and Hall (1969), partially displayed in figure 2. The bedrock walls of the valley consist primarily of the Precambrian Missoula Group, part of the Belt Supergroup. These rocks consist of 1.2 to 1.4 billion year old metasedimentary quartzites, argillites and dolomites. They have been altered since their deposition; most primary porosity has been lost. Relative to the adjacent alluvium, little flow is occurring where these rocks are not fractured.

Along the western edge of the valley, the early Cambrian Red Lion, Hasmark, Silver Hill, and Flathead formations crop out (figure 2). The Red Lion formation consists of interbedded siltstone, dolomite, and silty/muddy laminated dolomite. It is underlain by the Hasmark formation, consisting of a grey dolomite with scattered chert. Underlying the Hasmark formation is the Silver Hill formation. The Silver Hill formation is primarily grey limestone, with an interbedded shale-sandstone member. The bottom unit in the local Cambrian stratigraphy is the Flathead formation, a vitreous quartzite. Beneath the Flathead formation is an erosional unconformity underlain by the Belt Supergroup.

According to cross sections and mapping in Lewis (1998) and Hall (1969), Cambrian bedrock in the study area dips to the west-southwest by 10° to 12°. The Cambrian strata have been truncated by erosion along the valley margin, causing the bedrock surface to slope downwards. The Hasmark and Red Lion formations crop out west of the study area. Beneath the surface the Silver Hill formation and Flathead sandstone are truncated at an angle and are in contact with the valley fill. West of the study area, the section of Cambrian rock is truncated by the Albert Creek thrust, placing Belt Supergroup rocks on top of Cambrian rocks.

The area has undergone great structural alteration since these rocks were laid down, generating a structurally complex area. Compression during the cretaceous and extension during the tertiary has produced sets of northwest striking reverse and normal faults, which frequently separate mountain blocks. To the southwest of the Missoula Valley, the intrusion of
the Idaho Batholith during the late cretaceous produced a metamorphic core complex, with metamorphic grade increasing rapidly south of Lolo creek. Erosion has exposed large amounts of granitic rocks, creating the modern Bitterroot Range.

Smith (1992) performed a detailed investigation into the subsurface stratigraphy and structure of the Missoula Valley alluvium, as well as provided a geologic interpretation. The Missoula Valley opened during the Laramide orogeny and was filled during the tertiary with fine grained and volcanic deposits. The quaternary sediments that unconformably overlie the tertiary sediments were deposited by a complex system of lacustrine, fluvial, alluvial, and colluvial processes. They represent deposition by the ancestral Clark Fork and Bitterroot rivers, and sedimentation from glacial outwash, alluvial fan, and lacustrine processes during the fillings and drainings of Glacial Lake Missoula. These sediments can be divided into 3 general units: a lower gravel unit, a middle silty sand unit, and an upper gravel unit. These units are heterogeneous and can contain great lateral variation due to former stream channel migration.

Below the CFR in the study area, there is an estimated 50-100 feet of unconsolidated alluvium overlying Belt supergroup bedrock (Smith, 2006). This unconsolidated material thickens eastward to up to 200 feet thick below the former mill site. The subsurface stratigraphy of the alluvium in this area specifically is described by Smith (1992) and is also documented by drillers logs in the area. The top 10 to 40 feet consists primarily of fine to coarse gravel with discontinuous lenses of sand and clay. The next unit is composed predominantly of sand interbedded with lenses of silt, clay, and gravel, between 50 and 100 feet thick. Below this layer there is a somewhat laterally continuous layer of silty clay, 3 to 7 feet thick. Underlying the clay layer is another unit of heterogeneous sandy gravel. Many of the wells in this area are finished in this unit, which is 40 to 45 feet thick according to Smith (1992). All units except for the uppermost described are present east of the CFR, where alluvium is up to 200 feet thick (Smith, 2006). They are likely pinched out below the CFR as the bedrock surface slopes upward and crops out west of the river.
The quaternary alluvium within the valley is the primary aquifer for Missoula; it is a shallow, unconfined aquifer near the surface with many laterally discontinuous confining units creating some division between the upper and lower gravel units (Smith, 1992). Tracking how water moves in the aquifer is important for proper management of the aquifer. Quantification of the groundwater component of streamflow can be used to track how much water is exiting the aquifer. Woessner (1988) estimates that the natural discharge of the aquifer is 93% of its recharge; the majority of water that enters the aquifer runs through it and exits at another point.

