Isotope hydrology of precipitation, surface and ground waters in the Okanagan Valley, British Columbia, Canada

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Summary

Rapid population growth in the intermontane, semi-arid, Okanagan Valley of Western Canada has led to water shortages with increasing debate over competing water interests. Nevertheless, the relationships between the various water resources of the Okanagan remain poorly defined. Stable isotopes of hydrogen (\(\delta^2H\)) and oxygen (\(\delta^{18}O\)) were used to assess rainfall importance and the sources and flux of water to and from rivers and lakes in watersheds, and to evaluate the origin of ground water resources. Precipitation isotopes resulted in a meteoric water line of \(\delta^2H = 6.6 (\delta^{18}O) - 22.7\) for the Okanagan Valley. Isotopic seasonality in precipitation was evident, with summer precipitation clearly affected by local recycling of water vapor. The \(\delta^2H\) and \(\delta^{18}O\) of surface waters were more positive than mean annual precipitation, indicative of basin-scale evaporation of surface waters; however, Okanagan Lake and its downstream river and lake system were isotopically synchronous, indicating that they behaved as a single well-mixed hydrologic unit. Isotopic mass-balance modeling revealed \(\sim 35\%\) of inflow to the Lake Okanagan watershed was lost to evaporation, validating a meteorological water balance model for the region. Highland bedrock was recharged with snow-melt and early spring rains, with the isotopic composition dependent on elevation. Ground waters in the Valley bottom aquifers west of Osoyoos were recharged by irrigation water obtained from the Okanagan River system, with no evidence of recharge connections from the highland bedrock.

1. Introduction

The Okanagan Valley is located in the Southern Interior Plateau of British Columbia, Canada, and northern Washington State, United States (Fig. 1). In Canada, the Okanagan extends north–south for about 160 km, and covers approximately 8200 km². Population growth in the Valley has been increasing at an unprecedented rate over the past few decades. The population of Kelowna, the region’s largest city (Fig. 1), grew from \(\sim 19,000\) in 1971 to \(\sim 106,000\) in 2006, with similar growth trends for smaller towns throughout the Okanagan Valley (Statistics Canada; www.statcan.gc.ca). Kelowna’s population is projected to continue to grow at an annual rate of 1.5%, resulting in a population increase to \(\sim 162,000\) by 2030 (City of Kelowna; http://www.kelowna.ca).

Population growth in the Okanagan is fueled by a warm, intermontane semi-arid climate, and by abundant recreational opportunities and tourism. While the area sustains several freshwater lakes, including Okanagan Lake (100 km long and 230 m deep; Fig. 1), there is a looming water shortage in the Valley. The limited surface water reserves are completely managed, with about 70% of available water allocated for irrigated agriculture. The rapidly increasing population therefore competes with agriculture, natural vegetation, fisheries, and wildlife for available surface water, and exerts new pressure on the available ground water resources (Cohen and Kulkarni, 2001; Cohen et al., 2006; Merritt and Alila, 2004; Merritt et al., 2006; Neilsen et al., 2001). The competing water interests of various stakeholders are long recognized (Canada–British Columbia Consultative Board, 1974), and continue to be a topic of debate in a region where fresh water availability per capita ranks among the lowest in Canada (Statistics Canada; www.statcan.gc.ca).

Since the hydrology of semi-arid areas is particularly sensitive to short and long-term changes in climatic processes, several studies have attempted to consider aspects of climate in the Okanagan (Frederick and Major, 1997). Short and long-term changes in climate may alter the magnitude and timings of peak and low flows, which can result in uncontrolled variability in the quality and quantity of water for the aquatic environment, and for storage to meet both urban and agricultural requirements (Merritt et al., 2006). Consequently, the Okanagan Valley has been subjected to numerous computer modeling efforts, including those attempting...
to integrate climate to predict future water availability for crops (Merritt et al., 2006), assess ground water resources (Liggett and

Wolfe et al., 2007). The seasonal and climatically driven covariance between precipitation, surface waters, and ground waters in the Okanagan Valley remain poorly defined, and hence new data aimed at understanding of the regional water cycle and the connections between precipitation, surface waters, and ground waters in the Valley are needed to inform, manage, and protect the regional water resources for the future.

In water cycle studies, the stable isotopes of hydrogen ($^2$H) and oxygen ($^{18}$O) in H$_2$O have long been used as naturally occurring tracers to assess rainfall importance and the sources and flux of water to and from rivers and lakes in watersheds, and to evaluate the origin and age of ground water resources (Bowen, 2010; Clark and Fritz, 1997; Ferguson et al., 2007; Gibson and Edwards, 2002; Wolfe et al., 2007). The seasonal and climatically driven covariance between the $^2$H and $^{18}$O content of water in precipitation results in a robust “local meteoric water line” (LMWL) relationship (Craig, 1961), which can be linked to ground and surface water resources (e.g., rivers, lakes, wetlands) to assess the relative importance of seasonal precipitation contributions to western regional surface water reserves (Henderson and Shuman, 2009, 2010). In addition, evaporation of surface water enriches the heavy oxygen and hydrogen content of the remaining water to produce systematic linear deviations from the LMWL (Craig and Gordon, 1965; Gonfiantini, 1986). This local evaporation line (LEL) can be exploited to provide basin-scale first order estimates of the relative magnitude of evaporation and water inflow to and from individual surface water bodies (Gat, 1996; Gibson and Edwards, 2002). Regional or watershed scale isotope-based water balance studies have been established for various regions of North America (Gibson et al., 2005a) and in semi-arid prairie watersheds (Ferguson et al., 2007). No comprehensive isotope assessments of precipitation, surface, and ground waters in the Okanagan Valley have been undertaken.

The objectives of this research were to use the stable isotopes of water to: (1) establish a foundational LMWL for the intermontane Okanagan Valley by measuring the temporal stable isotope composition of precipitation at stations in the Valley; (2) compare the isotopic composition of Okanagan surface waters to precipitation to estimate surface water loss by evaporation; and (3) evaluate the source of water in shallow and deep ground water systems in the southern Okanagan. Taken altogether, the stable isotope data obtained from all key hydrological components will establish foundational data for future water supply assessment and modeling studies in the Okanagan Valley.

