Identification of groundwater discharge along the shoreline of large inland lakes in Southern Ontario

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Abstract

Groundwater discharge may be an important pathway for delivering pollutants into nearshore waters of large inland lakes, including the Great Lakes and Lake Simcoe in Southern Ontario, Canada. This pathway however is poorly understood and quantified. While field methods for evaluating groundwater discharge to surface waters in tributary and marine settings have been widely applied and are well tested, there are limited field methods available for evaluating groundwater discharge to large inland lakes, particularly at the regional-scale (i.e. 1-100 km).

The objective of this thesis was to evaluate suitable field methods for quantifying groundwater discharge into large inland lakes at different spatial scales, and to evaluate the spatial distribution and magnitude of groundwater discharge along shorelines in Nottawasaga Bay (Lake Huron) and Lake Simcoe. A combination of the field methods for quantifying groundwater discharge was first evaluated along a 17 km stretch of shoreline in Nottawasaga Bay near the Township of Tiny. Regional-scale radon-222 ($^{222}$Rn) and electrical resistivity tomography (ERT) boat surveys were conducted along the shoreline to identify potential groundwater discharge hotspots. From a management perspective, identification of groundwater discharge hotspots is needed so that water quality management efforts aimed at reducing groundwater pollution inputs can target these areas. Following the identification of a potential groundwater discharge hotspot area, a higher spatial resolution $^{222}$Rn survey was conducted in this area. Data from this survey indicated that groundwater discharge is the highest close to the shoreline with discharge decreasing offshore. A steady-state $^{222}$Rn mass balance model which considers the various sources and sinks of $^{222}$Rn from the coastal water column found groundwater discharge rates along the shoreline to range from $0.15 \pm 0.04$ - $5.11 \pm 1.23$ m$^3$ m$^{-1}$ d$^{-1}$. Six beach sites in the Nottawasaga Bay study area were characterized more closely with shore-normal transects of groundwater wells installed to determine the groundwater flux towards the lake. Detailed vertical temperature and hydraulic gradient profiles were also collected at select sites to characterize local-scale groundwater discharge patterns. While this work shows the successful application of $^{222}$Rn for evaluating nearshore groundwater discharge to large inland lakes, the use of local-scale methods including vertical temperature and hydraulic gradient methods was challenging due to shallow gravel-cobble sediment which prevented manual installation of equipment in the nearshore lake bed at many sites. Regional-scale $^{222}$Rn boat surveys were subsequently performed in Lake Simcoe to
identify groundwater discharge hotspots along 80 km of shoreline. Two potential groundwater discharge hotspot areas were identified, as well as two areas where indirect groundwater discharge (i.e. groundwater discharge to creeks which then flows into the lake) may affect the nearshore lake water quality. High spatial resolution surveys were conducted in the groundwater discharge hotspot areas with data indicating that groundwater discharge in these areas is higher near the shoreline and decreases offshore. Groundwater discharge rates in Lake Simcoe were estimated to range from $0.18 \pm 0.01 - 4.18 \pm 0.30$ m$^3$ m$^{-1}$ d$^{-1}$ from applying the steady-state $^{222}$Rn mass balance model. The $^{222}$Rn concentrations in the lake exhibited high temporal variability with preliminary analysis indicating this variability is due to varying wind speed and, to a lesser extent, precipitation. Better understanding of factors contributing to the temporal variability in $^{222}$Rn concentrations is needed for more accurate interpretation of the regional-scale $^{222}$Rn survey data. The combination of field methods evaluated in this thesis provides characterization of nearshore groundwater discharge to large inland lakes at multiple scales as required to develop more effective management plans to mitigate the contribution of groundwater pollutant inputs to nearshore waters.

Keywords

Groundwater discharge, $^{222}$Rn, Nottawasaga Bay, Lake Simcoe, Great Lakes, Field methods, Groundwater-surface water interactions
Co-Authorship Statement

The candidate is responsible for the collection and analysis of field data as well as writing all thesis chapters. Dr. Clare Robinson provided the initial motivation for this research, assisted with field work, and provided suggestions for data analysis and improvement of the thesis. The co-authorship breakdown of Chapters 3 and 4 are as follows:

Chapter 3: Multiple methods for characterizing groundwater discharge along a Lake Huron shoreline

Authors: Tao Ji, Richard Peterson, Kevin Befus, Clare Robinson

Contributions:

Tao Ji: Designed the field deployment plan, built and set up all field equipment except for the $^{222}\text{Rn}$ detection system and ERT survey equipment, and measured, analyzed and interpreted field data, performed lab experiment, and wrote the thesis chapter.

Richard Peterson: Together with Tao Ji, was responsible for conducting the $^{222}\text{Rn}$ survey, and provided advice for analysis of $^{222}\text{Rn}$ survey data.

Kevin Befus: Conducted the regional-scale ERT survey and ER inversion modeling.

Clare Robinson: Initiated research topic, aided in field investigations, provided advice for data analysis and revised the chapter draft.

