Ephemeral Ponds: Are They the Dominant Source of Depression-Focused Groundwater Recharge?

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Abstract Depression-focused recharge is a concept proposed to explain groundwater recharge in the prairie regions of North America. Topographic depressions in this hummocky landscape collect blowing snow and snowmelt, and occasional runoff during rainfall events. Wetland ponds that form in these depressions lose water to evaporation and infiltration. Some of this infiltration contributes to groundwater recharge, both to shallow aquifers in the weathered near-surface, and to underlying confined intertill aquifers. Here we focus on understanding recharge to the confined aquifers, which supply water for farms and rural communities. The isotopic composition of water in these aquifers shows little or no evaporative enrichment and is inconsistent with the average isotopic composition of the ponds. This observation appears to contradict the depression-focused recharge model. In this field study, we examine the isotopic composition of diverse types of wetland ponds and groundwater at the St. Denis National Wildlife Area, Saskatchewan, Canada. We use hydraulic head data to identify potential recharge and discharge ponds. Water in permanent recharge ponds that do not dry out every year have distinctly different isotopic signatures from the aquifers, suggesting that they cannot be the dominant source of recharge. Water in ephemeral recharge ponds, which are small and dry out quickly, have isotopic signatures identical to those of aquifers. We propose that ephemeral recharge ponds are the dominant source of depression-focused groundwater recharge in the prairies. We discuss why permanent recharge ponds may not be the main source of groundwater recharge and summarize our findings in a revised conceptual model.

Plain Language Summary Prairie wetland ponds have been identified as the primary sources of water for groundwater aquifers in the Northern Glaciated Prairie Plains of North America. The ponds, however, are diverse in their interactions with the subsurface and with one another. Our challenge was to compare the isotopic signature of groundwater with the range of isotopic signatures of the wetlands to determine which wetland ponds were the dominant source of groundwater recharge. Our study at St. Denis, Saskatchewan, a typical prairie setting, used hydrometric measurements and water isotopes. We found that groundwater is replenished from all recharge ponds but that ephemeral ponds—temporary ponds that dry out every year and typically disappear in spring—play the dominant role in groundwater replenishment. These ponds are abundant but seasonally short lived, losing mostly all their water by early summer to infiltration and plant transpiration. We summarize this in a revised conceptual model called the ephemeral wetland pond-focused recharge model.

1. Introduction

It is essential to quantify groundwater recharge rates to sustainably manage groundwater resources, assess the risks of groundwater contamination, and understand the role of groundwater in the hydrological cycle. However, before we can quantify the recharge rates, we must have sufficient qualitative understanding of the recharge mechanisms, which are diverse because it forms the basis of our conceptual model. In turn, the conceptual model of groundwater recharge informs the observations and/or mathematical models that could be useful in providing estimates of recharge rates and determining how these vary in time and space (Healy & Scanlon, 2010).

The glaciated prairie region of North America (an area of approximately 715,000 km²) extends from north-central Iowa in the United States to central Alberta in Canada (Figure 1a). The landscape is hummocky dotted with small depressions, known as “potholes” or “sloughs,” and uplands, whose formation is
attributed to deglaciation at the end of the Pleistocene (Lennox et al., 1988; van der Kamp & Hayashi, 1998). The geology for most of the region is characterized by a complex mix of shallow and weathered till layers of relatively high permeability overlying unweathered till aquitards that are mostly composed of glacial till (Lennox et al., 1988; van der Kamp & Hayashi, 2009). Groundwater within discontinuous confined aquifers in this region is an important water resource for farms and rural communities (van der Kamp & Hayashi, 1998). Groundwater recharge to these aquifers has been described by the well-established “depression-focused recharge” conceptual model (Berthold & Hayashi, 2004; Lissey, 1971; Noorduijn et al., 2018; van der Kamp & Hayashi, 2009). In this conceptual model, depressions collect water from wind-blown snow trapped by vegetation and snowmelt due to runoff over frozen soils (Gray & Landine, 1988; Hayashi et al., 1998a; Woo & Rowsell, 1993). This causes ponds to form, which act as focal points on the landscape for infiltration and evaporation (Bam & Ireson, 2018; Hayashi et al., 2003, 2016; Millar, 1971; Shjeflo, 1968; Sloan, 1972). Ponds exchange water vertically and laterally with shallow groundwater in the permeable, weathered layers of the glacial till, which we refer to as a transmission aquifer (0 to 6 m below ground level (bgl)). The direction and magnitude of these interactions vary temporally (with seasonal and interannual variability driven by meteorological factors) and spatially (where the variability is driven by hydraulic gradients and the position of the ponds in the landscape; Berthold & Hayashi, 2004; Brannen et al., 2015; Hayashi et al., 1998a, 1998b; Hayashi et al., 2016; Heagle et al., 2013; LaBaugh et al., 1998, 2000; Winter & LaBaugh, 2003; Winter & Rosenberry, 1998; Woo & Rowsell, 1993; Zebarth & De Jong, 1989). Underlying the shallow permeable weathered layers, are layers of unweathered clay-rich glacial till, which function as aquitards. The aquitard layers have varying thicknesses, and embedded between them are intermittent coarse-grained deposits, forming confined aquifers of varying extents and sizes, referred to here as intertill aquifers (van der Kamp & Hayashi, 2009). Water percolates slowly through these aquitards to recharge the intertill aquifers. The recharge rates to the intertill aquifers have been estimated by Darcy flux calculations as “a few mm to a few tens of mm per year” (Hayashi et al., 1998b; van der Kamp & Hayashi, 2009), and between 10 and 60 mm/year based on vertical flux estimates of pond hydrographs, groundwater recharge modeling, and chloride mass balance methods (Meyboom, 1966; Noorduijn et al., 2018; Pavlovskii et al., 2019; Rehm et al., 1982; Trudell, 1994; Zebarth & De Jong, 1989). The uplands of the depressions, by contrast, are scoured of snow by the wind, and snowmelt and rainfall infiltration are absorbed by the clay-rich soil which has a large water-holding capacity so that

Figure 1. The St. Denis National Wildlife area and, inset, a map of the Prairie Pothole Region (gray) of North America (van der Kamp & Hayashi, 2009), showing the location of St. Denis National Wildlife Area and the city of Saskatoon in Saskatchewan, Canada. Land use in 2014 is shown, along with the locations of piezometers and the major wetlands (numbered). The piezometers for which tritium samples have been collected are marked DP1, DP2, and 93LP3B. The metric grid uses the World Geodetic System 1984 geographic coordinate system.
most of the infiltrated water is lost to evapotranspiration during the summer (Hayashi et al., 1998a; Woo & Rowsell, 1993).

Although this conceptual model suggests that groundwater recharge through the clay-rich sediments originates from water collected within depressions, it does not follow that all depressions provide groundwater recharge. Some ponds, known as discharge ponds, have a water level below the potentiometric surface of the confined aquifer, and, as a result, they receive a net input of water from groundwater (Lissey, 1971; Sloan, 1972). In addition to being categorized according to recharge or discharge function, ponds are also classified on the basis of their interaction with other surrounding ponds: isolated ponds do not exchange water with other ponds, spilling ponds transmit water to surrounding ponds or streams, flow-through ponds receive input from and transfer output to surrounding ponds or streams via surface flow, and terminal ponds receive input from surrounding ponds or streams but do not spill. Furthermore, ponds are classified on the basis of the longevity of ponding, as in the five classical wetland classifications of Stewart and Kantrud (1971): I—ephemeral, II—temporary, III—seasonal, IV—semipermanent, and V—permanent (Eisenlohr et al., 1972; Hayashi et al., 2016). Here for simplicity, we do not distinguish between class I, II, or III, and refer to all ponds that dry up every year during the summer as ephemeral ponds. This results in three defining pond characteristics: permanence, groundwater interactions, and surface water interactions. These characteristics are in principle independent, so that any pond can be a combination of each of these three characteristics (e.g., a permanent potential recharge pond that is terminal). In this study, we explore how these characteristics may or may not determine whether a pond provides groundwater recharge to underlying confined aquifers. Discharge ponds are identified by their hydraulic head being below the potentiometric surface of the aquifer. In contrast, ponds with hydraulic heads above the potentiometric surface of the aquifer are considered potential recharge ponds.