2. Study location

The Okanagan Valley extends southward into the United States (Fig. 1), although the current study was conducted only in Canada (with the exception of one surface water sampling station located 10 km south of the international border). The intermontane Valley lies in the rain shadow of the Coast and Cascade mountains. In winter, moist air from the Pacific (Maritime Polar air) is forced over the Coastal and Cascade mountains by westerly winds, which cools and rains out on the windward (western) side of the mountains. These air masses are forced down the lee (eastern) side of the mountains into the Southern Interior Plateau, where it warms and dries. By the time Maritime Polar air reaches the Okanagan, much of its moisture has been lost (BC Ministry of Environment, February 2008b). Because surface waters in the Okanagan generally do not freeze over in winter, cold air pools within the Valley and thermal inversions trap local moisture, supporting low cloud formation (Klock and Mullock, 2001). Eastwards, the adjacent mountain range provides an effective barrier against the westward movement of Continental arctic air masses from the Prairies. In winter, a Pacific high pressure moves northward and diminishes the frequency and intensity of Pacific derived precipitation. By June, heat and surges of cool unstable air produce convective storm activity, resulting in rain and thundershowers (BC Ministry of Environment, February 2008b). Generally, weather systems from the west provide the majority of precipitation to the Okanagan in wintertime and spring, while localized convective systems provide most of the precipitation in summer (Klock and Mullock, 2001; Merritt and Allia, 2004).

The climate of the Okanagan Valley is classified as semi-arid. Overall, precipitation amounts in the basin are low (~647 mm/year), with rainfall decreasing from north to south. The mean annual precipitation is 447 mm in the city of Vernon, 320 mm in Kelowna, and 320 mm in Osoyoos. Most precipitation in the Okanagan occurs in the late winter to late spring period. Temperatures are approximately 2–3° cooler in the north of the Valley than in the south, and evapotranspiration (ET) is greater in the south than the north. Mean annual water deficit estimates in the Valley, based on the difference between total annual precipitation and total
Evapotranspiration (ET), range from about 300 mm/year in Vernon to 500 mm/year in Osoyoos with an overall average of 524 mm/year (Okanagan Basin Water Board – http://www.obwb.ca).

The Valley contains six major lakes (Table 1): the Wood, Kalamalka, Okanagan, Skaha, Vaseaux and Osoyoos Lakes (Fig. 1). The Okanagan, Skaha, Vaseaux and Osoyoos Lakes drain southward. Okanagan Lake discharges into the Okanagan River, which flows through Skaha, Vaseaux and Osoyoos Lakes before discharging into the Columbia River in Washington State (Merritt and Alila, 2004). Below Okanagan Lake, the Okanagan River is modified by numerous irrigation and flood control dams, channelization, river flow containment dykes and vertical drop structures (Osoyoos Lake Water Quality Society; http://www.olwqs.org/). Discharge and water levels are recorded at several stations along the Okanagan River. High discharge rates in springtime typically correspond to the snowmelt period or occasional heavy rains, while discharge spikes in the fall correspond to artificial releases from Okanagan Lake to meet the irrigation requirements of the south Okanagan. Osoyoos Lake outflow is regulated by the Zosel Dam, located 4 km south of Osoyoos in the United States.

Shallow, easily obtainable ground water resources in the southern Osoyoos region of the Valley are typically found in the valley-bottom benchland aquifers (Nasmith, 1962). These benchland aquifers are comprised of sand and gravel outwash terraces and silts ranging from 1 to 16 m in thickness (Nasmith, 1962; Piteau Associates, 1989). A surficial unconfined aquifer comprises the Valley benchlands west of Osoyoos Lake and is central to the ground water part of this study (Fig. 2). This aquifer is classified as a moderate demand and productivity unit, but highly vulnerable to contamination from surface sources (BC MOE, September 2008a). Based on the terraced topography and the undulating nature of the benchlands, this aquifer is likely discontinuous. Deeper confined ground water is present in coarse deposits underlying the glaciolacustrine and till deposits. Mountainous bedrock (highland) occur on either side of the benchlands to elevations of 900 m above sea level (masl) on Kiplooa Mountain and 1900 masl on Mt. Baldy (Fig. 2). These highland bedrock units are water bearing, and are comprised of fractured granitic intrusive rocks and metamorphic rocks. In the Valley bottom, granitic bedrock is encountered at depths of 51–123 m below ground surface (BG).

### Materials and Methods

#### 3.1. Precipitation Samples

Precipitation samples were collected from April 2006 to June 2010 using passive collectors (with paraffin oil added to prevent evaporation) at three collection stations along the Okanagan Valley at West Kelowna, Oliver, and Osoyoos West (Fig. 1 triangles). The stations at Osoyoos and Kelowna collected monthly integrated precipitation samples between April 2006–2008. Due to low and seasonally variable precipitation yields, additional event-based precipitation sampling was undertaken at Oliver between January 2008 and June 2010. Amount-weighted mean annual precipitation

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**Table 1**

<table>
<thead>
<tr>
<th>Lake</th>
<th>Length (km)</th>
<th>Surface area (km²)</th>
<th>Volume (km³)</th>
<th>Mean depth (m)</th>
<th>Maximum depth (m)</th>
<th>Water residence time (years)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Okanagan</td>
<td>113</td>
<td>344</td>
<td>24.6</td>
<td>76</td>
<td>242</td>
<td>53–60</td>
</tr>
<tr>
<td>Skaha</td>
<td>12</td>
<td>20</td>
<td>0.6</td>
<td>26</td>
<td>57</td>
<td>1.2</td>
</tr>
<tr>
<td>Vaseaux</td>
<td>4</td>
<td>2.8</td>
<td>0.02</td>
<td>6.5</td>
<td>27</td>
<td>0.03</td>
</tr>
<tr>
<td>Osoyoos</td>
<td>16</td>
<td>23</td>
<td>0.4</td>
<td>14</td>
<td>63</td>
<td>1.7</td>
</tr>
</tbody>
</table>

Fig. 2. Location of numbered surface and ground water sampling stations in the South Okanagan region of Osoyoos. Solid line denotes the Okanagan watershed boundary. Site details and data are reported in Supplemental materials.
Table 2
Surface water sampling stations in the Okanagan Valley, British Columbia, Canada. Station numbers are plotted in Figs. 1 and 2.