Chapter 4: Using $^{222}\text{Rn}$ as a tracer to quantify groundwater discharge along the Lake Simcoe shoreline

Authors: Tao Ji, Clare Robinson

Tao Ji: Designed the field deployment plan, built field equipment and instrumentation, measured, analyzed and interpreted field data, performed laboratory experiment and wrote the thesis chapter.

Clare Robinson: Initiated research topic, aided in field investigations, provided advice for data analysis and revised the chapter draft.
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Chapter 1

1 Introduction

1.1 Research background

Groundwater accounts for over 30% of the world’s freshwater (Shlklomanov, 1993). Almost nine million Canadians (30.3 % of the population) and 45.8 % of the population in Ontario rely on groundwater for municipal, domestic and rural use (Environment Canada, 2016). Despite its abundance and significance, groundwater is increasingly threatened by contamination caused by anthropogenic activities (Kalbus et al., 2006, Kidmose et al., 2015). Groundwater and surface water are inextricably linked and any changes in groundwater resources (water quantity or quality) can lead to the deterioration of surface waters and the related ecosystems (Grannemann et al., 2000). In response to the degraded water quality in many large inland lakes including the Laurentine Great Lakes (herein called the Great Lakes), there is an increasing need to evaluate the contribution of groundwater discharge in delivering contaminants into the lake. This information is required for the development of more effective water quality management programs.

Groundwater discharge into lakes is commonly referred to as lacustrine groundwater discharge (LGD). LGD is defined as all water that flows across the sediment-water interface to a lake regardless of its origin (Hayashi and Rosenberry, 2002, Lewandowski et al., 2013, Meinikmann et al., 2013, Meinikmann et al., 2015). As such LGD is driven by the groundwater hydraulic gradient, as well as processes that drive water exchange across the sediment-water interface in a lake setting (e.g. waves, current-bedform interactions). Groundwater discharge can be a significant pathway for transporting dissolved pollutants such as nutrients (e.g. nitrogen [N] and phosphorus [P]), chlorides and organic contaminants to lakes (Moore, 1996, Burnett et al., 2006). Although the amount of groundwater discharge into a lake may only be a small component of the lake water balance, concentrations of pollutants can be much higher in groundwater than in the receiving surface water (Taniguchi et al., 2002, Burnett et al., 2006,
Kornelsen and Coulibaly, 2014). LGD has been shown to play an important role in geochemical cycling and ecosystem functioning in lakes (Hayashi and Rosenberry, 2002, Meinikmann et al., 2013). For instance, many studies have shown that discharge of nutrient-enriched groundwater can alter the nutrient budget of lakes, leading to serious lake eutrophication issues (Grannemann et al., 2000, Kornelsen and Coulibaly, 2014, Meinikmann et al., 2015). Nutrients-driven lake eutrophication problems such as harmful and nuisance algae blooms have raised considerable public awareness and concern recently (Shaw et al., 1990, Evans et al., 1996, Winter et al., 2007, North et al., 2013, Kidmose et al., 2015). Although the importance of groundwater discharge to lakes is now widely recognized, groundwater inputs are still poorly understood and quantified for most lakes especially for large inland lakes such as the Great Lakes and Lake Simcoe. Identifying and quantifying groundwater discharge and associated pollutant loading is complex and challenging due to the difficulties in measuring these unseen fluxes (Burnett et al., 2006).

Methods for assessing and quantifying groundwater discharge to oceans (submarine groundwater discharge, SGD) have been improved greatly over the last few decades. There are many methods that have been successfully used to evaluate SGD and its impact on marine coastal waters. Approaches include natural isotope tracers (e.g. radium, radon-222 [$^{222}$Rn], oxygen-18 [$^{18}$O] and carbon-14 [$^{14}$C]), seepage meters, water mass balance models, hydraulic gradient methods (piezometers), heat tracer techniques and geophysical techniques (electric resistivity) (Taniguchi et al., 2002, Anderson, 2005, Burnett et al., 2006, Burnett et al., 2008, Povinec et al., 2008, Dimova et al., 2013, Ono et al., 2013). While many of these methods have also been successfully applied to quantify groundwater discharge to small inland lakes, there are limited applications of these methods in large inland lake settings such as the Great Lakes and Lake Simcoe. There is a need to identify suitable approaches for quantifying groundwater discharge into large inland lakes so that the influence of groundwater discharge on lake water quality can be evaluated. Each method is suitable for different spatial and temporal scales. Therefore, to understand the temporal and spatial variability of groundwater discharge and reduce the uncertainties in measurement estimates, it is recommended that a combination of...
multiple methods be used in evaluating groundwater discharge (Burnett et al., 2006, Kalbus et al., 2006).