Figure 3. Potentiometric map of the Missoula Valley from LaFave (2006)
The CFR is the primary source of recharge for the aquifer; leakage provides 90% of the aquifer recharge (Woessner, 1988). The CFR is hydraulically disconnected from the aquifer for 4 to 6 miles after entering the Missoula Valley (Miller, 1991). La Fave (2006) constructed a potentiometric surface map from logged water depths in wells (figure 3). North of the CFR, the head gradient dips northwest, trending towards the northwestern end of the valley. South of the CFR, the head gradient dips southwest towards the Bitterroot River. Figure 3 also shows there are artesian zones both upstream and downstream of the study area, indicating that flow paths deep in the aquifer rise in these areas.

Figure 4. Part of the alluvium thickness map from Smith (2006) containing the Missoula Valley

Smith (2006) compiled well log data to produce a map of the alluvium depth beneath the Missoula, Bitterroot, and Ninemile valleys, partially displayed in figure 4. The map indicates the valley alluvium is over 300 feet thick beneath the southeastern end of the Missoula Valley, and
thins rapidly to nothing near the bedrock valley walls. From southeast to northwest, the width of valley covered by alluvium decreases from 10 km around Missoula to 3 km near the end of the study area. Maximum alluvium thickness decreases along the same trend, from over 300 feet thick at the southeastern end to about 200 feet thick near the study area. The decrease in alluvium width and thickness represents a reduction of aquifer volume in this area. A large bedrock knob in the center of the valley also precedes the study area, further reducing aquifer volume near the study area. At the northwestern end of the Missoula Valley, bedrock pinches out the alluvium into a 700 meter wide, 60 meter deep section. All water exiting the Missoula Valley not lost to bedrock leakage and evapotranspiration passes through this area.
Methods

Field Methods

Figure 5. Aerial photograph of the study area showing sampling points and survey course.

Sampling took place on September 25th, 2019. We put on at the Harper’s Bridge fishing access site and collected 12 samples along an 8.7 km reach downstream (figure 5). All samples were collected in the thalweg, except for one sample taken in a side channel at 3.2 km. Sampling intervals were approximately 1 km for the first and last 3 samples, we sampled approximately every 0.5 km for the middle three samples. We collected samples using a continuously running pump, rinsing and then filling 900 ml plastic bottles at sampling points. We
partially capped the bottles then immediately squeezed out the remaining air to ensure minimal
gas exchange during sampling.

**Analytical methods**

We analyzed samples for $^{222}$Rn activity using a Durridge RAD7 spectral alpha decay
detector. The RAD7 is capable of detecting radon and thoron ($^{220}$Rn) in air and water through
alpha decay detection. $^{222}$Rn produces polonium-218 when it decays, which has a half-life of
3.05 minutes. The polonium settles on the detector and emits an alpha particle when it decays,
which the detector reads. The detector measures the radon in the air, radon in water is
measured by connecting the sample to the RAD7 in a closed loop. Air is pumped through the
water via an aerator, degassing the radon into the air within the closed loop. Air from the sample
then runs through the alpha decay detection chamber. Once the sample has been stripped, the
radon-laden air is circulated through the RAD7 and the alpha decay detector measures the
$^{222}$Rn content of the air. The $^{222}$Rn concentration in the sample is then calculated based on the
measured $^{222}$Rn in the air and the volume within the closed loop. In this study, we allowed the
samples to degas for two 5-minute runs before closing off the aerator. The RAD7 then
performed four 5-minute runs to count the $^{222}$Rn decays in the air. These runs were then
averaged, excluding the first two when the sample was degassing, to find the $^{222}$Rn in each
sample.

This method of measuring $^{222}$Rn has two major sources of error that must be accounted
for: radioactive decay of $^{222}$Rn between sampling and testing and humidity within the RAD7.
Samples were analyzed within 48 hours to minimize $^{222}$Rn decay prior to testing. The following
formula was used for decay correction:

$$C_o = C_m e^{\frac{t}{132.4}}$$
where $C_o$ (Bq/L) is the $^{222}$Rn content when the sample was taken, $C_m$ (Bq/L) is measured $^{222}$Rn content when the sample was tested, $t$ (hours) is the time duration between sampling and testing.