Despite its popularity, the depression-focused recharge conceptual model is difficult to reconcile with stable isotope observations from ponds and groundwater in the prairies. The isotopic signatures of ponds reflect precipitation water that is enriched in heavy isotopes (Fortin et al., 1991; Gibson et al., 2005; Jasechko et al., 2017; Mcmonagle, 1987; Pham et al., 2009). This enrichment occurs due to kinetic fractionation, which is enhanced by the fast diffusion of lighter $^1$H$_2$^{18}$O isotopologues relative to $^1$H$^2$O as well as H$_2$^{18}$O, from evaporating water surfaces into the open air during nonequilibrium evaporation (Clark, 2015; Gat, 1995a, 1995b; Gibson et al., 2005; Krabbenhoft et al., 1990; Payne, 1981). Pond water therefore plots below the meteoric water line, and we say it is enriched due to evaporation. Groundwater that is recharged by pond water should carry this enriched isotopic signature. On the other hand, groundwater recharged directly by infiltrating precipitation should have an isotopic composition consistent with the seasonally weighted long-term average precipitation inputs (Gat, 1995a; Gat & Gonfiantini, 1981; Gat & Tzur, 1967). Observations of groundwater to depths of 40 m in the glacial deposits, however, show isotopic signatures that resemble meteoric water with a bias toward snowmelt (Fortin et al., 1991; Fritz et al., 1987; Jasechko et al., 2017; Mcmonagle, 1987; Rehm et al., 1982). As a result, the groundwater isotope composition in the North American glacial deposits reflects neither pond water, nor the average precipitation water, nor a weighted combination of the two (which would also plot off the meteoric water line).

In this study, we use observations of hydraulic heads and isotopic composition of ponds and groundwater from the St. Denis National Wildlife Area (SDNWA), Saskatchewan, Canada. The SDNWA site provides a unique opportunity for this study because the existing piezometer infrastructure includes numerous piezometers at various depths installed in the subsurface beneath the centers of many ponds, allowing sampling of the groundwater beneath the ponds. Observations specific to this study, complemented by historic water analyses and pond depths records, are used to assess the applicability of the depression-focused conceptual recharge model and to provide a refined conceptual model for the recharge mechanism. The objective of this study is to determine which types of ponds and at what times they might provide recharge to a confined aquifer at 30 m bgl.

2. The Study Area

SDNWA is located in the prairies (Figure 1a) of Canada (106°060’W and 52°029’N). It was established by Environment and Climate Change Canada (ECCC) as a long-term study site to monitor ecological changes
since the 1960s (Henderson, 2013). SDNWA covers an area of 3.98 km² and is located approximately 35 km east of the city of Saskatoon (Figure 1b).

The climate is dry (subhumid) and seasonally frozen. The monthly mean temperature in Saskatoon varies between −14.7 °C in January and 18.7 °C in July (Wittrock & Beaulieu, 2015). Annual precipitation at SDNWA is approximately 360 mm/year, with around 80 mm/year falling as snow, but this is highly variable from year to year (Bam et al., 2019; Nachshon et al., 2014). The potential evaporation in the region exceeds total annual precipitation and is estimated at 700–800 mm/year, with annual open water (lake) evaporation at approximately 700 mm/year (Parsons et al., 2004).

The ponds at SDNWA are numbered (Figure 1b), and this numbering system is adopted in several studies, including this one. We studied eight ponds in detail for the years 1968 to 2016, where water level records exist (i.e., Ponds 50, 107, 108a, 109, 110, 117, 120, and 125). The ponds have varied water permanence and hydrologic functions and are classified as ephemeral, semipermanent, and permanent. The major surface water connections are shown in Figure 1b.

The vegetation at the site is diverse. Upland areas are covered by crops and planted native grasslands. Wetlands within depressions include a vegetated riparian zone around ponds. The riparian zones are covered either cattails \((\text{Typha latifolia})\) up to 1.5 m in height, or trees, known as “willow rings.” The willow rings comprise balsam poplar \((\text{Populus balsamifera})\), silver willow \((\text{Salix spp.})\) up to 8 m, or trembling aspen \((\text{Populus tremuloides})\) up to 10 m (Nachshon et al., 2014). After the ephemeral ponds have dried out, their centers are covered by sedges \((\text{Carex spp.})\), spike rush \((\text{Eleocharis pp.})\), cattails \((\text{Typha})\), and other species, unless they are not seeded with crops. Riparian zone plants all function as effective snow accumulators, enhance transpiration loses, and, in some cases, reduce open water evaporation by sheltering the pond from the wind (Hayashi, 1996; Nachshon et al., 2014; van der Kamp & Hayashi, 2009). During the study period, 2013 to 2016, floating aquatic species, including submergents such as watermilfoil \((\text{Myriophyllum exalbescens})\) and floating mats of duckweed \((\text{Lemna turionifera})\), were present in some of the ponds.

SDNWA soils developed on stratified, silty glaciolacustrine sediments with low hydraulic conductivity \((K)\), and glacial till of the Battleford and Floral formations of the Middle to Late Wisconsin age (Hayashi et al., 1998a). Orthic Dark Brown Chernozem soils are found in upland areas and on the midslopes and lower slopes Calcareous Dark Brown Chernozems \((\text{Entic Haplustolls})\) and Orthic Regosols \((\text{Typic Ustorthents})\) occurring on knolls all developed from moderate- to fine-textured unsorted glacial till typical of the northern glaciated plains (Miller et al., 1985; van der Kamp et al., 2003).

A geological cross section based on Hayashi (1996) and Miller et al. (1985) is shown in Figure 2a. The uppermost unit (up to 6 m in thickness) is weathered, fractured, and oxidized till, which forms a transmission aquifer (i.e., a permeable unit not capable of exploitation, as is the case for a normally defined aquifer; Brannen et al., 2015; van der Kamp & Hayashi, 2009). The saturated hydraulic conductivity \((K)\) within this layer declines exponentially with depth, from as high as 1,000 m/year in the highly fractured surface to as low as 0.001 m/year in the deep, unfractured base of the layer ~21 m bgl (van der Kamp & Hayashi, 1998, 2009). Below this upper unit is the less fractured till layer, which functions as an aquitard. Discontinuous sand and gravels are encountered sporadically within these deposits, with thicknesses from 0.1 to 1.0 m (Hayashi, 1996). A discontinuous 10-m-thick gravel-sand-silt layer is present at a depth of approximately 32–40 m bgl (Figures 2a and 2b). This layer is an intertill aquifer. This aquifer is part of the regional “Riddell sand \((\text{Qf-ms})\) layer” encountered at the same elevation \((525 \text{ m above sea level (asl)})\) in stratigraphic test hole no. 220582 south-east of Pond 50 (https://www.wsask.ca/Water-Info/Ground-Water/Mapping/Saskatoon-Mapsheet-73B). The position and depth of the intertill aquifer layer correspond with the “Qf‐ms layer” on the G-G’ cross section of the regional 1:250,000 hydrogeology map sheet which includes the SDNWA (MDH, 2011), suggesting that this aquifer is laterally extensive. This aquifer is in the same stratigraphic position (Upper Floral formation) as the Dalmeny aquifer, 50 km to the west of SDNWA, which has been described in detail by Fortin et al. (1991). Its stratigraphic position and depth are typical of many of the confined intertill aquifers that supply groundwater in the northern prairie region.

A large number of piezometers were installed in SDNWA between 1982 and 2012. All the piezometers were installed in the shallow fractured/oxidized till (transmission aquifer) or the unoxidized till (aquitard). The depth of these piezometers ranges from 1.2 to 6 m bgl in the shallow weathered till and >6–21 m bgl in
the aquitard. Collectively, water from these two units (i.e., the shallow weathered zone and the aquitards) are referred to as the shallow groundwater in later discussions. Two new piezometers were installed in the intertill aquifer during this study (described below). The locations of all piezometers used in this study are shown in Figure 1b. Miller et al. (1985) and van der Kamp and Hayashi (2009) showed that the water table beneath the uplands along the transect shown in Figure 1 was at about 548 to 553 m asl during the relatively dry years 1982 and 1994, with water table “mounds” beneath the depressions indicating that depression-focused recharge is dominant at this location during this period.

3. Methods

3.1. Piezometer Data

In October 2013 (during the course of this study), two piezometers were installed in the intertill aquifer, referred to as DP1 and DP2 (Figures 1b and 2a). DP1, northwest of Pond 109, is 39.6 m deep, and DP2, south-west of Pond 125, is 41.5 m deep. Drilling was performed using reverse circulation methods, with pond water as the drilling fluid. PVC pipes (5 cm in diameter) were installed, and both piezometers were screened (0.15–0.30 m) between 32 and 37 m bgl. Sand packs were placed around each of the screens, and 12–16 m of bentonite seals were placed above the sand packs. A grout seal and a mixture of sand and bentonite were placed on top of the bentonite seal. A Solinst Levelogger (LTC Levelogger Edge, Solinst Canada Ltd.) was installed in the two piezometers, which were loosely capped to allow air pressure inside the pipes to equalize with the atmosphere while preventing rainwater from entering them. Glacial deposits, consisting largely of clay-rich glacial till interbedded with thin sand and gravel, were encountered during the drilling in DP1 and DP2, with a sand and gravel layer at elevations of 521 and 519 m asl, respectively, which is the intertill aquifer. The drillers’ log for DP2 is shown in Figure 2b.