1.2 Research objective

This thesis is divided into three objectives. The first objective is to evaluate suitable field techniques for quantifying groundwater discharge into large inland lakes at different spatial scales (regional and local). The second objective is to evaluate the spatial patterns and quantities of groundwater discharge along shorelines of large inland lakes in Southern Ontario and link observed discharge groundwater patterns to hydrogeological characteristics of the nearshore area. The main field technique evaluated in this thesis for assessment of regional-scale groundwater discharge is the natural tracer $^{222}\text{Rn}$. The third objective, related to reducing uncertainty in this measurement techniques, is to evaluate the causes of temporal variability of $^{222}\text{Rn}$ concentrations in the lake water. Understanding large scale groundwater discharge patterns is the first critical step to better characterizing and quantifying groundwater as a potentially important non-point pollution source. The research presented in this thesis provides valuable information for water resource management in the study areas as well as methodologies that may be broadly applied to investigate groundwater discharge into large inland lakes.

1.3 Thesis outline

A concise description of the outline of this thesis is as follows:

Chapter 1: Introduction of the research background and research objectives.

Chapter 2: Literature review of previous work conducted to evaluate groundwater discharge into large inland lakes with a focus on applicable groundwater discharge measurement techniques and tools.

Chapter 3: Application of multiple field methods to quantify nearshore groundwater discharge along the eastern shore of Nottawasaga Bay. This chapter presents methods for identifying
shorelines areas with high groundwater discharge and shows how a combination of approaches can be used to estimate groundwater discharge rates and evaluate spatial variability in groundwater discharge.

Chapter 4: Application of multiple field methods to quantify nearshore groundwater discharge into Lake Simcoe. Field results are used to identify groundwater discharge hotspots and to evaluate factors controlling the temporal variability of $^{222}$Rn concentrations in the lake water.

Chapter 5: Summary of the research findings and recommendations for future work.
1.4 References


Chapter 2

2 Literature review

2.1 The importance of groundwater discharge to large inland lakes

Groundwater discharge to lakes (also called lacustrine groundwater discharge, LGD) has been shown to deteriorate water quality and ecosystem health in lakes (e.g. Moore, 1996, Moore, 2010, Schmidt et al., 2010, Smith and Swarzenski, 2012, Dimova et al., 2013). For instance, Kidmose et al. (2015) recently reported that 96% of total nitrogen (N) inputs to Lake Hampen in Denmark may be attributed to direct groundwater discharge and that these inputs may have caused changes in the benthic algae composition and the biodiversity at the sediment-water interface. Although magnitude of groundwater inputs are generally smaller than surface water (i.e. tributary) inputs, numerous studies have shown that groundwater discharge can be an important pathway for delivering pollutants to lakes particularly in areas where pollutant concentrations are elevated in aquifers compared to adjacent surface waters (Bottomley et al., 1984, Rosenberry et al., 2000, Sebestyen and Schneider, 2004, Lowry et al., 2007, Stets et al., 2010). Groundwater discharge to lakes occurs by 1) direct groundwater discharge whereby groundwater flows directly into lakes from the nearshore aquifer or offshore discharge points; or alternatively by 2) indirect groundwater discharge whereby groundwater discharges into tributaries which then flow into the lakes (Figure 2-1; Kalbus et al., 2006). This thesis focuses on evaluating direct groundwater discharge to nearshore waters in large inland lakes in Southern Ontario, specifically Nottawasaga Bay (Lake Huron) and Lake Simcoe (Figure 2-2).
Figure 2-1: Generalized direct and indirect groundwater flow systems in the Great Lakes Region (figure modified from Grannemann et al., 2000).

Figure 2-2: Map showing location of Nottawasaga Bay in Lake Huron and Lake Simcoe (figure modified from Northeast Michigan Lake Huron Watershed, 2014).
2.1.1 Groundwater discharge to the Laurentian Great Lakes

The Great Lakes which include Lake Huron, Lake Ontario, Lake Superior, Lake Michigan and Lake Erie, holds 18-20% of the world’s freshwater supplies. Aquifers in the Great Lakes Basin also contain large volumes of groundwater - approximately 4,200 km$^3$ - which is a major resource and an important link between the Great Lakes and their watersheds (Grannemann et al., 2000). Nearshore waters of the Great Lakes are of immense ecological, economical and recreational value (Austin et al., 2007). These areas, however, are being increasingly threatened by deteriorated water quality (Cherkauer and McKereghan, 1991, Haack et al., 2005). For instance, increasingly large harmful cyanobacterial blooms have been observed annually since 1995 in the western basin of Lake Erie (Bails et al., 2006, Rinta-Kanto et al., 2005). In 2014, these blooms caused the shutdown of the drinking water distribution system in the City of Toledo for successive days impacting over half a million people (Rinta-Kanto et al., 2005, Berman, 2014, Obenour et al., 2014, Steffen et al., 2014). While water quality management efforts have historically focused on mitigating point pollution sources, there is increasing recognition that non-point sources including groundwater discharge may play an important role in delivering pollutants to nearshore waters (Hartmann, 1990, Mitsch and Wang, 2000, International Joint Commission, 2012). Despite this increasing recognition, the magnitude of direct groundwater discharge and associated pollutant loading to nearshore areas of the Great Lakes remains poorly understood.