<table>
<thead>
<tr>
<th>Station ID</th>
<th>Location</th>
<th>Elevation (masl)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Okanagan river system</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Ellison</td>
<td>2</td>
<td>50.24°N</td>
</tr>
<tr>
<td>Peachland</td>
<td>3</td>
<td>49.78°N</td>
</tr>
<tr>
<td>Okanagan River</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Penticton</td>
<td>6</td>
<td>49.50°N</td>
</tr>
<tr>
<td>OK Falls</td>
<td>7</td>
<td>49.34°N</td>
</tr>
<tr>
<td>Oliver</td>
<td>13</td>
<td>49.11°N</td>
</tr>
<tr>
<td>Osoyoos, United States</td>
<td>109</td>
<td>48.94°N</td>
</tr>
<tr>
<td>Osoyoos Lake</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Mouth of River</td>
<td>15</td>
<td>49.08°N</td>
</tr>
<tr>
<td>Pumphouse</td>
<td>33</td>
<td>49.06°N</td>
</tr>
<tr>
<td>Central Basin</td>
<td>83</td>
<td>49.02°N</td>
</tr>
<tr>
<td>Canada–United States border</td>
<td>107</td>
<td>49.00°N</td>
</tr>
<tr>
<td>Other surface water bodies</td>
<td></td>
<td></td>
</tr>
<tr>
<td>McKinney Creek</td>
<td>14</td>
<td>49.12°N</td>
</tr>
<tr>
<td>Spotted Lake</td>
<td>21</td>
<td>49.08°N</td>
</tr>
<tr>
<td>Kettle Lake</td>
<td>49</td>
<td>49.05°N</td>
</tr>
<tr>
<td>Wildhorse</td>
<td>51</td>
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</tr>
<tr>
<td>Blue Lake</td>
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</tr>
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<td>Peanut Pond</td>
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<td>49.03°N</td>
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<td>Kilpoola Lake</td>
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</tr>
<tr>
<td>Taylor</td>
<td>69</td>
<td>49.03°N</td>
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<tr>
<td>Chapman</td>
<td>75</td>
<td>49.03°N</td>
</tr>
<tr>
<td>Haynes Creek</td>
<td>99</td>
<td>49.01°N</td>
</tr>
<tr>
<td>Yos Pond</td>
<td>105</td>
<td>49.01°N</td>
</tr>
</tbody>
</table>

(MAP) $\delta^2$H and $\delta^{18}$O values were obtained for monthly integrated stations using the precipitation amounts recorded at adjacent Environment Canada weather stations (http://www.climate.weatheroffice.gc.ca/climateData), and for the event based samples at Oliver by measuring the water collected in the precipitation events.

3.2. Surface water samples

Surface water samples were collected on a monthly basis where possible from May 2006 to June 2010 at 9 stations along the Okanagan River system (Fig. 1, solid circles); with additional smaller surface water (e.g. ponds, streams) samples collected opportunistically in the highlands on one occasion (Fig. 2). A numbered list of all surface-water stations is provided in Table 2. On the Okanagan River system (Fig. 1), two stations were located on Okanagan Lake at Ellison (2) and Peachland (3). Four stations were located on the Okanagan River at Penticton (6), Okanagan Falls (7), Oliver (13), and Oroville (109). Three stations were located on Osoyoos Lake – one at the mouth of the Okanagan River (15), in-lake (33) and near the Canada–United States border (107). Major surface water time series locations are shown in Fig. 1. The ten highland surface water stations in the Osoyoos area are shown in Fig. 2 (14, 21, 49, 51, 55, 63, 67, 69, 75, and 99).

3.3. Ground water samples

The ground water stations sampled in the south Okanagan are shown in Fig. 2. Thirty-two monitoring wells in the Valley bottom and benchlands in the Osoyoos area were sampled. Twenty of these (26, 28, 30, 35, 39, 42, 43, 44, 54a, 54b, 61, 62a, 62b, 66a, 66b, 70, 72a, 72b, 74a and 74b) were provincial observation wells. Eight (24, 25, 32, 37, 38, 46, 52 and 57) were domestic or nested wells (lettered subscripts denote nested wells in Table S3). Four (27, 29, 31, and 34) were research wells installed by the University of Saskatchewan in areas where subsurface irrigation drainage systems existed. In the highlands, 11 wells were sampled on Kilpoola Mountain (16, 17, 18, 19, 20, 22, 23, 40, 41, 48 and 50), and 38 wells on Anarchist Mountain (58–60, 64, 65, 68, 73, 76–80, 84–87, 89–98, 100–104, 106 and 108). To the east of the Okanagan watershed boundary, several wells were sampled at Bridesville (45, 53, 56 and 71) and on Mt. Baldy (11). Two springs east of Anarchist Mountain (81 and 82) and a spring on Mt. Baldy (12) were sampled. Two municipal supply wells (9 and 10) in the Town of Oliver were sampled. Access to Town of Osoyoos pump houses (47 and 88), as well as an irrigation tap (36) from a landowner in the northern portion of the study area serviced by (88), were used for the collection of irrigation water samples. All well, pump house, and spring locations and the location of the irrigation tap are shown in Fig. 2. Of the 87 ground water wells sampled in the study area, 25 were screened in the surficial sand and gravel aquifer, 10 in the top of the glaciolacustrine and/or till unit, one in the confined aquifer, one in the upper benchland, and 48 were in the highlands to the east and west of the Valley. One well (38) was from a water-bearing geologic unit present at the contact of the upper benchland and the highland bedrock. The Town of Oliver supply wells were completed in the surficial sand and gravel aquifer.

3.4. Water sampling methods

Unfiltered precipitation, surface water and ground water samples were collected by passive sampler, hand or mechanical pump, bucket or bailer, where appropriate. Osoyoos Lake was sampled using a Van Dorn sampler. All water samples were collected in 30 ml HDPE bottles and sealed. River and lake water samples were collected at a depth of about 0.3 m below the surface.

3.5. Stable isotopic analyses

Water samples were analyzed for $\delta^2$H and $\delta^{18}$O at Environment Canada, Saskatoon, Canada. Isotopic analyses were conducted by laser spectroscopy using a Los Gatos Research DLT-100 liquid isotope water analyzer (Los Gatos Inc., California) using the methods described in (Lis et al., 2008). Laboratory standards, previously calibrated to the VSMOW2-SLAP2 reference waters were INV1 ($\delta^2$H = −214‰, $\delta^{18}$O = −28.2‰) and ROD2 ($\delta^2$H = −4.9‰, $\delta^{18}$O = −1.3‰). INV1 and ROD3 were used to normalize the results to the VSMOW-SLAP scale by assigning $\delta^{18}$O and $\delta^2$H values of −55.5‰ and −428‰, respectively, to the SLAP reference water. The laboratory precision was ±1.0‰ for $\delta^2$H and ±0.2‰ for $\delta^{18}$O (Lis et al., 2008).
4. Results and discussion

4.1. Isotopes of precipitation

The δ2H and δ18O values of 106 precipitation monthly and event samples collected in the Okanagan Valley between April 2006 and June 2010 are presented in Fig. 3. The precipitation isotope data are tabulated in the Supplemental materials (Table S1). A Kruskall–Wallis non-parametric test revealed there was no difference (p = 0.65) in the hydrogen and oxygen isotopic data collected between Oliver (n = 65) and Osoyoos and Kelowna (n = 45), at the south and north ends of the Valley. Hence, all event and monthly precipitation in the Okanagan Valley were pooled as a single data-set that was representative of the entire Okanagan Valley. The δ2H and δ18O values of precipitation in the Valley ranged from −177‰ to −42‰ and −23.0‰ to −3.0‰, respectively. A least squares regression of the precipitation isotope data resulted in highly significant Okanagan Valley meteoric water line (OMWL): δD = 6.6 δ18O − 22.7 (n = 106, r² = 0.93, Fig. 3). The amount-weighted mean δ2H and δ18O values of precipitation (MAP) over the sampling period in the Okanagan Valley were −102‰ and −12.5‰, respectively. This amount-weighted MAP was more negative than Victoria, BC, to the west (−77‰ and −10.2‰, respectively) (Fritz et al., 1987), and more positive than Calgary, Alberta, to the east (−136‰ and −17.9‰, respectively (Peng et al., 2004). The progressive evolution towards more negative δ2H and δ18O MAP values inland and eastwards from the Pacific coast is controlled, in part, by well-known Rayleigh rain-out processes as vapor masses from the Pacific preferentially lose 2H and 18O as they move over the Coast Range (Yonge et al., 1989). The monthly precipitation isotope samples revealed the expected temperature-dependent seasonality, with more negative δ2H and δ18O values occurring in the winter (November–April) and more positive values occurring in the summer and early fall (May–October).