A set of seven additional shallow piezometers were also instrumented with Solinst Leveloggers that were programmed to yield hourly output. The barometric pressure at the on-site meteorological station (Figure 1b) was used to remove the effects of atmospheric pressure on the measured total heads. A geodetic survey was conducted between Ponds 109 and 125 in October 2014 to correct the groundwater levels in the intertill aquifer to heads above sea level.

3.2. Pond-Level Data

The water depth at the deepest point of the major ponds at SDNWA, including Ponds 50, 107, 108a, 109, 110, 117, 120, and 125, has been measured manually on a biweekly or monthly basis during the summer since 1968 using a measuring rod with a staff gauge for the water level (#209-SG-1 Staff Gauge, Hoskin
Scientific Limited, BC, Canada) at designated poles in each pond. The difference between the manual measurements (corrected to datum elevation) and the depths measured using a conventional level survey is approximately ±25 mm (Conly et al., 2004). Between 2013 and 2016, the water levels in Ponds 50, 109, and 5340 were monitored hourly using Solinst Leveloggers.

3.3. Water Sampling

Samples of precipitation (snow and rainfall), pond water, and groundwater were collected from the SDNWA for isotope analysis. The sample locations are indicated in Figure 1b, and sampling details are described below.

Precipitation: Wet-dry rain collectors placed at two locations on poles 1 m above the ground away from vegetation at the SDNWA were used to collect rainfall samples during the summers of 2013 and 2014. The collectors consisted of Canadian rain gauge designs Type-B and MSC (Meteorological Service of Canada). The Type-B gauge is a solid white Acrylonitrile Butadiene-Styrene Terpolymer plastic constructed to reduce both evaporative and wetting losses. MSC is made of copper, and Type-B is an upgraded version. The receiver in Type-B can hold 25 mm of rain and was graduated to the nearest 0.2 mm. The overflow of up to 250 mm of rain was retained by the body of the gauge (Metcalfe et al., 1997). Rainfall samples \((n = 24)\) were collected immediately after events when possible. Prior to the installation of collectors, the receiver flasks were carefully cleaned and filled with ~5 mm of paraffin oil (liquid petrolatum) to prevent evaporation of the water samples. The paraffin oil effectively prevents evaporation from the sample container during a collection period of one month (IAEA, 1997, 2004; WMO, 1972). Used in most rain sampling stations for stable isotopes, this standard technique results the smallest evaporative losses and isotope shifts (Gröning et al., 2012; Michelsen et al., 2018; van Geldern et al., 2014). Precipitation amounts were recorded manually and compared well with digital records from the meteorological station at the site. Twenty-three snow samples were collected in mid-March 2013 just before the snow season ended. The snow samples were scooped from the ground into Ziploc® plastic bags, sealed, and taken to the Isotope Geochemistry laboratory of the National Hydrology Research Centre in Saskatoon. The sealed samples stored in plastic bags were allowed to melt under laboratory temperatures before being transferred into polyethylene bottles, capped, and stored for stable isotope analysis.

Pond water: Pond water samples \((n = 251)\) were collected primarily during the open water season in 2013 and 2016 from eight ponds (50, 107, 108a, 109, 110, 117, 120, and 125; Figure 1b). The first samples were collected in May (i.e., immediately after snowmelt when the ponds became accessible), and the collection continued throughout the summer up to the end of October (i.e., before pond freeze-up). The water samples were collected midway between or a few meters from the banks of the ponds using 50-mL polypropylene bottles. After being rinsed with pond water, the bottles were immersed in the pond water, filled, and sealed underwater (Mook, 2001).

Groundwater: Water samples \((n = 82)\) from 21 piezometers installed in the shallow transmission aquifer, aquitard, and the intertill aquifer units (Figure 1b) were sampled for stable water isotope analysis using a bailer or mechanical pump. The samples were collected from piezometers installed in uplands, pond riparian zones (i.e., the pond edges), and centers (i.e., beneath the ponds) of Ponds 50, 107, 108a, 109, 110, 117, 120, and 125 in August 2013, October 2014, and February 2016. The water samples were collected in 25- and 100-mL polypropylene bottles following the protocols described in Mook (2001). Five water samples were collected at depths of 5 and 41 m from piezometers 94W7, 82W9, 93LP3B, DP1, and DP2 in July 2016 and stored in 1,000-mL polypropylene bottles for ultralow-tritium \((^3\text{H})\) analysis (Figure 1b and Table 1), following the Institute of Geological and Nuclear Sciences (GNS), New Zealand protocol.

3.4. Archived Stable Isotope Data

In addition to the data collected above, stable isotope analyses of precipitation for Saskatoon (i.e., from 186 monthly aggregated samples) were retrieved from ECCC records. These data were collected by ECCC between June 1993 and November 2014. The ECCC office in Saskatoon is the closest site with historical isotope records to the St. Denis site. The precipitation samples were collected using a wet-dry rain collector placed on a rooftop (8 m above the ground and 1 m from the floor of the roof) away from the trees. Sampling was done manually on an event basis and aggregated after each month.
Pond and groundwater isotopes were supplemented with monthly historical water isotope data collected from the eight ponds \((n = 90)\) and 31 piezometers \((n = 54)\) between 1993 and 2012. Some piezometers could not be sampled because they were either destroyed by farmers cultivating the land, flooded, or inaccessible during most of the summers when the research was being conducted.

### 3.5. Stable Isotope and Tritium Analyses

Isotope ratios of hydrogen and oxygen \((\delta^2H\text{ and } \delta^{18}O)\) and ultralow-level tritium in groundwater were measured at the ECCC office in Saskatoon, Canada and Rafter Radiocarbon Laboratory, National Isotope Centre, GNS Science in New Zealand, respectively.

Between 1993 and 2007, the precipitation, ponds, and groundwater samples were analyzed for \(\delta^2H\) and \(\delta^{18}O\) using isotope-ratio mass spectrometry. The methods for the water isotope-ratio \((^{2}H/^{1}H\text{ and } ^{18}O/^{16}O)\) analysis followed standard procedures for isotope-ratio mass spectrometry analysis (Begley & Scrimgeour, 1997; Coleman et al., 1982; Eiler & Kitchen, 2001; Epstein & Mayeda, 1953; Karhu, 1997; Kelly et al., 2001; Socki, 1999). All isotope ratios, \(^{18}O/^{16}O\) and \(^{2}H/^{1}H\), were measured relative to internal standards. These ratios were calibrated using Standard Light Antarctic Precipitation (SLAP) and Vienna-SMOW (V-SMOW). The data are reported in the delta notation after normalization following Coplen (1988):

\[
\delta (\%) = \left( \frac{R_s}{R_{V-SMOW}} - 1 \right) \times 1,000
\]

where \(R_s\) represents either the \(^{18}O/^{16}O\) or the \(^{2}H/^{1}H\) ratio of the sample and \(R_{V-SMOW}\) is the \(^{18}O/^{16}O\) or the \(^{2}H/^{1}H\) ratio of the V-SMOW. An analytical reproducibility of \(\pm 0.1\%\) for \(\delta^{18}O\) and \(\pm 1.0\%\) for \(\delta^2H\) and measurement accuracies were \(\pm 0.5\) and \(\pm 4\%\) for \(\delta^{18}O\) and \(\pm 0.1\) and \(\pm 0.4\%\) for \(\delta^2H\) (Lis et al., 2008).

The water samples collected in 2013 and 2016 were analyzed using a Los Gatos Research DLT-100 liquid isotope water analyzer, which was coupled to a CTC LC-PAL liquid autosampler (Los Gatos Inc., CA). The measurement method is described in Lis et al. (2008) and IAEA (2009). Precalibrated internal laboratory standards and a blank were used in sample runs and the results were normalized to the VSMOW-SLAP \(\delta\) scale.

The low-level detection of tritium \((^3H)\) counting was done using a Quantulus™ low-level counter. The analytical procedure is described in Morgenstern and Taylor (2009). The \(^3H\) concentrations are in tritium units (TU); the precision at an average tritium concentration measurement is 98.5%, and the detection limit is \(\pm 0.025\) TU (Morgenstern & Taylor, 2009).