High relative humidity reduces the efficiency of the detector by slowing the settling of polonium-218, allowing some to decay before it can be detected. High humidity will cause the RAD7 to read low. To reduce the humidity within the closed loop, an inline desiccant tube was used. The system was also purged for 10 minutes after each run to remove moisture from aerating the sample.

**Numerical Methods**

To model the results of $^{222}$Rn sampling, we used RADIN13, a 1-dimensional stream transport model. It calculates the $^{222}$Rn activity in the stream with distance as a function of groundwater inflow, hyporheic exchange, gas exchange, radioactive decay, and evaporation.

The model simulates the following equations from *Cook et al.* (2006) and used by *Horne* (2017):

\[
\frac{\partial Q}{\partial x} = I(x) - P(x) - E(x)
\]

\[
Q \frac{\partial c}{\partial x} = I(c_i - c) + wEc - kw\lambda c + dw\lambda c + \frac{\gamma hw\theta}{1 + \lambda t_h} + \frac{\lambda hw\theta}{1 + \lambda t_h} c
\]

where $Q$ (m$^3$/day) is river streamflow during the survey, $x$ (m) is distance along the direction of flow, $I$ (m$^3$/m/day) is groundwater inflow rate per unit of river length, $P$ (m$^3$/m/day) is the rate of water loss through direct pumping, $E$ (m/day) is the evaporation rate, $c$ (Bq/L) is the concentration of $^{222}$Rn in the river, $c_i$ (Bq/L) is the concentration of $^{222}$Rn in groundwater inflow, $w$ (m) is the width of the river, $k$ (m/day) is the gas transfer velocity across the river surface, $d$ (m) is the mean river depth, $\lambda$ (1/day) is the radioactive decay constant, $\gamma$ (Bq/L/day) is the production rate of $^{222}$Rn in the hyporheic zone, $h$ (m) is the mean depth of the hyporheic zone, $\theta$ is the hyporheic zone porosity, and $t_h$ (days) is the mean residence time for hyporheic zone water.
Modeling parameters were taken from previous work and literature on the area. Most parameters were taken from Horne (2017), whose study area was close by and very similar to this one. From Horne (2017), we use an evaporation rate of 5 mm/day, a constant hyporheic depth of 0.1 m, a mean hyporheic zone residence time of 0.25 days, a hyporheic production rate of 0.2 Bq/L/m, and a decay constant of 0.18 days\(^{-1}\) for \(^{222}\)Rn. Streamflow was taken from a USGS gauging station on the CFR after its confluence with the Bitterroot river. River widths were estimated from NAIP aerial photography at each sample point. Average depths were calculated from the widths and the streamflow for that day.

Two modeling parameters, \(k\) and \(c_i\), were not constrained by field parameters, and estimates from previous literature did not adequately fit modeled instream \(^{222}\)Rn concentrations to measured \(^{222}\)Rn concentrations. These two parameters were varied independently to constrain upper and lower bounds on groundwater discharge. We performed a sensitivity analysis on these parameters to see how they affected the final modeled groundwater discharge. Modeling parameters and river geometry are tabulated in tables 1 and 2.

**Table 1. Parameters used in RADIN**

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
<th>Units</th>
</tr>
</thead>
<tbody>
<tr>
<td>Evaporation Rate ((E))</td>
<td>5</td>
<td>mm/day</td>
</tr>
<tr>
<td>(^{222})Rn Decay Constant ((\lambda))</td>
<td>0.18</td>
<td>1/days</td>
</tr>
<tr>
<td>Hyporheic Zone Depth ((h))</td>
<td>0.1</td>
<td>m</td>
</tr>
<tr>
<td>Hyporheic Zone Residence Time ((t_h))</td>
<td>0.25</td>
<td>days</td>
</tr>
<tr>
<td>Hyporheic Zone Porosity ((\theta))</td>
<td>0.4</td>
<td></td>
</tr>
<tr>
<td>Hyporheic Zone (^{222})Rn Production Rate ((\gamma))</td>
<td>0.2</td>
<td>Bq/L/day</td>
</tr>
<tr>
<td>Streamflow ((Q))</td>
<td>74.76</td>
<td>m(^3)/s</td>
</tr>
<tr>
<td>Gas Exchange Velocity ((k))</td>
<td>25 +/- 5</td>
<td>m/s</td>
</tr>
<tr>
<td>(^{222})Rn Inflow Concentration ((c_i))</td>
<td>35 +/- 10</td>
<td>Bq/L</td>
</tr>
</tbody>
</table>
Table 2. River geometry used in RADIN