The magnitude of direct groundwater discharge into the Great Lakes is generally thought to be much smaller than surface water (tributary) inputs. Nevertheless, groundwater may be enriched with pollutants (e.g. nutrients, chloride, organic contaminants, metals), and therefore the groundwater discharge may be an important pathway for delivering pollutants to the Great Lakes (Grannemann et al., 2000, Coon and Sheets, 2006, Kornelsen and Coulibaly, 2014). For instance, shallow unconfined aquifers adjacent to the Great Lakes have a high susceptibility to contamination due to high populations residing in shoreline areas (e.g. septic systems, leaky sewers) as well as high intensity agriculture in these areas. Groundwater contaminants in nearshore aquifers may be delivered to the lake via direct groundwater discharge (Grannemann
et al., 2000, Haack et al., 2005). Studies have reported that non-point source chloride (Cl\(^-\)) and nitrate (NO\(_3^-\)) flow through the shallow aquifers into the Great Lakes (Hill, 1990, Boutt et al., 2001). For instance, Cherkauer et al. (1992) applied a two-dimensional finite-element transport model in Door Peninsula, Wisconsin and estimated that around 33% and 38% of the total Cl\(^-\) and NO\(_3^-\) that entered the surficial aquifer of the Green Bay Basin was transported into Lake Michigan via direct groundwater discharge.

Prior studies have attempted to quantify direct groundwater discharge to the Great Lakes with most of them using water budget or numerical modeling approaches. Further most prior studies have focused on Lake Michigan. Table 2-1 provides a summary of previous studies that have quantified direct groundwater discharge to the Great Lakes. Bergstrom and Hanson (1962) estimated groundwater discharge into Lake Michigan to be 22.7 m\(^3\) s\(^{-1}\) using a water budget method. Considering a more realistic thickness for sand and fine-grained aquifers, Cartwright et al. (1979) calculated the groundwater discharge rate to Lake Michigan as 189.7 m\(^3\) s\(^{-1}\). Grannemann and Weaver (1999) estimated that Lake Michigan has the largest amount of direct groundwater discharge (76.5 m\(^3\) s\(^{-1}\)) amongst all of the Great Lakes because it has the greatest area of sand and gravel aquifers near the shore. More recently, Feinstein et al. (2010) constructed a regional-scale groundwater flow model with which they estimated the direct groundwater discharge into Lake Michigan to be 9.61 m\(^3\) s\(^{-1}\). Despite efforts to quantify direct groundwater discharge rates using the above approaches, there is currently limited field data available to quantify estimates.

Quantifying groundwater discharge to surface water bodies is challenging because discharge typically exhibits high spatial and temporal variability (Dimova et al., 2015). Limited field methods are available for identifying areas of high direct groundwater discharge (herein called groundwater discharge hotspots) and quantifying groundwater discharge rates, particularly at the regional-scale (> 1 km), for large inland lakes. Developing suitable field methods for identifying direct groundwater discharge hotspots and quantifying groundwater discharge rates is urgently needed to develop more effective and targeted groundwater monitoring and
protection plans to manage the contribution of groundwater to degraded water quality in nearshore waters.

Table 2-1: Summary of studies of direct groundwater discharge to the Great Lakes.

<table>
<thead>
<tr>
<th>Location</th>
<th>Method</th>
<th>Groundwater discharge flux</th>
<th>Reference</th>
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<tr>
<td>Lake Huron and Lake Michigan</td>
<td>Water budget</td>
<td>350 (m$^3$ km$^{-1}$ d$^{-1}$)</td>
<td>Bergstrom and Hanson (1962)</td>
</tr>
<tr>
<td>Western Lake Michigan</td>
<td>Water budget</td>
<td>110 (m$^3$ km$^{-1}$ d$^{-1}$)</td>
<td>Skinner and Borman (1973)</td>
</tr>
<tr>
<td>Lake Michigan</td>
<td>Piezometers</td>
<td>8200 (m$^3$ km$^{-1}$ d$^{-1}$)</td>
<td>Cartwright et al. (1979)</td>
</tr>
<tr>
<td>Western Lake Michigan</td>
<td>Water table</td>
<td>580-880 (m$^3$ km$^{-1}$ d$^{-1}$)</td>
<td>Cherkauer and Hensel (1986)</td>
</tr>
<tr>
<td>Lake Michigan</td>
<td>Seepage meter</td>
<td>107-671 (m$^3$ km$^{-1}$ d$^{-1}$)</td>
<td>Cherkauer and McKereghan (1991)</td>
</tr>
<tr>
<td>Eastern Lake Michigan</td>
<td>Groundwater flow modelling</td>
<td>553 (m$^3$ km$^{-1}$ d$^{-1}$)</td>
<td>Sellinger (1995)</td>
</tr>
<tr>
<td>Western Lake Ontario</td>
<td>Water table</td>
<td>4423 (m$^3$ km$^{-1}$ d$^{-1}$)</td>
<td>Harvey et al. (2000)</td>
</tr>
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<td>Lake Michigan</td>
<td>Water budget</td>
<td>3.45×10$^4$ (m$^3$ km$^{-1}$ d$^{-1}$)</td>
<td>Grannemann et al. (2000)</td>
</tr>
<tr>
<td>Grand Traverse Bay (Lake Michigan)</td>
<td>Groundwater flow modelling and GIS</td>
<td>1000-2000 (m$^3$ km$^{-1}$ d$^{-1}$)</td>
<td>Boutt et al. (2001)</td>
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<td>Groundwater flow modelling</td>
<td>440 (m$^3$ km$^{-1}$ d$^{-1}$)</td>
<td>Hoaglund et al. (2002)</td>
</tr>
<tr>
<td>Northern Lake Michigan</td>
<td>Groundwater flow modelling</td>
<td>3456 (m$^3$ km$^{-1}$ d$^{-1}$)</td>
<td>Hoaglund et al. (2002)</td>
</tr>
<tr>
<td>Lake Huron</td>
<td>Water table</td>
<td>7.31×10$^{-2}$-8.31×10$^{-1}$ (m d$^{-1}$)</td>
<td>Crowe and Meek (2009)</td>
</tr>
<tr>
<td>Lake Michigan</td>
<td>Groundwater flow modelling</td>
<td>3.13×10$^4$ (m$^3$ km$^{-1}$ d$^{-1}$)</td>
<td>Feinstein et al. (2010)</td>
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</table>