### 4. Results

#### 4.1. Long-Term Pond Isotope Composition and Hydraulic Head Data

Long-term observations of the pond water heads in all eight ponds are shown in Figure 3 for the period 1968 to 2016. The gray shaded areas extend from the final reading in the fall to the first reading in the spring and delineate when and where the ponds carried water from one year to the next. Water-level changes are driven by climatic variability and reflect how individual ponds function regarding surface inflows and outflows and groundwater interactions. Ponds 107, 108a, and 117 are ephemeral. They dried out every year of the study.

| Sample ID | Water type       | Date sampled | Depth (m) | \(\delta^2H(\%)\) | \(^{3}H\) (TU) | sigTU* |
|-----------|------------------|--------------|-----------|-------------------|----------------|--------|
| DP1       | Intertill aquifer| 10/07/2016   | 39.6      | −155.1            | 0.01           | 0.02   |
| DP2       | Intertill aquifer| 10/07/2016   | 41.5      | −155.1            | 0.22           | 0.02   |
| 94W7      | Shallow aquifer  | 10/07/2016   | 5.3       | −139.8            | 5.94           | 0.08   |
| 82W9      | Shallow aquifer  | 10/07/2016   | 6.0       | −140.4            | 4.89           | 0.06   |
| 93LP3B    | Aquitard         | 14/11/2013   | 21.5      | −149.1            | 1.27           | 0.04   |

*SigTU is the 1-standard-deviation uncertainty of the measurements.
usually by early summer. Pond 110 is also considered ephemeral because it dried out most years, with a few exceptions in the post-2004 period. Ponds 109 and 120 are more complex, having characteristics of permanent and ephemeral ponds at different times. Prior to 2004 these ponds often dried out over the summer, but from 2004 to 2016, neither pond dried out due to exceptionally high winter and summer precipitation. The size and behavior of Pond 120 are similar to those of Pond 109, but there is a maximum
capacity beyond which Pond 120 spills (into Pond 125), while Pond 109 continues to fill, leading to notable differences between the water levels of these ponds from 2009 on. Pond 50 has never spilled, and, as a result, is considered terminal. It is also characterized as permanent because it almost always contained water, except during the extreme drought of 2000–2004, when it dried out. Pond 109 has always been considered a terminal pond that does not spill, although, as shown in Figure 3, it is possible that a small amount of spill may have occurred in recent years. Pond 125 is a flow-through pond with its water balance dominated by inflow from Pond 1 which in turn is fed by a large watershed to the north, and outflows to Pond 97, then to 90, and ultimately filling the large terminal Pond 5340, which is the lowest point on the site (see Figure 1b and Shaw et al. (2012) for more details about pond-to-pond connections). Pond 125 dried out twice: in the early 1990s and in the 2000–2004 drought.

Figure 4 compares the variation of pond water levels with pond water $\delta^{18}O$ over time for the two most data rich ponds: 50 and 109. The seasonal variation of $\delta^{18}O$ is similar in both ponds: relatively depleted isotopic composition in spring due to the filling of the pond by snow meltwater, followed progressively by enrichment through the summer/fall due to evaporation. The absence of water in the ponds is attributed to the major drought period 2001–2004. In the period before 2006, Pond 109 was an ephemeral pond, drying out most years, and, as a result, the pond water composition in the spring was highly depleted, reflecting snowmelt. By contrast, in the unusually wet period from 2009 to 2016, Pond 109 behaved like a permanent pond, carrying water over from the previous year. As a result, the snowmelt mixed with enriched pond water, resulting in a less negative isotope signal in the spring.

Observations of the hydraulic heads in the intertill aquifer DP1 and DP2 and in Ponds 50, 107, 108a, 109, 110, 117, 120, 125, and 5340 are shown in Figure 5 for the period after DP1 and DP2 were installed in 2013. These data show the potential groundwater-surface water interactions between the intertill aquifer and these individual ponds. Over the two-year monitoring period, there was no significant head gradient between the two deep piezometers, suggesting that the intertill aquifer has a relatively high K. The seasonal response of the heads in the aquifer was subtle, but present, suggesting that the aquifer receives highly attenuated and delayed seasonally varying recharge through the aquitards. For this period, the heads in Ponds 120, 109, 110, and 117 were 3–6 m higher than those in the intertill aquifer; therefore, these are all potential recharge ponds. The heads in Ponds 50 and 125 were roughly the same, or perhaps moderately elevated, compared with the heads in the aquifer, suggesting that there is a minimal exchange between these ponds and the aquifer, and that these ponds are therefore not potential recharge ponds. The head in Pond 5340 was about 3 m below the intertill aquifer, indicating that this is a discharge pond.

4.2. Stable Isotopes of Water

The isotopic composition of pond water and groundwater (i.e., the shallow transmission aquifer, aquitard, and confined aquifer) at SDNWA (Figure 6) shows distinct isotopic differences among the different water types. The precipitation data from SDNWA plot on the Saskatoon local meteoric water line (SLMWL) and show distinct seasonal variability, with snow being more depleted than summer rainfall. The pond isotope samples plot along a pond evaporation line with a slope of 5.6, starting on or close to the SLMWL in the spring and becoming progressively enriched relative to snow through the summer months. The shallow groundwater data are similar to the pond data, but subject to less evaporative enrichment relative to the pond water. The slight variability in the shallow groundwater isotope data is attributed to the spatial differences in locations and the depth of the piezometers in the landscape. The intertill aquifer data plot on the SLMWL and are biased toward the snow end of the precipitation spectrum.

Local-scale (spatial separations of tens of meters) variations in the isotope composition of the ponds and surrounding groundwater were explored using samples collected from the piezometers in the uplands and from the pond edges (i.e., in the riparian zone) and beneath them (Figure 7). The isotope samples of the intertill aquifer plot on the SLMWL, intersecting with some of the pond waters and are biased toward the depleted end of the pond water. The pond samples plot along the local evaporation line. In some cases, the pond waters intersect the SLMWL. Ponds 50 and 125 have the most enriched isotopic compositions compared to the rest of the ponds and the intertill aquifer. The low evaporation line of Pond 125 is as result of the fact that it is a flow-through pond and dominated by inflows from a larger watershed (see Figure 1). The isotope samples from these ponds also did not plot on the SLMWL. As discussed above, based on the hydraulic gradients, Ponds 50 and 125 are not potential recharge ponds with respect to the intertill aquifer and may or
may not be groundwater-discharge ponds. The distinctly different isotopic composition of these ponds and of the intertill aquifer is consistent with the idea that these ponds do not contribute recharge to the intertill aquifer, but may be fed in part by groundwater discharge from the aquifer and subsequently enriched by

![Figure 4](image-url) 

*Figure 4.* Pond water levels and $\delta^{18}$O compositions. Red dashed line and shaded area represent the mean and standard deviation of $\delta^{18}$O compositions before and after 2009.

![Figure 5](image-url) 

*Figure 5.* Hydraulic head in Ponds 50, 107, 108a, 109, 110, 117, 120, 125, and 5340, and the intertill aquifer, DP1 and DP2, between January 2013 and December 2015.
evaporation. Of the remaining ponds (109, 120, 108a, 107, 117, and 110), which are all potential recharge ponds, all have some isotope samples that plot on the SLMWL and overlap with or are near the intertill aquifer water. As shown in Figures 7 and 8, the ponds are further separated into ephemeral potential recharge ponds (107, 108a, 117, 109 prior to 2006, and 110) and permanent potential recharge ponds (109 since 2006 and 120). For Pond 109, there is a clear difference in isotopic composition between when the pond is ephemeral (prior to 2006) and permanent (since 2006), as shown in Figures 4 and 8.

The isotopic results for the shallow groundwater beneath the ponds and the pond edges, shown in Figure 7, seem to show that shallow groundwater near the ephemeral ponds overlaps with the intertill aquifer water and extends slightly toward the least evaporated pond waters. The shallow groundwater near the permanent discharge ponds is distinctly different from the aquifer water. The isotopic composition of groundwater beneath the uplands is variable, sometimes extending toward and even overlapping with the aquifer water (Ponds 50, 109, and 125), and sometimes quite distinct (Pond 120 and some data points from Ponds 109 and 125).