<table>
<thead>
<tr>
<th>Sample</th>
<th>Distance (m)</th>
<th>Width (m)</th>
<th>Depth (m)</th>
</tr>
</thead>
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<tr>
<td>1</td>
<td>0</td>
<td>115</td>
<td>0.65</td>
</tr>
<tr>
<td>2</td>
<td>885</td>
<td>113</td>
<td>0.66</td>
</tr>
<tr>
<td>3</td>
<td>1855</td>
<td>130</td>
<td>0.57</td>
</tr>
<tr>
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<td>2744</td>
<td>113</td>
<td>0.66</td>
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<td>5</td>
<td>3262</td>
<td>134</td>
<td>0.55</td>
</tr>
<tr>
<td>6</td>
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<td>101</td>
<td>0.74</td>
</tr>
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<td>112</td>
<td>0.66</td>
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<tr>
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<td>129</td>
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<td>10</td>
<td>6462</td>
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<tr>
<td>11</td>
<td>7679</td>
<td>104</td>
<td>0.71</td>
</tr>
<tr>
<td>12</td>
<td>8690</td>
<td>138</td>
<td>0.54</td>
</tr>
</tbody>
</table>

RADIN13 uses user input groundwater discharge to model instream $^{222}$Rn. Modeled instream $^{222}$Rn is fit to measured instream $^{222}$Rn by varying the amount and location of groundwater discharge. This method provided constraints on the quantity and spatial variation in groundwater discharge to the CFR. Cook et al (2006) performed a sensitivity analysis on this model to find which parameters it is most sensitive to. Due to uncertainty and assumptions made with parameters, our estimated discharge has error. Because the CFR is a shallow ($d < 5$ m) and wide river, gas exchange is the primary controller of radon loss to the atmosphere, while in a deeper river, radioactive decay would dominate. Large variability $k$, such as in this stretch of the CFR, is the largest source of error in $Q$, as the model assumes gas transfer is constant.

The model is also sensitive to $^{222}$Rn inflow concentrations. A lower $k$ requires more groundwater inflow to get to the same instream concentration as a higher groundwater inflow.
concentration with less groundwater inflow. A given section of stream could have a large variety of combinations of inflow concentrations and inflow rates, creating a problem of non-uniqueness. Because groundwater \(^{222}\text{Rn}\) concentrations have not been quantified in this area, we have varied the groundwater inflow concentrations and gas transfer velocities, then fit the model to the data. This method creates a range of inflow and gas exchange conditions that lead to the same instream distribution of \(^{222}\text{Rn}\).

Cook et al (2006) also found that because hyporheic zone residence times are usually low, radon contributions from the hyporheic zone are controlled more by the depth of the hyporheic zone than by residence time. Errors in hyporheic zone depth could cause large errors in \(Q\). The model is relatively insensitive to pumping losses if they are small compared to streamflow. Surface waters in our reach are not actively pumped to a large degree, so we assume pumping losses are 0.

Because \(^{222}\text{Rn}\) is a radioactive gas that decays over a short timeframe, there is no \(^{222}\text{Rn}\) present in the atmosphere. Therefore, surface water in equilibrium will have no radon content. \(^{222}\text{Rn}\) introduced into streams by groundwater will diffuse rapidly into the atmosphere; \(^{222}\text{Rn}\) will only be present close to points of groundwater inflow (Cook et al, 2003). In shallow streams, gas exchange is the primary method for \(^{222}\text{Rn}\) loss (Cook et al, 2006). Gas exchange can be highly variable in rivers with alternating runs of quiescent and agitated water or white water, and can rapidly change by orders of magnitude (Kokic et al, 2018; Hall et al, 2012). The scale length, which is the distance \(^{222}\text{Rn}\) decays to \(1/e\) of its original, described in Cook et al (2006), is:

\[
L = \frac{Q}{kw + dw\lambda}
\]

Where \(L\) is scale length (m), \(Q\) is river streamflow (m\(^3\)/day), \(k\) is gas transfer velocity from stream to atmosphere (m/day), \(d\) is mean stream depth (m), \(w\) is mean stream width (m), and \(\lambda\) is the radioactive decay constant for \(^{222}\text{Rn}\) (1/day).
Scale length ($L$) was evaluated to be between 1.5 km and 2.7 km. Because the CFR is relatively shallow for much of its width through the study area, more $^{222}\text{Rn}$ is lost to gas exchange than to radioactive decay. Using lower gas exchange constants lead to lower scale lengths. Scale length is representative of the distance $^{222}\text{Rn}$ is detectable downstream from the point it was input. In order for modeled groundwater discharge to be accurate, sampling intervals must be shorter than scale length. We sampled along intervals of 1 km and 500 m, less than half the shortest calculated scale lengths.