2.1.2 Groundwater discharge to Lake Simcoe

Lake Simcoe is the largest inland lake in Southern Ontario aside from the Great Lakes. Lake Simcoe is very important for drinking water supply, fisheries as well as tourism and recreational activities (Palmer et al., 2011, North et al., 2013). The water quality and ecosystem in Lake Simcoe continues to be threatened by anthropogenic activities with current water
quality issues including elevated phosphorus (P) and Cl\(-\) levels (Evans et al., 1996, Eimers et al., 2005, Roy and Malenica, 2013). The decline of fish populations (e.g. whitefish and herring) in recent years has caused severe economic loss and raised considerable alarm regarding the water quality (Eimers et al., 2005). The Lake Simcoe Protection Plan (LSPP) was approved by the Federal Government in 2009 in recognition of the urgency to protect and restore the water quality and ecosystem health of Lake Simcoe (Ontario Ministry of the Environment, 2009).

Water quality management activities for Lake Simcoe including pollutant loading estimates have focused on tributary inputs. The magnitude of direct groundwater discharge to Lake Simcoe and the subsequent contribution of groundwater to pollutant loading is poorly understood. Lewis et al. (2007) applied seismo-stratigraphic techniques in Lake Simcoe and identified submarine hollows (potential locations for offshore groundwater discharge) in the floor of Kempenfelt Bay (located in the west of Lake Simcoe). Winter et al. (2007) estimated that septic systems may account for 5-7% of annual P inputs to Lake Simcoe, but the total P discharged from groundwater to the lake was not quantified. Roy and Malenica (2013) measured contaminant concentrations in the shallow groundwater along the shores of Kempenfelt Bay. They found high concentrations of contaminants (e.g. NO\(_3\)-, ammonium and chlorinated solvents) suggesting that urban groundwater may be delivering contaminants to the lake. North et al. (2013) observed large differences in P settling coefficients as well as O\(_2\) concentrations between Kempenfelt Bay and the main lake basin – these differences may be caused by direct groundwater discharge into Kempenfelt Bay. While these previous studies provide evidence of groundwater inputs to Lake Simcoe, field investigations have not been conducted to quantify groundwater inputs. More importantly, from a monitoring and management perspective, there is a need to identify areas that may be hotspots for direct groundwater discharge so that future groundwater management efforts can target these areas.
2.2 Methods to quantify groundwater discharge to large inland lakes

Quantifying the magnitude of groundwater discharge to surface waters is challenging due to low groundwater seepage rates combined with high spatial and temporal variability. Many field methods and tools have been developed and applied to quantify groundwater discharge to surface waters. These include naturally occurring isotopes (e.g., Radon-222 ($^{222}$Rn), radium isotopes, uranium isotopes, carbon-14 ($^{14}$C), tritium ($^{3}$H)), seepage meters, nested piezometers, groundwater monitoring wells, numerical groundwater modeling and heat tracer techniques (Shaw et al., 1990, Froehlich et al., 2005, Haack et al., 2005, Burnett et al., 2006, Kalbus et al., 2006, Dimova et al., 2013). Selection of an appropriate method (or suite of methods) needs to consider the spatial and temporal scale of interest and the advantages and limitations of each method. For example, while seepage meters and nested piezometers are useful simple techniques for quantifying local-scale (1 - 100 m of shoreline) groundwater discharge, these methods are not able to adequately characterize groundwater discharge over large areas due to the heterogeneous nature of groundwater discharge. Alternatively, regional-scale (1-100 km) groundwater discharge can be characterized by tracers including radium isotopes and $^{222}$Rn, but these methods integrate discharge rates over large areas and give limited local-scale understanding (Loaiciga and Zektser, 2003, Burnett et al., 2006, Kalbus et al., 2006, Dulaiova et al., 2010). Often, a combination of methods that are able to characterize groundwater discharge at different scales is recommended (Burnett et al., 2006, Kalbus et al., 2006). The objective of this thesis is to both evaluate the magnitude of direct groundwater discharge to nearshore areas of the Great Lakes and Lake Simcoe and moreover to assess the suitability of methods for quantifying groundwater discharge into large inland lakes. The following sections provide a review of field methods that were used to evaluate direct groundwater discharge to Nottawasaga Bay and Lake Simcoe in this thesis.
2.2.1 $^{222}$Rn as a tracer