### 4.3. Tritium, $^3$H

Tritium samples were collected from five groundwater sources (Table 1). All samples were tritiated, except in the case of DP1, where the very low TU value may be considered nontritiated. Shallow groundwater had the highest $^3$H values, and values decreased with depth. Since the cessation of atmospheric nuclear tests, $^3$H concentrations in precipitation in the region have dropped to an average of 10 TU, although nuclear power plants may add small concentrations (Gleeson et al., 2015; Wassenaar, 1995). Given that the half-life of $^3$H is 12.3 years, then a very approximate calculation (ignoring diffusion and dispersion) suggests groundwater ages of around 10 years at 5.3–6.0 m bgl (i.e., modern water), 37 years at 21.5 m bgl (i.e., a mix of submodern and modern water), and 70 and 120 years in two samples from the intertill aquifer (i.e., submodern water; Clark & Fritz, 1997). The calculated travel times suggest slow advective flow in the till at pore water
velocities in the range of 0.27–0.6 m/year. This pore water velocity is higher than previously reported for Darcy flux and model calculations (Noorduijn et al., 2018; van der Kamp & Hayashi, 1998, 2009) and may be uncertain due to the heterogeneous nature of the till matrix, leading to varying degrees of diffusion/dispersion (Hendry et al., 2011, 2015; Simpkins & Bradbury, 1992). Fracture flow within the till could deliver water at flow velocities greater than those reported for piston flow. The intertill aquifer $^3$H concentrations are likely too low to be a direct product of bomb-pulse tritium but could be produced by the mixing of older water with bomb-pulse affected water.

5. Discussion

5.1. Is There Any Groundwater Recharge to the Confined Aquifer System?

The first question that we need to consider is this: is it possible that the groundwater in the intertill aquifer is not recharged by recent precipitation at all, but was recharged thousands of years ago by glacial meltwater? Stable isotopes of water are unable to answer this question directly, since the isotopic composition of recent snow and snowmelt ($\delta^{18}O$ between $-25\%o$ and $-20\%o$; Figure 6) is consistent with the composition of water of “glacial age” (i.e., more than 10,000 years old, as shown by Keller et al. (1988), with $\delta^{18}O$ values around

Figure 7. Isotopic composition of ponds, shallow groundwater, and intertill aquifer.
The observation of tritium in the aquifer and the overlying aquitard is consistent with slow percolation of recharge water, which in the aquifer might be prebomb pulse water, 70–120 years old. The most likely interpretation is that this is evidence that the aquifer is being recharged. However, there is a possibility that cannot be completely ruled out that the tritium concentrations are the result of contamination of the intertill aquifer with drilling fluid. Studies that have looked at similar aquifers in the North American Pleistocene glacial till have reported different answers to this question. Some works show significant $^3$H values in aquifers at depths of 20–60 m bgl (Bacon & Keller, 1998; Fortin et al., 1991; Gleeson et al., 2015; Pavlovskii et al., 2019). Other studies have reported groundwater and pore water at 30–46 m depths with an age of 20,000 years, but these aquifers generally were located within very thick glacial till aquitards with very low $K$ (Fortin et al., 1991; Hendry & Wassenaar, 1999; Keller et al., 1988).

### 5.2. Potential Recharge Areas

Potential recharge areas are identified as parts of the landscape where the free water surface (ponds) or ground surface (uplands) are located above the potentiometric surface of the confined aquifer. Ponds or uplands below the potentiometric surface can be ruled out as a potential source of recharge and are designated as discharge areas. The potentiometric surface at SDNWA is at an elevation of 549 m asl, and is assumed to be laterally extensive and uniform over the site. The digital elevation model of SDNWA shows that the potential recharge area makes up 73% of the 4.0–km$^2$ total land area. Of the potential recharge area, only about 8% was water covered during the wet conditions after 2006 (i.e., the ponds that have been formally mapped; Figure 1). This may be an underestimation, as some ephemeral ponds may not be registered as wetlands in wetland inventories based on aerial-image analysis (Hayashi et al., 2016). The discharge areas, which make up 27% of the landscape, were 70% water covered. Hence, the potential recharge areas are predominantly covered by uplands, while the discharge areas are predominantly covered by ponded depressions during wet climatic conditions.

Turning to the specific ponds that were studied, Pond 50 has been reported to be a discharge pond (Heagle et al., 2013; Nachshon et al., 2013; Pennock et al., 2014) due to the high salinity of the pond water, and local-scale head gradients beneath the pond, pond edges, and the shallow transmission aquifer and the aquitards (Heagle et al., 2013). We show that the head difference between Pond 50 and the intertill aquifer is small, and, since 2013 (when observations in the intertill aquifer became available), the head in the pond has been slightly above the aquifer. It is probably correct that, historically, Pond 50 functioned as a discharge pond (Heagle et al., 2013), albeit with only moderate gradients and small fluxes, and that these fluxes may have stalled in the recent period. Pond 5340 is likely to be a much more important discharge pond for water from the intertill aquifer via the aquitards between the two units, as indicated by free water surface elevation (typically 3 m lower than the potentiometric head of the aquifer) and the high salinity of the pond water.

![Figure 8. Temporal variations in Pond 109 isotopic composition (a) 1993–2005 when the pond was ephemeral and (b) 2012–2016 when the pond was permanent. Also included are the shallow groundwater that surrounded Pond 109 and the intertill aquifer at SDNWA.](image-url)
Pond 125 functions both as a discharge and flow-through pond (Brannen et al., 2015) and is part of the drainage network of a larger watershed (~10 km²), so the water balance of this pond is likely dominated by surface processes. All the other ponds studied were potential recharge ponds, with heads at least 3 m higher than the head in the intertill aquifer.

5.3. Ephemeral Versus Permanent Ponds

We can gain further understanding of which of the potential recharge ponds might actually be a source of recharge to the intertill aquifer based on stable isotope compositions of the ponds and groundwater. The composition of stable isotopes in samples from the intertill aquifer plot on the SLMWL (i.e., with no evidence of evaporative enrichment) and reflect a dominance of snow (i.e., relatively depleted compared to the mean precipitation signal). Water in ponds can contain both rainfall and snowmelt and is subject to evaporative enrichment. Variability in isotopic composition between ponds is due primarily to permanence: permanent ponds carry enriched evaporated water over from previous years, while ephemeral ponds are reset each spring with the meteoric snowmelt signal (Bam & Ireson, 2018).

The only place and time when the isotopic signatures of pond water and intertill aquifer water are the same are in the ephemeral ponds during the spring, before the pond water is enriched by late spring and summer evaporation. To explain this, we consider the seasonal pattern of infiltration, transpiration, and groundwater recharge from ephemeral ponds. Ephemeral ponds dry out after the spring snowmelt period, typically within a few weeks, but sometimes persist for a few months into the summer (Hayashi et al., 2016; Rehm et al., 1982; Stewart & Kantrud, 1971). Immediately postmelt, the pond volume is at its maximum (van der Kamp & Hayashi, 2009). Because head gradients between the pond and the aquifer are also at their maximum, infiltration rates below the pond are too (Hayashi et al., 2003). Water that infiltrates into the ground below the pond can either contribute to deeper groundwater recharge or be lost laterally to transpiration from the riparian zone (Berthold & Hayashi, 2004; Hayashi et al., 1998a; Millar, 1971; Parsons et al., 2004; Siegel, 1988; van der Kamp & Hayashi, 2009). In the early spring, evaporation and transpiration rates are at their minimum, and thus, groundwater recharge is at its maximum at the time when the pond isotopic signal is the same as that of snow melt with little or no evaporative enrichment (Mohammed et al., 2019; Pavlovskii et al., 2019). As time progresses, the amount of transpiration increases as riparian vegetation becomes active, meaning less of this water that infiltrates below the pond becomes deep groundwater recharge. Furthermore, even without the enhanced transpiration losses, the enriched isotopic signal observed in the ponds relative to the aquifer water later in the season is associated with progressively smaller volumes of water in the pond and thus smaller volumes of recharge. After the ponds dry up, the most recently infiltrated pond water is lost to root uptake by vegetation in the previously ponded area (Hayashi et al., 2016; Meyboom, 1966), thereby further reducing the amount of enriched water relative to snow melt that infiltrates into the soil beneath the depressions.