Results

Figure 6. Aerial photograph showing distribution and concentration of $^{222}\text{Rn}$ samples
Analysis of the samples revealed that there are measurable quantities of $^{222}\text{Rn}$ through the entire stretch sampled. $^{222}\text{Rn}$ concentrations started at 465 mBq/L near Harper’s bridge, with peaks of 1150 mBq/L at 2 km and 877 mBq/L at 6.5 km (figure 6). Lowest concentrations were 465 mBq/L at the start of sampling, 441 mBq/L at 4.3 km, and 470 mBq/L at 8.7 km. Instream $^{222}\text{Rn}$ rose quickly over the first 2 km’s of the survey, remained steady for 1 km, then dropped sharply over 200 m. The sample after the first peak (sample 5) was taken in a side channel containing a long riffle, which may have caused the sharp drop. Instream $^{222}\text{Rn}$ then rose more gradually for 2.5 km, then dropped gradually to the end of the survey course.

Figure 7. Modeled groundwater inflow and instream $^{222}\text{Rn}$ for $k=25$, $c_i=30$. Red squares represent measured instream $^{222}\text{Rn}$ concentrations, green line is modeled instream $^{222}\text{Rn}$.

The model variation with 25 m/s for $k$ and 30 Bq/L for $c_i$ fits the data well but does not greatly exceed values in literature. Radon inflow concentration for this variation was within 25%
of the highest groundwater $^{222}$Rn measurement of 24 Bq/L from Ward (1997), whose measurements varied between 8-24 Bq/L. Because of the heterogeneity of groundwater $^{222}$Rn concentrations in the aquifer, it is possible the study area could have a higher groundwater $^{222}$Rn content. Gas transfer velocity can vary by orders of magnitude in rivers with changing levels of aeration and turbulence (Hall et al., 2012). The studied section of river is shallow and has a diverse streambed morphology, leading to a large heterogeneity in $k$. Because $k$ is modeled homogenously, a representative value for the entire reach must be chosen. Large values of $k$ were easier to fit to the data; however, larger values would indicate a very turbulent, well aerated river. We used the lowest possible $k$ values that would still fit the data well, as the CFR in this reach lies somewhere in between a well aerated, highly turbulent stream with many large rapids, and a quiescent river with few disturbances.

The modeling results in figure 7 show that there are 2 primary discharge zones in the study area. The first discharge region begins before the study area with a discharge per unit length of 120 m$^3$/day/m from 0-1 km, 150 m$^3$/day/m from 1-1.5 km, 230 m$^3$/day/m from 1.5-2 km, 130 m$^3$/day/m from 2.5-3 km, dropping to 0 after 3 km. The 2nd area of discharge begins at 4.5 km with a discharge per unit length of 150 m$^3$/day/m from 4.5-5 km, 110 m$^3$/day/m from 5-6 km, 120 m$^3$/day/m from 6-6.5 km, 50 m$^3$/day/m from 6.5-7 km, 30 m$^3$/day/m from 7 to 8 km, dropping to 0 after 8 km. Between 4.5 km and 7 km, and after 8 km, no discharge was modeled.
Figure 8. Modeled groundwater inflow and instream $^{222}$Rn using end member $k$ and $c_i$ values. A: $k=20$ and $c_i=45$, B: $k=20$ and $c_i=30$, C: $k=30$ and $c_i=45$, D: $k=30$ and $c_i=30$. 
Parameter Sensitivity

Plots of groundwater discharge and modeled instream $^{222}$Rn for end member $k$ and $c_i$ variations are displayed in figure 8. Modeling results averaged to a total discharge of $5.55 \times 10^5$ m$^3$/day over the entire reach, with a standard deviation of $1.22 \times 10^5$ m$^3$/day. Highest modeled discharge was $8.0 \times 10^5$ m$^3$/day, lowest modeled discharge was $3.7 \times 10^5$ m$^3$/day. Increasing $k$ while holding $c_i$ constant caused discharge to increase, while increasing $c_i$ and holding $k$ constant caused discharge to increase.