$^{222}$Rn is a naturally occurring isotope that has been widely used as a tracer to assess groundwater discharge into the ocean, small inland lakes and streams (Corbett et al., 1997, Burnett et al., 2001, Burnett and Dulaiova, 2003, Burnett et al., 2006, Kluge et al., 2007, Dimova et al., 2009, Dimova and Burnett, 2011, Ono et al., 2013, Dimova et al., 2015). Generally, for a natural tracer to be suitable for evaluating groundwater discharge: 1) the tracer must be conservative; 2) the concentration of the tracer in groundwater should be higher relative to its concentration in surface water; 3) measurement of the tracer should be relatively straightforward (Moore, 1996, Cable et al., 1996, Povinec et al., 2008). $^{222}$Rn is a nuclide of the $^{238}$U decay series and a daughter nuclide of $^{226}$Ra. $^{222}$Rn is primarily produced by $^{226}$Ra decay and is delivered to surface waters by sediment diffusion and groundwater discharge (Swarzenski, 2007, Charette et al., 2008). The half-life of $^{222}$Rn is 3.8 days, so it is suitable for studying nearshore groundwater discharge as nearshore processes often have a similar time scale as the half-life of $^{222}$Rn (Burnett et al., 2001, Burnett et al., 2007, Charette et al., 2008, Dimova et al., 2013). $^{222}$Rn is a conservative gas that typically has a higher concentration in groundwater than in surface water (Burnett and Dulaiova, 2006, Povinec et al., 2012). Previous studies have shown that $^{222}$Rn is a useful tracer for evaluating groundwater discharge at both local (0.1-1 km of shoreline) and regional (1-100 km of shoreline) spatial scales (Mulligan and Charette, 2006, Dulaiova et al., 2010, Smith, 2012). Automated and continuous measurements of $^{222}$Rn in surface water can be performed using portable RAD7 (Durridge Co., Inc.) monitoring units and these commercial units have been used in studies of direct groundwater discharge into coastal areas around the world (Burnett et al., 2001, Dulaiova et al., 2005).

A steady-state $^{222}$Rn mass balance model (Figure 2-3) which considers the various sources and sinks of $^{222}$Rn in water column inventory is often adopted to estimate groundwater discharge rates from the $^{222}$Rn measurements (Cable et al., 1996, Burnett et al., 2001, Schmidt et al., 2010, Smith and Swarzenski, 2012). The mass balance is given as:

$$0 = J_{diff} + J_{gw} - J_{atm} - J_{mix} + [z(\lambda_{Rn} C_{Ra} - \lambda_{Rn} C_{w})]$$

(1)
where $z$ is the average depth of the water column (m); $C_w$ is the measured $^{222}$Rn concentration in the surface water (dpm L$^{-1}$); $J_{\text{mix}}$ is the loss of $^{222}$Rn in the water column due to offshore mixing (dpm m$^{-2}$ d$^{-1}$); $J_{\text{atm}}$ is the loss of $^{222}$Rn due to atmospheric evasion (dpm m$^{-2}$ d$^{-1}$); $J_{\text{diff}}$ is the $^{222}$Rn diffusion from sediment (dpm m$^{-2}$ d$^{-1}$); $J_{gw}$ is the $^{222}$Rn delivered to the surface water by groundwater discharge (dpm m$^{-2}$ d$^{-1}$); $\lambda_{\text{Ra}}C_{\text{Ra}}$ is the depth-integrated $^{222}$Rn production from $^{226}$Ra (dpm m$^{-2}$ d$^{-1}$); $\lambda_{\text{Rn}}C_{\text{Rn}}$ is depth-integrated in situ decay of $^{222}$Rn based on its half-life (dpm m$^{-2}$ d$^{-1}$). Multiplying $z$ by $\lambda_{\text{Rn}}C_{\text{Ra}}$ and $\lambda_{\text{Rn}}C_{\text{Rn}}$ yields $^{222}$Rn production from $^{226}$Ra ($J_{\text{prod}}$, dpm m$^{-2}$ d$^{-1}$) and the loss of $^{222}$Rn through decay ($J_{\text{decay}}$, dpm m$^{-2}$ d$^{-1}$), respectively.

Figure 2-3: $^{222}$Rn mass balance model for estimating groundwater discharge (modified from Burnett and Dulaiova, 2003). The sources of $^{222}$Rn to the water column include groundwater discharge ($J_{gw}$); diffusive flux of $^{222}$Rn from sediments ($J_{\text{diff}}$) and $^{222}$Rn production from $^{226}$Ra ($J_{\text{prod}}$). The losses include mixing with offshore waters ($J_{\text{mix}}$); atmospheric evasion ($J_{\text{atm}}$) and in-situ decay of $^{222}$Rn ($J_{\text{decay}}$).