Winter precipitation δ2H and δ18O in the Okanagan Valley had lower isotopic variance and plotted closer to the Canadian Meteoric Water Line (CMWL) and to the Victoria-Calgary meteoric water lines than the late spring and summer rains (Fig. 3). The OMWL exhibited a slightly lower slope and a considerably lower intercept than the CMWL and the MWLs of Victoria and Calgary. The lower slope and intercept could be attributed to the convective recycling effects that drive late spring and summer rains. Water vapor evaporated from surface water bodies can mix with air masses moving over these areas, and the mixing of 2H and 18O enriched recycled water vapor results in a lower slope (Peng et al., 2004). Ingraham and Taylor (1986) showed that rain downwind from where air masses receive recycled water vapor are more enriched in 2H and 18O than if rainout occurred without any addition of recycled water vapor. In semi-arid regions, summer rains from thunderstorms that obtain moisture mainly from local evapotranspiration are also known to produce low slopes (Peng et al., 2004). This is due to the aforementioned mixing, and to secondary evaporation as raindrops beneath the cloud-base evaporate as they fall to the ground surface and undergo subsequent isotopic fractionation (Dansgaard, 1964). This recycling effect is generally seen in the slopes of summer meteoric water lines, because secondary evaporation of snowfall in the winter is usually negligible (Fritz et al., 1987). In the Okanagan, the summer months are characterized by high evapotranspiration rates and temperatures and low relative humidity. These conditions drive evaporation in the Okanagan watershed, and this evaporated moisture plays a key role in driving summer precipitation events (i.e., thunderstorms). The low slope of the OMWL, particularly in summer thus revealed that summer precipitation in the Okanagan contains recycled water as a result of incorporation of moisture from evaporation of local surface water bodies and irrigated soils.

The deuterium excess (δ-excess) value of precipitation in the Okanagan Valley ranged between −40.5‰ and +10.8‰, with an amount weighted mean value of −5.2 ± 9.4‰, compared to the global weighted mean δ-excess of +10‰ (IAEA, 2006). The δ-excess values were lowest (and more variable) in the summer (early fall and highest (and less variable) in the winter, as shown in Fig. 4a. The δ-excess values for precipitation in the Okanagan Valley, particularly during the summer, further supported the secondary processes noted above – the incorporation of local recycled water vapor into the air mass and secondary evaporation of rainfall during small rain events. The variable δ-excess values during winter (generally −11‰ to +11‰) may reflect higher humidity during formation of Pacific vapor masses (low δ-excess) and non-equilibrium conditions that occurred during the formation of snow (high δ-excess) (Merlivat and Jouzel, 1979).

The δ18O values of monthly integrated precipitation collected from the Okanagan Valley and the average ground air temperature (T) for the sampling interval were plotted on a δ18O versus temperature diagram (Fig. 4b). The significant relationship between δ18O values and average air temperature was δ18O = 0.337 − 16.6 (r² = 0.53, n = 38). The observed isotopic depletion rate for δ18O per degree Celsius drop in air temperature for the Okanagan was between −0.3‰ and −0.4‰.

4.2. Isotopes of surface waters

The δ2H and δ18O values of the Okanagan River and lake system samples were plotted on a δ2H–δ18O diagram, and compared to the OMWL and MAP (Fig. 5). All surface water isotopic data are tabulated in the Supplemental materials (Table S2). The δ2H and δ18O values of the Okanagan River system ranged from −11.1‰ to −9.8‰ (mean = −10.3 ± 3‰) and −13.5‰ to −9.9‰ (mean = −11.5 ± 0.6‰), respectively. The δ2H and δ18O values plotted in a tight cluster to the right of MAP, but within the observed 95% confidence interval (CI) for Okanagan Valley precipitation. Linear regressions of temporal δ2H versus δ18O for each riverine and lake station revealed weak (r² < 0.2) to no isotopic trends (not shown), with a very weak trend overall for all Okanagan Lake and River surface waters combined (r² = 0.12, n = 140). Despite the large seasonal isotopic variations observed in precipitation in the Valley, the seasonal variation in δ2H and δ18O values of water in the Okanagan River system was very small, with a range of only 13‰ and 3.8‰ for δ2H and δ18O, respectively, over the timeframe of the study. The fact that synchronous isotopic patterns were
observed in Okanagan Lake and its downstream river and lake stations suggested that little evaporation of water occurred during the short travel time in the Okanagan River system following discharge from Okanagan Lake. Similarly, an analysis of covariance (ANCOVA) of the stable isotopes by date and sampling location remained unknown how these water bodies might change over time.

Additionally, the \( \delta^2H \) and \( \delta^{18}O \) values of several sampled upland ponds, streams and small closed basin lakes in the southern Okanagan region were plotted on Fig. 5. The \( \delta^2H \) and \( \delta^{18}O \) values of these smaller surface water bodies ranged widely from \(-81\%e\) to \(-85\%e\) (mean \(-92\%e\)) and \(-10.7\%e\) to \(-4.7\%e\) (mean \(-8.5\%e\)), respectively. The \( \delta^2H \) and \( \delta^{18}O \) values of surface waters on Kilpoola Mountain (67, 55, 21, 51) ranged from \(-66\%e\) to \(-49\%e\) (mean \(-56\%e\)) and \(-3.4\%e\) to \(-1.8\%e\) (mean \(-2.4\%e\)), respectively. The positive \( \delta^2H \) and \( \delta^{18}O \) values of samples to the upper right and outside of the observed 95% CI of the OMWL clearly indicated that evaporation of some closed basin surface water bodies had occurred. Conversely, the \( \delta^2H \) and \( \delta^{18}O \) values of Taylor (69), Chapman (75) and Haynes Creeks (99) were more negative and ranged from \(-125\%e\) to \(-120\%e\) (mean \(-123\%e\)) and \(-16.3\%e\) to \(-15.4\%e\) (mean \(-15.9\%e\)), respectively. The \( \delta^2H \) and \( \delta^{18}O \) values of Yos Pond (105) were \(-116\%e\) and \(-12.2\%e\), respectively. Those of McKinney Creek (14) were \(-130\%e\) and \(-17.3\%e\), respectively. These negative \( \delta^2H \) and \( \delta^{18}O \) values of surface waters on Anarchist Mountain and Mt. Baldy were indicative of snowmelt, spring precipitation, or highland ground water sources with no to limited subsequent evaporation. It should be noted that no comprehensive time series were available on these smaller upland surface water bodies, thus it remained unknown how these water bodies might change over the course of time.