Sensitivity of modeled discharge to variations of $k$ and $c_i$ are shown in figures 9 and 10. Model simulations show that $dQ/dk$ and $dQ/dc_i$ are dependent on each other. $dQ/dk$ increases with increasing $c_i$; for the lowermost line in figure 9, $c_i=45$ Bq/L, for the uppermost line $c_i=30$ Bq/L. $dQ/dc_i$ decreases with increasing $k$; for the uppermost line in figure 10 $k=30$ m/s, for the lowermost line $k=20$. Both of these trends are non-linear, $Q$ increases more rapidly as $k$ and $c_i$ grow. The absolute magnitude of $dQ/dk$ is larger than the absolute magnitude of $dQ/dc_i$, suggesting that the model is more sensitive to $k$ than it is to $c_i$.

![Total Discharge vs. Gas Transfer Velocity](image)

Figure 9. Total modeled discharge plotted against $k$. Average slope is 19447.66.
Figure 10. Total modeled discharge plotted against $c_i$. Average slope is -15228.50
Discussion

Figure 11. Aerial photograph showing distribution of groundwater discharge zones

Based solely on the quantity and variable distribution of $^{222}$Rn in the CFR through the study area, it is clear that groundwater is discharging to this section of river. Uncertainty in $k$ and $c_i$ creates problems solving the mass balance to quantify the amount of groundwater discharging to the CFR. Higher $k$ requires more groundwater discharge to the river to model the same amount of instream $^{222}$Rn, in order to make up for the increased escape of $^{222}$Rn to the atmosphere. Higher $c_i$ requires less discharge, as more $^{222}$Rn is being introduced per unit volume of discharge, allowing the model to simulate the same amount of instream $^{222}$Rn with less groundwater inflow.
Because we were unable to quantify these values, we used values that could provide an adequate fit to measured instream $^{222}$Rn concentrations, with some basis in literature. Gas transfer velocity can vary by orders of magnitude on stretches of river with riffles or whitewater runs, of which the study area contained many (Kokic et al., 2018). Riffles and whitewater would raise the average $k$ value through the study area, despite stretches of flat water with $k$ values that are probably similar to those used in Horne (2017) further upriver. Although values for $c_\text{i}$ used in modeling were higher than those found in Ward (1997), Ward does postulate that $c_\text{i}$ would be higher near valley margins. Additionally, the presence of ancestral Bitterroot River sediments on the western side of the valley could cause higher $c_\text{i}$ values than on the eastern side, as the Bitterroot River has a large, proximal source of material that would raise subsurface $^{222}$Rn equilibrium concentrations: the Bitterroot lobe of the Idaho Batholith.

Despite variations in $k$ and $c_\text{i}$, modeled groundwater discharge locations showed little change in spatial distribution, because $k$ is a property of the CFR and $c_\text{i}$ is a consequence of flow paths taken, rather than their controller. The primary control on groundwater flow paths are subsurface stratigraphy, hydraulic conductivity, and structural configuration. Descriptions of the area by Smith (1992) and Lewis (1998) indicate that the Missoula Valley alluvium can support groundwater discharge to the CFR in this area much better than the bedrock valley walls, which have a lower hydraulic conductivity and dip away from the river.

Potentiometric mapping from LaFave (2006) indicates that flow paths converge at the CFR along the western edge of the valley (figure 3). Mapping of alluvium thickness indicates that alluvium volume decreases as you move west through the study area, especially close to valley margins and underneath the CFR in this area. There are also bedrock knobs and an area of shallow alluvium to the east of the start of the survey in figures 2 and 4. These features could serve to channel flow paths toward the southern end of the study area, which is marked as an artesian zone in figure 3.
The artesian zone in figure 3 also coincides with the first groundwater discharge zone (figure 11). These factors provide strong evidence that groundwater from both the deep and shallow aquifers is discharging to this area. Overall valley geometry and alluvium depth mapping show that alluvium cross-sectional area decreases towards the northwest end of the valley. As alluvium volume decreases, excess groundwater within the alluvium must be discharged. As flow paths converge and the valley narrows, groundwater discharge will be focused in topographically low areas (Freeze and Witherspoon, 1967). The 2nd groundwater discharge zone precedes a chokepoint where alluvium volume decreases, and the potentiometric surface directs flow paths straight at the bedrock rise below the river. These factors cause groundwater discharge in this area.