If a permanent pond provided recharge to the aquifer, the isotopic composition of that water would be not seasonally biased but would reflect the average isotopic composition of the pond, that is, a mix of rainfall, snowmelt, and evaporatively enriched waters carried over from previous years. The isotopic data indicate that the permanent potential recharge ponds at SDNWA (Pond 109 since 2006) are distinct from the aquifer water, leading to the conclusion that permanent recharge ponds cannot be the dominant source of groundwater recharge to the aquifer. It is still possible that they do provide some recharge, but that this is smaller in volume than the recharge provided by ephemeral ponds or uplands. The smaller volume could be due to a combination of spatial averaging (for example at SDNWA, more area is covered by ephemeral ponds and uplands than by permanent ponds, which make up less than 8% of the potential recharge area), and temporal averaging of groundwater recharge signals over a long time span, perhaps roughly the last 100 years (see Eisenlohr et al., 1972; Clark et al., 2002; S.C. Fritz et al., 2000; Hayashi et al., 2016; Johnson et al., 2004; LaBaugh et al., 2016; Laird et al., 2003; Levy et al., 2018; Ryberg et al., 2016; Shapley et al., 2005; van der Kamp et al., 2008; Winter & Rosenberry, 1998; Winter & Wright, 1977; Yu et al., 2013 and articles in Wetlands, Volume 36, Supplement 2, 2016). During this period, the now permanent ponds may have behaved as ephemeral ponds, as did Pond 109 prior to 2009. It is also possible that the permanent ponds supply disproportionately lower rates of groundwater recharge due to differences in the hydraulic properties. Studies have shown that infiltration rates are higher under the ephemeral ponds (Bam & Ireson, 2018;
Millar, 1971; Mohammed et al., 2019; Pavlovskii et al., 2019; Rehm et al., 1982; Shjeflo, 1968; Sloan, 1972), and suggest that large permanent ponds accumulate considerably more low-K sediments (Richardson et al., 1994).

5.4. Uplands Versus Depressions

Another possibility is that recharge to the intertill aquifer is not only depression focused but also supplied by uplands. The groundwater head measurements in upland piezometers in this study vary between 544 and 553 m asl, compared with the piezometric head in the confined aquifer of 549 m asl, suggesting that flow directions are variable and also strongly influenced by shallow, transient lateral interactions with the depressions (Brannen et al., 2015; Hayashi et al., 1998a; Heagle et al., 2013; Miller et al., 1985). The isotopic compositions of water from upland piezometers, as shown in Figure 7, are also variable, with some data overlapping with the confined aquifer waters (those around Pond 50 and some data around Ponds 109 and 125) and others not (those around Ponds 110 and 120 and some data around Ponds 109 and 125). These data are therefore inconclusive with regard to the significance of upland recharge to the intertill aquifer. We can rule out infiltration of rainfall in the uplands as the dominant source of recharge, since the aquifer waters are biased toward snow (see Figure 6b; also reported in Fortin et al., 1991; Fritz et al., 1974, 1987; Jasechko et al., 2017). This is unsurprising because rainfall that infiltrates is intercepted by the high water-retention capacity clay-rich soils and is expected to be lost to evapotranspiration in this semiarid environment (Hayashi et al., 1998a; Woo & Rowsell, 1993). However, where the upland samples are similar to the aquifer (Figure 7), we cannot rule out the possibility that some recharge also occurs by preferential
infiltration of snowmelt water through macropores in the frozen soils beneath the uplands (Hayashi et al., 1998a; Hendry et al., 2015; Maule et al., 1994; Mohammed et al., 2019; Stumpp & Hendry, 2012). The infiltrating water would have to move down through cracks in the frozen soils, as observed by van der Kamp et al. (2003) and not be affected by subsequent evaporative enrichment.

5.5. The Refined Hypothesis

In the above discussion we have conclusively shown that permanent potential recharge ponds cannot be the dominant source of groundwater recharge to the intertill aquifer at St. Denis. We show that ephemeral potential recharge ponds could be the dominant source of recharge, but so could snowmelt infiltration from upland areas. We hypothesize that the ephemeral ponds will be the dominant source of recharge, since they collect more snow and snowmelt than the uplands. Our conceptual understanding of this system is represented in Figure 9. The model is consistent with the existing conceptual models on groundwater-surface water interactions in the prairies (e.g., Hayashi et al., 2003; Lissey, 1971; Stumpp & Hendry, 2012; van der Kamp & Hayashi, 2009) and provides additional details about which ponds recharge the intertill aquifers and when this recharge occurs. The model depicts three types of ponds: a permanent potential recharge pond, a permanent discharge pond, and five ephemeral potential recharge ponds. Each pond has a surface water catchment from which it collects the snow and snowmelt. Because there are more ephemeral ponds, they cover the largest effective area. Although all ponds lose water through the summer due to evaporation and infiltration only the ephemeral ponds dry out completely. They therefore may be overlooked in mapping exercises and by farmers. Recharge from the permanent potential recharge ponds is a relatively small component of the total recharge.

Care should be taken in generalizing these findings to the wider prairie region, and in particular, the relative importance of ephemeral ponds and upland areas in supplying recharge is still uncertain. Pavlovskii et al. (2019) and Mohammed et al. (2019) studied infiltration and recharge processes at various prairie sites in Alberta using tracer and physical hydrological observations. Both studies concluded that recharge was depression focused, consistent with our findings, but that preferential infiltration of snowmelt into frozen soils was highly significant. Winter and Rosenberry (1998) and Winter & LaBaugh (2003) reported significant upland recharge, based on water table rises, for a hummocky moraine landscape with a permanent grass land cover at the Cottonwood Lake Area in North Dakota. It is likely that variations in geology, climate, and land cover throughout the prairies lead to differences in the importance of uplands versus ephemeral ponds for supplying recharge.

6. Conclusions

Based on field observations of hydraulic head, tritium, and stable isotopes of water, we propose a refined depression-focused recharge hypothesis: that groundwater recharge in the Canadian prairies is depression focused and dominated by snowmelt water from ephemeral ponds. Head data were used to identify potential recharge areas, and stable isotopes were used to show that permanent ponds within these recharge areas could not be the dominant source of groundwater recharge. We were unable to be conclusive about the possible significance of snowmelt infiltration from uplands as a source of groundwater recharge. We also showed that tritium observations were inconclusive, although they were not inconsistent with our hypothesis. These findings suggest that if we wish to enhance the sustainability of groundwater resources in the prairies, it may be important to focus wetland conservation efforts on preserving the small, often overlooked depressions that host ephemeral ponds but which play a critical role in providing groundwater recharge.

References

Bacon, D. H., & Keller, C. K. (1998). Carbon dioxide respiration in the deep vadose zone: Implications for groundwater age dating. *Water Resources Research*, 34(11), 3069. https://doi.org/10.1029/98WR02045

Bam, E. K. P., Brannen, R., Budhathoki, S., Ireson, A. M., Spence, C., & van der Kamp, G. (2019). Meteorological, soil moisture, surface water, and groundwater data from the St. Denis National Wildlife Area, Saskatchewan, Canada. *Earth System Science Data*, 11(2), 553–563. https://doi.org/10.5194/essd-11-553-2019

Bam, E. K. P., & Ireson, A. M. (2018). Quantifying the wetland water balance: A new isotope-based approach that includes precipitation and infiltration. *Journal of Hydrology*, 570, 185–200. https://doi.org/10.1016/j.jhydrol.2018.12.032

Begley, I. S., & Scrimgeour, C. M. (1997). High-precision $^{2}H$ and $^{18}O$ measurement for water and volatile organic compounds by continuous-flow pyrolysis isotope ratio mass spectrometry. *Analytical Chemistry*, 69(8), 1530–1535. https://doi.org/10.1021/ac960935r
Berthold, S., & Hayashi, M. (2004). Integrated hydrogeological and geophysical study of depression-focused groundwater recharge in the Canadian prairies. *Water Resources Research, 40*, W06S05. https://doi.org/10.1029/2003WR002982

Brannen, R., Spence, C., & Ireson, A. (2015). Influence of shallow groundwater-surface water interactions on the hydrologic connectivity and water budget of a wetland complex. *Hydrological Processes*, 29(18), 3862–3877. https://doi.org/10.1002/hyp.10563

Clark, I. (2015). *Groundwater geochemistry and isotopes* (1st ed.). Boca Raton, FL: Taylor & Francis Group, LLC. https://doi.org/10.1201/9781498790425

Eisenlohr, W. S., Schjei, M., Groening, M., Lutz, H. O., Roller (2010). *(unpublished doctoral dissertation). Waterloo, Canada: University of Waterloo.

Hayashi, M., van der Kamp, G., & Rudolph, D. L. (1998a). Water and solute transport between a prairie wetland and adjacent upland. 1. Water balance. *Journal of Hydrology*, 207(1–2), 42–55. https://doi.org/10.1016/S0022-1694(98)00096-5

Hayashi, M., van der Kamp, G., & Rudolph, D. L. (1998b). Water and solute transport between a prairie wetland and adjacent upland: 2. Chloride cycle. *Journal of Hydrology*, 207(1–2), 56–67.