Ellins et al. (1990) first used a $^{222}$Rn mass balance model to calculate groundwater discharge into surface waters in Puerto Rico. Cable et al. (1996) later constructed a linked benthic exchange-horizontal transport model which was the first time that $^{222}$Rn benthic flux was used to quantify submarine groundwater discharge into the northeastern Gulf of Mexico. Corbett et al. (1997) applied $^{222}$Rn to trace groundwater into Par Pond in South Carolina, and
demonstrated that groundwater discharge was an important component in the lake water budget (accounting for 10%-33% of total input). Dimova et al. (2013) also more recently used the $^{222}$Rn steady-state model to quantify groundwater discharge into small lakes in Florida. Although $^{222}$Rn and application of the steady-state $^{222}$Rn mass balance model has been widely applied to estimate groundwater discharge rates, there are still considerable uncertainties and limitations associated with its application (Burnett et al., 2007). The uncertainties in groundwater discharge rate calculations are propagated errors associated with all sources and sink terms in the $^{222}$Rn mass balance model (Figure 2-3).

Quantifying the $^{222}$Rn concentration of the groundwater end-member ($C_{gw}$) represents a major uncertainty in using the $^{222}$Rn mass balance model to estimate the specific groundwater flux ($q_{gd}$, m d$^{-1}$). In applying the mass balance model, $q_{gd}$ is determined by dividing the estimated $^{222}$Rn groundwater flux ($J_{gw}$, dpm m$^{-2}$ d$^{-1}$) by the $^{222}$Rn concentration of the groundwater end-member ($C_{gw}$, dpm L$^{-1}$) (Burnett and Dulaiova, 2003, Smith and Swarzenski, 2012):

$$ q_{gd} = \frac{J_{gw}}{C_{gw}} $$

(2)

Natural geological heterogeneities result in spatially variable concentrations of $^{222}$Rn in the groundwater end-member and therefore assigning one representative value for $C_{gw}$ is challenging (Dulaiova et al., 2008). Dimova et al. (2013) showed that the uncertainties in $q_{gd}$ due to estimation of $C_{gw}$ were more than 50%. Corbett et al. (2000) found that $^{222}$Rn concentrations were generally higher in deeper sediments compared with surficial sediments as $^{222}$Rn in surficial sediment may escape to the atmosphere. An alternative method used to determine a representative groundwater end-member concentration is performing sediment equilibrium experiment whereby surface bed sediment is placed in a sealed chamber with overlying surface water for 21 days to allow $^{222}$Rn and its parent, $^{226}$Ra, to reach equilibrium (Corbett et al., 1998, Santos et al., 2009, Kranrod et al., 2015). Gonneea et al. (2008), on the other hand, showed $^{226}$Ra which is the source of $^{222}$Rn in groundwater is strongly related to the presence of manganese (Mn) and iron (Fe) (hydr)oxides solid phases. The abundance of these
(hydr)oxides and thus $^{226}$Ra and $^{222}$Rn are strongly controlled by the subsurface geochemical conditions (in particular Eh and pH).

Another challenge in applying the $^{222}$Rn mass balance model is estimating and reducing the uncertainties associated with $^{222}$Rn loss by atmospheric evasion ($J_{atm}$) (Burnett and Dulaiova, 2006, Dulaiova and Burnett, 2006, Burnett et al., 2007). There are various methods for quantifying $^{222}$Rn loss due to atmospheric evasion but often empirical equations are used to calculate this flux (MacIntyre et al., 1995, Dulaiova and Burnett, 2006, Dimova et al., 2013):

$$J_{atm} = k(C_w - \alpha C_{air})$$

(3)

where $C_{air}$ is the measured $^{222}$Rn activities in atmosphere. $k$ is the gas-transfer coefficient (m h$^{-1}$) and $\alpha$ is the partitioning coefficient of $^{222}$Rn between water and air (dimensionless) given by:

$$k(600) = 0.45 \times \overline{u}_{10}^{1.6} \times (S\ell/600)^{-0.5}$$

(4)

$$\alpha = 0.105 + 0.405\exp(-0.05027T)$$

(5)

where $S\ell$ is the Schmidt number; $\overline{u}_{10}$ is the wind speed at a 10 m height above the water surface (km h$^{-1}$) and $T$ is the temperature at the water-air interface ($^\circ$C). $^{222}$Rn evasion to the atmosphere and therefore the $^{222}$Rn inventory in surface water is influenced by various factors including wind speed, water temperature and currents. For example, Burnett and Dulaiova (2006) observed $^{222}$Rn inventories in coastal waters of Donnalucata, Italy to change in response to high winds (10 m s$^{-1}$). In a case study in Dor Beach, Israel, $^{222}$Rn inventories considerably decreased during a storm (Burnett et al., 2007). Therefore, uncertainty in quantifying $^{222}$Rn losses to the atmosphere can cause difficulties in applying the steady-state mass balance model under some conditions (e.g. large winds, high precipitation and waves).