The \( \delta^2H \) values of two surface water stations in Okanagan Lake, four Okanagan River stations and four stations in Osoyoos Lake were plotted versus sampling date and volumetric discharge in Fig. 6. Although this time series was just 2 years, several observations could be made. The lower \( \delta^2H \) and \( \delta^{18}O \) values for all lake and river stations tended to occur in the months of May-June following snowmelt, often with increased discharge from Okanagan Lake, whereas the higher \( \delta^2H \) and \( \delta^{18}O \) values occurred in September and October. The riverine sampling stations downstream of
Okanagan Lake closely matched the isotopic composition of Okanagan Lake discharge at all times. The Okanagan Lake surface water samples were only collected in 2006, and exhibited considerable spatial isotopic spread, unsurprising given the short water residence time of <2 years (Table 1) may preserve some seasonal changes in Okanagan River inputs. The Okanagan River water samples did not exhibit any systematic isotopic enrichment in a downstream direction (see above), even though increasing temperatures, lower amounts of precipitation and higher rates of evapotranspiration at the south end of the Valley contribute to a change in biogeoclimatic zone from Ponderosa Pine to Bunchgrass near Skaha Lake (BC MOE, September 2008a). This finding was also not unexpected since Okanagan Lake and Okanagan River are located in the same Valley bottom within an elevational range of 65 m and over a spatial distance of only 140 km (e.g. short riverine residence time). Variations in \( \delta^2H \) and \( \delta^{18}O \) values among riverine sampling stations during each monthly sampling event were very small. The near negligible spatial variations in \( \delta^2H \) and \( \delta^{18}O \) values of surface water samples associated with the Okanagan River system, along with the close isotopic correspondence with Okanagan Lake, revealed that the Okanagan Lake and its downstream river and lake system essentially behave as a single hydrologic unit, rather than as a series of interconnected and independently evaporating lakes and rivers, and verified that the overwhelming contributor of water into downstream main-stem water bodies was Okanagan Lake. Given the comparatively long residence time of water in Okanagan Lake (>50 years, Table 1), low annual isotopic variation (<10 \( \delta^2H \)), in its discharge compared to a wide annual isotopic range observed in precipitation (~130\% for \( \delta^2H \)) was not unexpected.

4.3. Isotopes of ground water

All \( \delta^2H \) and \( \delta^{18}O \) values of ground water samples from the southern Okanagan part of the study area are plotted in Fig. 7 on a \( \delta^2H \) versus \( \delta^{18}O \) diagram against the OMLW. All ground water isotopic data are tabulated in the Supplemental materials (Table S3). The data can be classified into two distinct isotopic groupings: the highland bedrock and the Valley bottom benchland aquifers.

![Fig. 7. Plot of \( \delta^2H \) versus \( \delta^{18}O \) of Valley benchland and highland aquifers in the Osoyoos region of the Okanagan watershed, compared to the OMLW. Circles, triangles and diamonds denote the three groupings of highland ground waters. Inverted triangles denote ground water from benchland aquifers. The gray shaded area denotes the range of Okanagan Lake and River system surface water. Specific Valley ground water outliers groups (38, 62, 50) are discussed in the text. Dashed lines encompass the observed 95\% CI of precipitation.](image-url)

4.3.1. Highland bedrock ground water

The \( \delta^2H \) and \( \delta^{18}O \) values of highland bedrock (Fig. 7) on Kilpoola Mountain ranged from −149.6\% to −123\% (mean = −129 ± 8\%e) and −19.6\% to −16.1\% (mean = −16.9 ± 1.1\%e), respectively, with the exception of well 50, where \( \delta^2H \) and \( \delta^{18}O \) values were −67\% and −3.5\%, respectively. The \( \delta^2H \) and \( \delta^{18}O \) values of all highland ground water plotted to the lower left in the OMLW diagram (Fig. 7), within the zone for snowmelt and spring precipitation. The highland bedrock ground water samples were binned into three isotopic sub-groups, primarily related to elevation (see below). The first group consisted of seven highest elevation wells (17 and 19 on Kilpoola; 79, 86, 94 and 103 on Anarchist Mountain; and 56 at Bridesville) where ground water exhibited the most negative \( \delta^2H \) and \( \delta^{18}O \) values. These \( \delta^2H \) and \( \delta^{18}O \) values ranged from −149.6\% to −139\% (mean = −144 ± 3\%e) and −19.6\% to −18.0\% (mean = −18.6 ± 0.5\%e), respectively. The second mid-elevation grouping consisted of 31 samples from 29 wells (16 and 22 on Kilpoola; 58, 60, 65, 68, 77, 78, 84, 85, 87, 90, 91, 92, 93 [2], 95, 96 [3], 97, 98, 100a, 100d, 102, 104, 106 and 108 on Anarchist Mountain; 45, 53 and 71 at Bridesville; and 11 on Mt. Baldy, and two springs (81 and 82 on Anarchist Mountain). Ground water in this group had \( \delta^2H \) and \( \delta^{18}O \) values that were more positive than the first group, and ranged from −138\% to −130\% (mean = −134 ± 2\%e) and −17.8\% to −15.5\% (mean = −16.8 ± 0.5\%e), respectively. The third group consisted of 21 mainly lower elevation (above the benchlands) samples from 17 wells (18, 20, 22 [1 and 2], 23, 40, 41 and 48 on Kilpoola; and 59, 64, 73, 76, 80, 89 [1 and 2], 93 [1], 96 [1 and 2], 100b and 100c on Anarchist Mountain, and one spring (22 on Mt. Baldy). Ground water in the third group had more positive \( \delta^2H \) and \( \delta^{18}O \) values than the other highland ground water that ranged from −129\% to −123\% (mean = −126 ± 2\%e) and −16.7\% to −15.6\% (mean = −16.2 ± 0.3\%e), respectively. Ground water exhibiting more negative \( \delta^2H \) and \( \delta^{18}O \) values may be indicative of: (1) higher elevation recharge via snowmelt and precipitation, or (2) wells that intersected deep fractures recharged with old ground water from colder climatic conditions with little mixing of older ground water with young ground water. Highland ground water with more positive \( \delta^2H \) and \( \delta^{18}O \) values could be a result of wells that intersected fractures recharged with younger ground water. Surface waters on Kilpoola did not appear to be a predominant source of recharge for bedrock aquifers (albeit based on limited surface water data above), as the \( \delta^2H \) and \( \delta^{18}O \) values of surface waters in the Kilpoola area ranged from −67\% to −49\% (mean = −59 ± 8\%e) and −3.5\% to −1.8\% (mean = −2.6 ± 0.8\%e), respectively. The only exception was the ground water from well 50, where \( \delta^2H \) and \( \delta^{18}O \) values were similar to those of nearby surface water sample 51 (~66\% and −3.4\%, respectively). Surface water bodies on Anarchist Mountain did not appear to be a major source of recharge for bedrock in the highlands east of Osoyoos. The \( \delta^2H \) and \( \delta^{18}O \) values of surface waters at Haynes Creek, Chapman and Taylor ranged from −125\% to −120\% (mean = −123 ± 3\%e) and −16.3\% to −15.4\% (mean = −15.9 ± 0.4\%e), respectively. Overall, ground water at higher elevations on Kilpoola Mountain, Anarchist Mountain, Bridesville and Mt. Baldy were more depleted in \( ^2H \) and \( ^{18}O \) than samples from lower elevations. This could be attributed to the altitude effect, where at higher elevations average temperatures were cooler, resulting in precipitation that was comparatively isotopically depleted in \( ^2H \) and \( ^{18}O \) (Clark and Fritz, 1997). The altitude effect was confirmed by \( \delta^2H \) values obtained from the 60 highland wells plotted against the surface elevation of each well (Fig. 8, excluding outliers 17, 11, 12). Regression analysis revealed that ~54\% of the \( \delta^2H \) isotopic variation in highland ground water could be explained by elevation alone, with an observed isotopic depletion rate of ~20\% in \( \delta^2H \) per 1000 m rise in elevation. Combining \( \delta^{18}O \) and elevation into a single analysis of
covariance (ANCOVA), over 99% \( (p < 0.001) \) of the variance in \( \delta D \) could be explained by the interaction with \( \delta ^{18}O \) and elevation. The isotopic patterns in highland bedrock aquifers were therefore a result of the combined effect of elevation, the prevalence of snowmelt or precipitation derived recharge, and possibly the mixing of old or younger ground water. Because of the lack of information here on ground water age and flow paths for each ground water station, it was not possible to further delineate alternate ground water sources without further detailed study. The observed significant altitude effect over a short spatial distance also suggested that the application of water mass balance studies in watersheds that have extreme topography may require the generation of low and high altitude MWLs or MAPs for watershed isotopic water balance modeling purposes (see below).