Another possible source of groundwater discharge in this area is from the smaller alluvial valleys created by the incision of Deep Creek, Albert Creek, and Rock Creek. These valleys extend into the mountain block west of the study area. Incised alluvial valleys such as these tend to concentrate groundwater flow from the surrounding mountain blocks, as they are topographically low points with a higher hydraulic conductivity (Welch, 2012). While these creeks were not flowing during the study period, it is still possible groundwater from the shallow alluvial groundwater systems beneath these streams is discharging to the CFR.

Both discharge zones appear to be preceded by a smaller tributary entering the valley. It is possible that they could be contributing to their respective discharge zones. Since the first discharge zone begins before the start of sampling, it is uncertain whether the mouth of Deep Creek marks its beginning; however, the coincidence of the discharge zone with the location of artesian conditions in the aquifer indicate that the Missoula Valley Aquifer is likely the principal source of discharge. The 2nd discharge zone has its peak discharge at the mouth of Rock Creek; it is likely that groundwater from the Rock Creek alluvium is contributing to discharge to the CFR. If it were the sole source of groundwater discharge for this zone, we would expect peak $^{222}\text{Rn}$ concentrations right after the mouth of Rock Creek, followed by a gradual drop as
the CFR degassed. The peak $^{222}$Rn concentration is 2 km downstream from Rock Creek, and $^{222}$Rn is distributed broadly through the zone. The same is true for the first discharge zone and Deep Creek. Therefore, while these smaller alluvial valleys may be adding to groundwater discharge in this area, the primary contribution is from the Missoula Valley aquifer to the west.

**Conclusions**

In this study, we investigated the spatial distribution of instream $^{222}$Rn to locate and quantify groundwater discharge suggested by previous work. We used RADIN13 to model groundwater discharge quantities and locations from observed instream $^{222}$Rn. We performed a literature review on the Missoula Valley aquifer and nearby bedrock walls to interpret nearby geology to explain these distributions in the context of topography, subsurface hydraulic conductivity, and structural figuration.

We detected instream $^{222}$Rn concentrations as high as 1148 mBq/L, and a second peak as high as 877 mBq/L. *Cook et al* (2003) found that the Daly River in Northern Territory, Australia had $^{222}$Rn concentrations of up to 3000 mBq/L during their study period. Their modeling indicates groundwater is discharging to the Daly river, causing streamflow to increase by over 100% through their studied reach. While our instream $^{222}$Rn concentrations were not as high as those found in *Cook et al* (2003), our results clearly indicate there is significant groundwater discharge in the study area, affirming previous research that postulated this stretch of the CFR has areas of groundwater discharge.

Modeling results indicate that the study area was receiving at least $3.7\times10^5$ m$^3$/day, up to $8.0\times10^5$ m$^3$/day, broken into two zones. The first zone began before the start of sampling and extended for 2.5 km, reaching peak discharge near its end. The 2nd zone began 4.5 km after the start of sampling and extended for 2.5-3.5 km, with peak discharge at the beginning of the zone. These discharge zones are likely caused by 2 sources of groundwater. The primary source is groundwater discharge from the Missoula Valley aquifer due to a decrease in overall valley width along the CFR’s flow path, and a decrease in alluvium depth underneath the CFR.
Smaller alluvial valleys with mouths adjacent to the CFR could also contribute to groundwater discharge in these zones.

These findings indicate that the CFR and Missoula Valley aquifer are constantly interacting along its course through the valley. Water introduced to the aquifer earlier along the CFR flows through the subsurface as valley width increases, eventually finding its way back to the CFR when the valley begins to narrow. The aquifer serves as a large reservoir for the CFR, providing a temporally buffered base flow. Future work could involve better quantification of $k$ and $c$, in this area, to better constrain discharge quantities. A model that would allow for spatial variation in $k$ would also benefit future attempts to quantify discharge volumes on rivers with high spatial variability of $k$, such as the CFR and other rivers with heterogenous streambed morphologies. Additionally, the transience of discharge could be investigated by sampling at different times, to better understand how the CFR and Missoula Valley aquifer interact over time.
References