Hayashi, M., van der Kamp, G., & Schmidt, R. (2003). Focused infiltration of snowmelt water in partially frozen soil under small depressions. *Journal of Hydrology*, 270(3), 214–229. https://doi.org/10.1016/S0022-1694(02)00287-1

Heagle, D., Hayashi, M., & van der Kamp, G. (2013). Surface-subsurface salinity distribution and exchange in a closed-basin prairie wetland. *Journal of Hydrology*, 478, 1–14. https://doi.org/10.1016/j.jhydrol.2012.05.054

Healy, R. W., & Scanlon, B. R. (2010). Estimating groundwater recharge (1st ed., Vol. 1). Cambridge, UK: Cambridge University Press.

Henderson, D. (2013). St. Denis national wildlife area management plan. Environment Climate Change Canada, Gatineau. https://doi.org/10.3666/CW61-325/2013E-PDF

Hendy, M. J., Barbour, S. L., & Schmeling, E. E. (2015). Defining near-surface groundwater flow regimes in the semi-arid glaciated plains of North America. *Isotopes in Environmental and Health Studies, 51*(3), 203–213. https://doi.org/10.1080/10256016.2016.1069266

Hendy, M. J., Barbour, S. L., Zettl, J., Chostner, V., & Wassenaar, L. I. (2011). Controls on the long-term downward transport of $^3$H of water in a regionally extensive, two-layered aquitard system. *Water Resources Research, 47*, W06505. https://doi.org/10.1029/2010WR010044

Hendy, M. J., & Wassenaar, L. I. (1999). Implications of the distribution of $^3$H in pore waters for groundwater flow and the timing of geologic events in a thick aquitard system. *Water Resources Research, 35*(6), 1751–1760. https://doi.org/10.1029/1999WR900046

IAEA (1997). Technical procedure for cumulative monthly sampling of precipitation for isotopic analyses. Vienna, AT: International Atomic Energy Agency.
IAEA (2004). Technical procedure for sampling. IAEA-WMO Programme on Isotopic Composition of Precipitation: Global Network of Isotopes in Precipitation (GNIP). Download from GNIP Homepage at International Atomic Energy Agency under: <http://www-naweb.iaea.org/nac/nib/documents/userupdate/sampling.pdf>

IAEA (2009). Laser spectroscopic analysis of liquid water samples for stable hydrogen and oxygen isotopes. Isotope Hydrology Section International Atomic Energy Agency, p. 49.

Jasechko, S., Wassenaar, L. I., & Mayer, B. (2017). Isotopic evidence for widespread cold-season-biased groundwater recharge and young streamflow across central Canada. Hydrological Processes, 31(12), 2196–2209. https://doi.org/10.1002/hyp.11175

Johnson, W. C., Boettcher, S. E., Poiani, K. A., & Gunterseugen, G. (2004). Influence of weather extremes on the water levels of glaciated prairie wetlands. Wetlands, 24(2), 385–398. https://doi.org/10.1672/077–5212(2004)24[385:WET8OT]2.0.CO;2

Kachur, J. A. (1997). Catalytic reduction of water to hydrogen for isotopic analysis using zinc containing traces of sodium. Analytical Chemistry, 69(22), 4728–4730. https://doi.org/10.1021/ac9704467

Keller, C. K., Van Der Kamp, G., & Cherry, J. A. (1988). Hydrogeology of two Saskatchewan tills: 1—Fракtures, bulk permeability, and spatial variability of downward flow. Journal of Hydrology, 101(1–4), 97–121. https://doi.org/10.1016/0022-1694(88)90303-3

Kelly, S. D., Heaton, K. D., & Berenot, P. (2001). Deuterium/hydrogen isotope ratio measurement of water and organic samples by continuous-flow isotope ratio mass spectrometry using chromium as the reducing agent in an elemental analyzer. Rapid Communications in Mass Spectrometry, 15(15), 1283–1286. https://doi.org/10.1002/rcm.303

Krahenbuhl, D. P., Bowser, C. J., Anderson, M. P., & Valley, J. (1990). Estimating groundwater exchange with lakes: 1—The stable isotope mass balance method. Water Resources Research, 26(10), 2445–2453.

LaBaugh, J. W., Mushet, D. M., Rosenberry, D. O., Euliss, N. H. Jr., Goldhaber, M. B., Mills, C. T., & Nelson, R. D. (2016). Changes in pond water levels and surface extent due to climate variability at solute sources to closed-basin prairie- pothole wetlands, 1979 to 2012. Wetlands, 36(2), 343–355. https://doi.org/10.1007/s11267–016-0808–x

LaBaugh, J. W., Winter, T. C., & Rosenberry, D. O. (1998). Hydrologic functions of prairie wetlands. Great Plains Reviews, 1(1), 17–37.

Laird, K. R., Cumming, B. F., Wunsm, S., Rusak, J. A., Oglesey, R. J., Frits, S. C., & Leavitt, P. R. (2003). Lake sediments record large shifts in moisture regimes across the northern prairies of North America during the past two millennia. Proceedings of the National Academy of Sciences, 100(5), 2483–2488. https://doi.org/10.1073/pnas.0530193100

Lennox, D. H., Maathuis, H., & Pederson, D. (1988). Region 13, western glaciated plains. In W. Back, J. S. Rosenshein, & P. Seaber (Eds.), The Geology of North America, Vol. 2–2, pp. 115–128. Boulder, CO: The Geological Society of America.

Levy, Z. F., Rosenberry, D. O., Moucha, R., Mushet, D. M., Goldhaber, M. B., LaBaugh, J. W., et al. (2018). Drought-induced recharge promotes long-term storage of porewater salinity beneath a prairie wetland. Journal of Hydrology, 557, 391–406. https://doi.org/10.1016/j.jhydrol.2017.12.005

Lis, G., Wassenaar, L. I., & Hendry, M. J. (2008). High-precision laser spectroscopy D/H and 18O/16O measurements of microliter natural water samples. Analytical Chemistry, 80(1), 287–293. https://doi.org/10.1021/ac701716q

Lissey, A. (1971). Depression

Meyboom, P. (1966). Unsteady groundwater flow near a willow ring in hummocky moraine. Journal of Hydrology, 4, 38–62.

Michelsen, N., van Geldern, R., Rolfsen, Y., Bauer, I., Schulz, S., Barth, J. A. C., & Schüth, C. (2018). Comparison of precipitation collectors used in isotope hydrology. Chemical Geology, 488(1January), 171–179. https://doi.org/10.1016/j.chemgeo.2018.04.032

Millar, J. B. (1971). Shoreline-area ratio as a factor in rate of water loss from small sloughs. Journal of Hydrology, 14, 259–284.

Miller, J. J., Acton, D. P., & St. Arnaud, R. J. (1985). The effect of groundwater on soil formation in a morainal landscape in Saskatchewan. Canadian Journal of Soil Science, 65(2), 293–307. https://doi.org/10.4141/cjss85–033

Mohammed, A. A., Pavlovskii, I., Cey, E. E., & Hayashi, M. (2019). Effects of preferential flow on snowmelt partitioning and groundwater recharge in frozen soils. Hydrology and Earth System Sciences, 23(1), 5017–5030. https://doi.org/10.5194/hess-23-5017-2019

Mook, W. G. (2001). Environmental isotopes in the hydrological cycle: Principles and applications, Volume 1: Introduction: Theory, Methods, Review. International Hydrological Programme IHP-V. Paris/Vienna: International Atomic Energy Agency and United Nations Educational, Scientific and Cultural Organization. Retrieved from http://www.iaea.org/water/EarthResources/149/environmental-isotopes-in-the-hydrological-cycle-principles-and-applications.html}

Morgenstern, U., & Taylor, C. B. (2009). Ultra-low-level tritium measurement using electrolytic enrichment and LSC. Isotopes in Environmental and Health Studies, 45(2), 96–117. https://doi.org/10.1080/10603560902931194

Nachshon, U., Ireson, A., van der Kamp, G., Davies, S., & Wheeler, H. S. (2014). Impacts of climate variability on wetland salinization in the North American prairies. Hydrology and Earth System Sciences, 18(4), 1251–1263. https://doi.org/10.5194/hess-18-1251-2014

Nachshon, U., Ireson, A., van der Kamp, G., & Wheeler, H. S. (2013). Sulfate salt dynamics in the glaciated plains of North America. Journal of Hydrology, 499, 188–199. https://doi.org/10.1016/j.jhydrol.2013.07.001

Nachshon, U. S. L., Hayashi, M., Mohammed, G. A., & Mohammed, A. A. (2018). A coupled soil water balance model for simulating depression-focused groundwater recharge. Vadose Zone Journal, 17, 170176. https://doi.org/10.2136/vzj2017.10.0176

Parsons, D. F., Hayashi, M., & van der Kamp, G. (2004). Infiltration and solute transport under a seasonal wetland: Bromide tracer experiments in Saskatoon, Canada. Hydrological Processes, 18(11), 2011–2027. https://doi.org/10.1002/hyp.1345

Pavlovskii, I., Hayashi, M., & Cey, E. E. (2019). Estimation of depression-focused groundwater recharge using chloride mass balance: Problems and solutions across scales. Hydrogeology Journal, 27(6), 2263–2278. https://doi.org/10.1007/s10040–019–01993–2

Payne, B. R. (1981). Practical applications of stable isotopes to hydrological problems. Chapter 13. In J. R. Gat & R. Gonfiantini (Eds.), Stable Isotope Hydrology, IAEA Technical Report Series (Vol. 210, pp. 303–334). Vienna, Austria: IAEA.