4.3.2. Valley aquifers

The \( \delta ^{2}H \) and \( \delta ^{18}O \) of the Valley bottom benchland aquifers in the Osoyoos region ranged from \(-124\% \) to \(-87\% \) (mean = \(-103 \pm 5\%) \) and \(-16.1 \) to \(-8.6\% \) (mean = \(-11.5 \pm 1.0\%) \), respectively. The \( \delta ^{2}H \) and \( \delta ^{18}O \) values plotted just above the midpoint of the OMWL and MAP, between spring and summer precipitation, and slightly to the right albeit within the 95% CI of the OMWL (Fig. 7). The majority of the \( \delta ^{2}H \) and \( \delta ^{18}O \) values of the benchland aquifer ground water were identical in isotopic composition to that of Okanagan Lake, as denoted by the gray ellipse in Fig. 7. The \( \delta ^{2}H \) and \( \delta ^{18}O \) values of ground water in the benchland aquifers of the Osoyoos area were significantly more positive than those of the highland aquifers in the surrounding region, unequivocally revealing that the benchland aquifers are not (currently) being recharged by groundwater or surface water from the Kilpoola highlands (Fig. 7), nor from up-gradient sources recharged by mean annual local precipitation. The only exception was the sample obtained from well 38, which plotted in the same range as the Group 3 lower elevation highland aquifers. Station 38 was a test well located in the uppermost part of the Valley benchlands at the boundary with the bedrock, and was above the lower Valley irrigation zones. The isotope data from station 38 was closer to that of snowmelt and spring precipitation and to highland ground water. The other exceptions were six samples that plotted above and to the right of the benchland cluster (Fig. 7). These were from nested multilevel wells (62 series) that were recharged by surface water from Peanut Pond, whose waters were isotopically enriched by evaporation (Fig. 7).

The \( \delta ^{2}H \) and \( \delta ^{18}O \) values of irrigation water used in the Osoyoos area ranged from \(-108\% \) to \(-101\% \) (mean = \(-105 \pm 3\%) \) and \(-12.8\% \) to \(-10.6\% \) (mean = \(-11.7 \pm 0.7\%) \), respectively. The Okanagan River system feeds Osoyoos Lake, which is the sole source of irrigation water for the Osoyoos study area, and the \( \delta ^{2}H \) and \( \delta ^{18}O \) values of irrigation water taps were similar to surface waters from the Okanagan River system. The insignificant differences in mean \( \delta ^{2}H \) and \( \delta ^{18}O \) values of shallow ground water (i.e., \(-103\% \) and \(-11.5\% \)) and lake irrigation water (i.e., \(-105\% \) and \(-11.7\% \)) may be indicative of very slight isotopic enrichment occurring between the application of irrigation water to the ground surface and its subsequent infiltration to the ground water table, or from partial evaporation of irrigation water by vapor diffusion during its residence in the top few meters of the unsaturated soil zone (Coplen et al., 2000).

Using the linear relationship between the isotopic composition of highland ground water and sample elevation (Fig. 6), and using an average ground surface elevation of 300 masl for the Osoyoos Valley benchland aquifers, the \( \delta ^{2}H \) of shallow ground water in the Valley bottom based on this relationship should be about \(-120\% \). This modeled value was significantly more negative than what was actually observed in the shallow ground water, further supporting that shallow ground water in the Valley bottom and benchlands is not currently naturally recharged by highland ground water or by mean precipitation occurring at the elevation of the Valley bottom and benchlands. The \( \delta ^{2}H \) and \( \delta ^{18}O \) values of ground water in the confined aquifer below the surficial benchland aquifers ranged from \(-108\% \) to \(-106\% \) (mean = \(-107 \pm 1\%) \) and \(-12.8\% \) to \(-11.1\% \) (mean = \(-11.7 \pm 1.0\%) \), respectively. Deeper confined ground water exhibited \( \delta ^{2}H \) and \( \delta ^{18}O \) values that were slightly more positive than the highlands and upper benchland, but similar to ground water in the Valley bottom and benchlands that were predominantly recharged by irrigation water.