Pennock, D., Bedard-Haucon, A., Kiss, J., & van der Kamp, G. (2014). Application of hydropedology to predictive mapping of wetland soils in the Canadian Prairie Pothole Region. Geoderma, 235–236, 199–211. https://doi.org/10.1016/j.geoderma.2014.07.008

WMO Programme on Isotopic Composition of Precipitation: Global Network of Isotopes in Precipitation (GNIP). Download from GNIP Homepage at International Atomic Energy Agency under: <http://www-naweb.iaea.org/nac/nib/documents/userupdate/sampling.pdf>

IAEA-WMO Programme on Isotopic Composition of Precipitation: Global Network of Isotopes in Precipitation (GNIP). Download from GNIP Homepage at International Atomic Energy Agency under: <http://www-naweb.iaea.org/nac/nib/documents/userupdate/sampling.pdf>
BAM ET AL. 19 of 19

Yu, G., Sauchyn, D., & Li, Y. F. (2013). Drought changes and the mechanism analysis for the North American Prairie. *Journal of Hydrology*, 593(3–4), 293–314. https://doi.org/10.1016/j.jhydrol.2014.07.011

Rehm, B. W., Moran, S. R., & Groenewold, G. H. (1982). Natural groundwater recharge in an upland area of central North Dakota, U.S.A. *Journal of Hydrology*, 59(3–4), 293–314. https://doi.org/10.1016/0022-1694(82)90093-2

Richardson, J. L., Amdt, J. L., & Freeland, J. (1994). Wetland soils of the prairie pothole. *Advances in Agronomy*, 52, 121–171.

Ryberg, K. R., Vecchia, A. V., Akyüz, F. A., & Lin, W. (2016). Tree-ring-based estimates of long-term seasonal precipitation in the Souris River Region of Saskatchewan, North Dakota and Manitoba. *Canadian Water Resources Journal*, 41(3), 412–428. https://doi.org/10.5589/cwrj.2016.1164627

Shapley, M. D., Johnson, W. C., Engstrom, D. R., & Osterkamp, W. R. (2005). Late-Holocene flooding and drought in the Northern Great Plains, USA, reconstructed from tree rings, lake sediments and ancient shorelines. *Holocene*, 15(1), 29–41.

Shaw, D. A., Vanderkamp, G., Conly, F. M., Pietronio, A., & Martz, L. (2012). The fill-spill hydrology of prairie wetland complexes during drought and deluge. *Hydrological Processes*, 26(20), 3147–3156. https://doi.org/10.1002/hyp.8390

Stjeljo, J. (1968). Evapotranspiration and the water budget of prairie potholes in North Dakota: U.S. Geological Survey Professional Paper, 585-B, 49.

Siegel, D. I. (1988). The recharge

Pham, S. V., Lecquigny, P. R., McGowan, S., Wissel, B., & Wassenaar, L. I. (2009). Spatial and temporal variability of prairie lake hydrology as revealed using stable isotopes of hydrogen and oxygen. *Limonology and Oceanography*, 54(1), 101–118. https://doi.org/10.4319/lo.2009.54.1.0101

van Geldern, R., Baier, A., Subert, H. L., Kowol, S., Balk, L., & Barth, J. A. C. (2014). Pleistocene paleo

Winter, T. C., & Wright, H. E. (1977). Paleohydrologic phenomena recorded by lake sediments. *Eos, Transactions American Geophysical Union*, 58(4), 188–196. https://doi.org/10.1029/EO058i004p00188

Winter, T. C., & Rosenberry, D. O. (1998). Hydrology of Prairie pothole wetlands during drought and deluge: A 17

van der Kamp, G., Hayashi, M. (1998). The groundwatter recharge function of small wetlands in the semi-arid northern prairies. *Great Plains Research*, 8(1), 39–56.

van der Kamp, G., & Hayashi, M. (2009). Groundwater-wetland ecosystem interaction in the semiarid glaciated plains of North America. *Hydrogeology Journal*, 17(1), 203–214. https://doi.org/10.1007/s10040-008-0367-1

van der Kamp, G., Hayashi, M., & Gallén, D. (2003). Comparing the hydrology of grassed and cultivated catchments in the semi-arid Canadian prairies. *Hydrological Processes*, 17(3), 559–575. https://doi.org/10.1002/hyp.1157

van der Kamp, G., Keir, D., & Evans, M. S. (2008). Long-term water-level changes in closed-basin lakes of the Canadian Prairies. *Canadian Water Resources Journal*, 33(January 2008), 23–38. https://doi.org/10.4296/cwrj3301023

van Geldern, R., Baier, A., Subert, H. L., Kowol, S., Balk, L., & Barth, J. A. C. (2014). Pleistocene paleo-groundwater as a pristine fresh water resource in southern Germany: Evidence from stable and radiogenic isotopes. *Science of the Total Environment*, 496, 107–115. https://doi.org/10.1016/j.scitotenv.2014.07.011

Wassenaar, L. I. (1995). Evaluation of the origin and fate of nitrate in the Abbotsford Aquifer using the isotopes of 15N and 18O in NO3. *Applied Geochemistry*, 10(4), 391–405. https://doi.org/10.1016/0883-2927(95)00013-A

Winter, T. C. (1999). Relation of streams, lakes, and wetlands to groundwater flow systems. *Hydrogeology Journal*, 7(1), 28–45. https://doi.org/10.1007/s100400500178

Winter, T. C. (2000). The vulnerability of wetlands to climate change: A hydrologic landscape perspective. *Journal of the American Water Resources Association*, 36(2), 305–311. https://doi.org/10.1111/j.1752-1688.2000.tb04269.x

Winter, T. C., & LaBough, J. (2003). Hydrologic considerations in defining isolated wetlands. *Wetlands*, 23(3), 532–540.

Winter, T. C., & Rosenberry, D. O. (1998). Hydrology of Prairie pothole wetlands during drought and deluge: A 17-year study of the Cottonwood Lake wetland complex in North Dakota in the perspective of longer term measured and proxy hydrological records. *Climatic Change*, 40, 189–209.

Winter, T. C., & Wright, H. E. (1977). Paleohydrologic phenomena recorded by lake sediments. *Eos, Transactions American Geophysical Union*, 58(4), 188–196. https://doi.org/10.1029/EO058i004p00188

Wittrock, V., & Beaulieu, C. (2015). Climate reference station Saskatoon annual summary 2014. Saskatoon, Saskatchewan, Canada. WMO. (1972). Evaporation losses from storage gauges. In *Distribution of precipitation in mountainous areas. Geilo Symposium* (Norway, 31 July–5 August 1972), (vol. II, Technical Papers, WMO-No. 326). Geneva, CH: World Meteorological Organization, pp. 96–102.

Woo, M. K., & Rowell, R. D. (1993). Hydrology of a prairie slough. *Journal of Hydrology*, 146, 175–207. https://doi.org/10.1016/0022-1694(93)90275-E

Yu, G., Sauchyn, D., & Li, Y. F. (2013). Drought changes and the mechanism analysis for the North American Prairie. *Journal of Arid Land*, 5(1), 1–14. https://doi.org/10.1007/s11627-013-0136-4

Zeban, B. J., & De Jong, E. (1989). Water flow in a hummocky landscape in central Saskatchewan, Canada, II. Saturated flow and groundwater recharge. *Journal of Hydrology*, 110(1–2), 181–198. https://doi.org/10.1016/0022-1694(89)90243-6