4.4. Okanagan Valley isotopic water balance

According to the Okanagan Basin Water Board (http://www.obwb.ca), insufficient local data exists to directly quantify evaporation from the Okanagan Basin. In the Okanagan Valley it is currently estimated that about 12% of the precipitation falling within the watershed reaches Okanagan Lake, with roughly 85% lost to the combined processes of evapotranspiration from plant uptake and to evaporation, and the remainder discharging from the Valley through the river system (Nordin, 2005). Preliminary basin-scale quantitative evaporation estimates are made on the basis of several meteorological models, with evaporative loss estimates ranging widely from 350 mm to 1000 mm/year. The most parsimonious model produced evaporation estimates on the order of 25 mm/year from combined surface runoff of 78–82 mm/year, or approximately 30.5–32.1% of water lost by evaporation (http://www.obwb.ca).

While isotopic enrichment of water occurs during evaporation, evapotranspiration by plants does not result in any isotopic partitioning of \( ^{2}H \) and \( ^{18}O \) (Clark and Fritz, 1997). As a result, the \( \delta ^{2}H \) and \( \delta ^{18}O \) values of surface water at any given location in a watershed should reflect the isotopic composition of water derived from precipitation and any subsequent effects of evaporation as it travels through a watershed (Ferguson et al., 2007). Thus, the \( \delta ^{2}H \) and \( \delta ^{18}O \) values of surface water in the Okanagan system should reflect net evaporation of water within contributing creeks, rivers and lakes prior to watershed discharge. Further, only minor evaporation occurs during the short transit time from Okanagan Lake through the Okanagan River system, as evident by the lack of anisotopic evaporation trend from the OMWL (see above).

In semi-arid regions like the Okanagan, the isotopic composition of surface water can be modified by non-equilibrium evaporation, whereby water molecules containing the heavier isotopes \( ^{2}H^{2}O \) and \( ^{2}H^{18}O \) diffuse at a lower rate than water molecules containing the lighter isotopes \( ^{1}H^{2}O \) (Gat, 1981;
Surface waters that have undergone evaporation become more enriched in $^2$H and $^{18}$O than the resulting water vapor (Froehlich et al., 2005). The degree of progressive isotopic enrichment of the evaporating surface waters is dependent on meteorological factors such as relative humidity, water temperature and salinity (Gonfiantini, 1986; Karim et al., 2011; Rozanski et al., 2001). The isotopic composition of surface waters subjected to evaporation evolve along a line at a slope lower than that of the global and/or local meteoric water line (LMWL) (i.e., <8), that originates at the initial isotopic composition of the water prior to evaporation (Karim et al., 2011; Rozanski et al., 2001). The low slope trend exhibited by increasing $^2$H and $^{18}$O content is referred to as the local evaporation line (LEL) (Gibson et al., 2005b). Evaporating surface water bodies are therefore isotopically characterized by (1) a displacement of the isotopic composition of the residual non-evaporated water below the LMWL in the $\delta^2$H-$\delta^{18}$O diagram along a LEL; and (2) a limiting isotopic enrichment of the heavy isotopic water molecules (Horita et al., 2008). In conditions where relative humidity is less than 100%, water-vapor isotopic back exchange is minimized as water vapor leaving the system is no longer in equilibrium with the water (Karim et al., 2011). Gat (2008), summarizing Craig (1957), indicated that the non-equilibrium fractionation factor for $^{18}$O at normal earth surface temperatures is greater than the equilibrium fractionation factor by almost a factor of 2.

The Craig and Gordon model (1965) was developed for evaluating the isotopic composition of the evaporative flux of water into an open (under-saturated) atmosphere. It describes the evaporative flux relative to three regions above the water surface: (1) an interface (boundary) layer that is 100% water-saturated and in isotopic equilibrium with the underlying water column; (2) a laminar layer where transport of water vapor is dominated by molecular diffusion and non-equilibrium enrichment takes place when humidity is less than one; and (3) a turbulent section present at the base of the open atmosphere and where eddy diffusion and turbulent transport occur (Horita et al., 2008; Karim et al., 2011). Isotopic fractionation occurs in the interface and laminar layers, however, it does not occur in the turbulent section thus giving an isotopic profile that is a mixing profile (Karim et al., 2011).

The equation for the isotopic enrichment of the $\delta^{18}$O of water with respect to the $\delta^{18}$O of vapor is:

$$\delta^{18}O_{v,b} = \left( \frac{\delta^{18}O_{v} + \Delta \delta^{18}O_{b,v}}{\delta^{18}O_{b,v}} \right)$$

The equation for the enrichment of the $\delta^{18}$O of vapor with respect to the $\delta^{18}$O of water is the reciprocal:

$$\delta^{18}O_{v} = \delta^{18}O_{b} + \Delta \delta^{18}O_{v,b}$$

The key parameter in the Craig and Gordon (1965) model is the isotopic composition of the evaporation moisture ($\delta_{e}$), $\delta_{e}$ can be estimated by the equation developed by Craig and Gordon (1965) and modified by Gibson (2002) to use water isotope data in the per mil notation:

$$\delta_{e} = \left( \delta_{v} - \delta_{h_{v}} - \epsilon^{'} \right) / \left( 1 - h + \Delta \epsilon / 1000 \right)$$

where $\delta_{e}$ is the measured isotopic composition of the surface water; $h$ is the measured (modeled or assumed) relative humidity normalized to the surface temperature; $\delta_{h_{v}}$ is the isotopic composition of ambient atmospheric moisture; $\epsilon^{'}$ is the net isotope fractionation between the liquid and vapor phases; $\epsilon = \epsilon + \Delta \epsilon$. When calculating $\delta_{e}$ the isotopic composition of ambient atmospheric moisture, $\delta_{h_{v}}$ is usually presumed to be in equilibrium with the isotopic composition of precipitation, $\delta_{p}$, and therefore, $\delta_{e} = \delta_{p} - \epsilon$ (Gat and Bowser, 1991).

Wolfe et al. (2007) define a theoretical local evaporation line (LEL) that can be determined using three parameters: $\delta_{p}$, the mean amount-weighted isotopic composition of precipitation (measured or modeled); $\delta_{SSa}$, the measured isotopic composition of water in a basin in which evaporation equals inflow ($E/I = 1$) from precipitation, $\delta_{p}$, under isotopic steady-state conditions; and $\delta_{e}$, the theoretical maximum (limiting) isotopic enrichment of water in a desiccating basin. The theoretical LEL defines a straight line that extends from $\delta_{p}$ through $\delta_{SSa}$ to $\delta_{e}$ and represents the isotopic evolution of water in a closed basin with inputs of mean amount-weighted precipitation ($\delta_{p}$) evaporating towards $\delta_{e}$ (Wolfe et al., 2007). Deviations from the LEL occur as a result of differing amounts of seasonal precipitation or unaccounted for ground water influx, which may cause a decrease in the intercept of the LMWL and LEL from $\delta_{p}$ to $\delta_{groundwater}$, and an increase in the slope of the observed relationship between $\delta^{2}$H and $\delta^{18}$O relative to the theoretical LEL (Pham et al., 2009; Wolfe et al., 2007).

The evaporation-to-inflow ratio ($E/I$) can be used to determine the relative influence of evaporation and water inflow to the water balance of a surface water body, as in Gibson et al. (1996), Gibson (2002), Wolfe et al. (2007), Ferguson et al. (2007) and Pham et al. (2009). The $E/I$ ratio thus provides a first order estimate of the fraction of water lost by evaporation (Ferguson et al., 2007). The equation for $E/I$ for $\delta^{18}$O (from Ferguson et al., 2007) incorporates the $\delta_{e}$ parameter, and is expressed as:

$$E/I_{\delta^{18}O} = \left( \delta^{18}O_{v} - \delta^{18}O_{I} \right) / \left( \delta^{18}O_{I} + 1 \right) \left( \delta^{18}O_{v} \right)$$

where $\delta_{v}$ is the isotopic value of input waters into the water body, and $\epsilon = \alpha \epsilon$, where $\alpha$ is the isotope fractionation factor. The remaining parameters were defined above. Calculation of $E/I_{\delta}$ assumes the surface water is (1) a closed watershed receiving water only from precipitation; (2) well mixed and (3) in isotopic steady-state (Gibson and Edwards, 2002). In view of assumption 1, $\delta_{i}$ can be estimated as the isotopic composition of precipitation ($\delta_{p}$).

To determine a LEL for the Okanagan Valley watershed, several parameters needed to be defined. The $\delta_{p}$ value was the isotopic composition of water (precipitation) prior to any evaporation. The value used here was the mean isotopic composition of sixty highland ground water samples as being representative of mean annual highland precipitation input and infiltration, rather than the lower elevation MAP collected in the Valley precipitation stations at Osoyoos, given that most of the watershed recharge area is located well above the Valley drainages. The $\delta_{SSa}$ parameter was determined from (1) the mean isotopic composition of 138 surface water samples collected from the Okanagan Lake and River system ($\delta_{SSa}$), and (2) the mean isotopic composition of 10 upland and closed basin surface water samples, excluding those from Anarchist Mountain and Mt. Baldy where significant evaporation did not appear to have taken place ($\delta_{groundwater}$). The $\delta_{e}$ and $\delta_{SSa}$ parameters are listed in Table 3.

The $\delta_{p}$, $\delta_{SSa}$, and $\delta_{groundwater}$ values were plotted on a $\delta^{2}$H versus $\delta^{18}$O diagram (Fig. 9). A straight line extending through $\delta_{p}$, $\delta_{SSa}$ and $\delta_{groundwater}$ resulted in the equation $\delta^{2}$H = 5.0$\delta^{18}$O – 48.4 for the LEL of the Okanagan Valley. The slope of the LEL (i.e., 5.0) was within the
Table 3

<table>
<thead>
<tr>
<th>Parameter</th>
<th>$\delta^2$H (‰)</th>
<th>$\delta^{18}$O (‰)</th>
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</tr>
<tr>
<td>$\delta_{SSa}$</td>
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<td>-11.0</td>
</tr>
<tr>
<td>$\delta_{SSb}$</td>
<td>-78</td>
<td>-6.0</td>
</tr>
</tbody>
</table>

4.4.4. Calculation of E/I ratio

The E/I ratio for the Okanagan Valley was calculated assuming that all evaporation of surface water within the upper Okanagan watershed took place within creeks and tributaries and Lake Okanagan prior to discharging through the Okanagan river system (see above). The E/I ratio was calculated in terms of $^{18}$O using Eq. (4), and $f$ was the residual water fraction. $\Delta^{18}$O_e was the difference between $\delta^{18}$O_vap and $\delta^{18}$O_water, or in this case between $\delta^{18}$O_{SSa} and $\delta^{18}$O_{P}. For $\delta^{18}$O_{P} was the equilibrium fractionation factor for vapor-water exchange based on a mean annual temperature of 9.7 °C, and $\Delta^{18}$O_{SSa} the kinetic enrichment factor for vapor-water exchange and was based on a slope of 5.0 and a corresponding relative humidity $(h)$ of 78% (Gonfiantini, 1986). The residual water fraction hereby calculated for the Okanagan Lake and River system was approximately 65%, resulting in an average evaporative loss of up to ~35%. This determination was similar to that calculated using the E/I equation.

Potential errors in both E/I and average evaporative loss estimates are substantive and may be due to: (1) uncertainty in the relative amount of evaporation of surface waters taking place within watershed creeks and tributaries prior to entering Okanagan Lake; (2) the relative humidity value (78%) taken from Gonfiantini (1986) which was based on the slope of the LEL for the Okanagan Valley. The relative humidity value used was within the range of measured humidity values for the Okanagan Valley (52–78%); however, it may not be representative of the less humid and drier conditions in the summer/early fall when evaporative isotopic enrichment likely takes place; (3) the temperature used in the calculations (9.7 °C) was the mean annual temperature for the Okanagan Valley.
5. Conclusions

The multi-year collection of precipitation isotopes in the inter-montane, semi-arid, Okanagan Valley of Western Canada resulted in a local meteoric water line of $dD = 6.6(\delta^{18}O) – 22.7$, with annualized amount-weighted mean $\delta^2H$ and $\delta^{18}O$ values of precipitation of $–10.2\%$ and $–12.5\%$, respectively. Seasonality in the isotopic composition of precipitation was evident, where summer precipitation was further affected by processes such as recycling of water vapor and secondary evaporation. The $\delta^2H$ and $\delta^{18}O$ values of surface waters were more positive and offset from the weighted mean annual precipitation in the Okanagan Valley, indicative of basin-scale evaporation of surface waters. However, the Okanagan Lake and its river system were isotopically synchronous, indicating that they behaved as a single well-mixed hydrologic unit. First order isotopic mass-balance modeling suggested about 35% of precipitation in the Okanagan Valley watershed was lost to evaporation prior to and during the ~50 year residence time in Okanagan Lake. This estimate yielded similar results to existing meteorological models. Highland bedrock was recharged with snowmelt and early spring rains, with the isotopic composition largely dependent on elevation. Ground waters in the Valley bottom and benchlands west of Osoyoos Lake, on the other hand, were currently recharged by irrigation water from the Okanagan River system at Osoyoos Lake with little evidence of recharge from the highland bedrock.

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Appendix A. Supplementary material

Supplementary data associated with this article can be found in the online version, at doi:10.1016/j.jhydrol.2011.09.032.

References