While $^{222}$Rn has been successfully applied for quantifying groundwater discharge into the ocean and small freshwater lakes, its suitability to measuring groundwater discharge to the large inland lakes such as the Great Lakes and Lake Simcoe is unclear. This research will
evaluate the use of $^{222}\text{Rn}$ as a tracer to quantify groundwater discharge to large inland lakes including assessment of the advantages and shortcomings of this measurement approach for this setting.

2.2.2 Groundwater hydraulic gradient

The groundwater hydraulic gradient is the driving force for groundwater discharge to surface waters, as groundwater flows in the direction of decreasing hydraulic gradient. Using Darcy’s Law (Eqn. (6)), specific groundwater flux ($q_{gd}$, m d$^{-1}$) can be calculated by (Darcy, 1856):

$$q_{gd} = -K \frac{dh}{dl}$$

(6)

where $K$ is the hydraulic conductivity of the porous media (m d$^{-1}$); $h$ is the hydraulic head (m); $L$ is the distance between hydraulic head measurements (m); and $\frac{dh}{dl}$ is the hydraulic gradient.

Groundwater monitoring wells and multi-level nested piezometers are generally used to measure hydraulic heads (Freeze and Cherry, 1979). For calculating horizontal groundwater flow, the hydraulic gradient is the difference of hydraulic head between monitoring wells spaced at a known distance, $L$. For calculating vertical groundwater flow, the hydraulic gradient is the difference of hydraulic head in piezometers with openings at depths spaced at a known distance, $L$ (Kalbus et al., 2006). $K$ can be determined by various methods such as grain size analysis, permeameter tests, slug and bail tests as well as pumping tests (Hazen, 1892, Cooper et al., 1967, Freeze and Cherry, 1979, Shepherd, 1989, Kelly and Murdoch, 2003, Kalbus et al., 2006).

*Horizontal hydraulic gradient measurements:* Measurement of groundwater heads adjacent to surface water bodies is the most common approach for estimating groundwater discharge rates to surface waters (Turner, 1998, Sophocleous, 2002, Gibbes et al., 2007). Horizontal hydraulic gradients can be calculated from groundwater levels measured in monitoring wells located at known locations from a surface water body. The equipment (groundwater monitoring wells) is easy to install particularly in permeable nearshore sediments and the calculation is
straightforward (Kalbus et al., 2006). This method provides point hydraulic gradient and groundwater discharge estimates along a shoreline making this method appropriate for small-scale studies of groundwater discharge conditions along a shoreline (Kalbus et al., 2006, Meinikmann et al., 2013). The largest challenge in using hydraulic gradient measurements to calculate the groundwater discharge is the accurate determination of $K$ (Eqn. (6)) (Mulligan and Charette, 2006). $K$ estimates often range by orders of magnitude depending on the method used to quantify it. $K$ is also highly spatially variable and can vary several orders of magnitude over small distances (Devlin and McElwee, 2007). Further, it is important to note that this measurement technique estimates only the terrestrial (inland) groundwater discharge whereas other measurement techniques (e.g. heat tracer techniques, $^{222}$Rn, vertical hydraulic gradients) also include water that is recirculating across the sediment-water interface in the estimated groundwater discharge rate.

*Vertical hydraulic gradient measurements:* Groundwater discharge to a surface water body may be determined by measuring vertical hydraulic gradients directly below the sediment-water interface (Cherkauer and McKereghan, 1991, Harvey et al., 2000). Multi-level piezometers are typically used to measure vertical hydraulic gradients with different techniques used to measure the hydraulic heads. Pressure transducers may be installed in each piezometer, or alternatively nested mini-piezometers may be directly attached to a differential manometer thereby measuring the vertical water pressure difference only (Cey et al., 1998, Kelly and Murdoch, 2003, Anderson, 2005). Oil-water rather than air-water differential manometers can be used as the head difference that can be read off the manometer board is amplified when using an oil-water manometer.

Similar to the horizontal hydraulic gradient measurement, the vertical hydraulic gradient method is well established, the equipment is easy to install in the field and the data analysis is straightforward. This method however provides localized point estimates of $q_{gd}$ and therefore this method is not suitable for quantifying regional-scale groundwater discharge. A large number of nested piezometers needs to be installed to accurately determine groundwater discharge rates along even a small length of shoreline ($10 – 100$ m). The method is useful for
assessing localized spatial heterogeneity in discharge rates. It is important to note that $q_{gd}$ calculated using this method includes both terrestrial (inland) groundwater discharge and any water that is recirculating across the sediment-water interface. This can lead to large temporal variability in $q_{gd}$ particularly when there is high wind and wave activity near the shoreline.

Studies often combine both horizontal and vertical hydraulic gradient measurements with other field methods to calculate the groundwater discharge. For instance, Rosenberry et al. (2008) used groundwater monitoring wells, nested piezometers as well as seepage meters to quantify the groundwater discharge to a small lake and recommended that using more than one method to quantify groundwater discharge increases the confidence in the estimated values. Kishel and Gerla (2002) applied the horizontal hydraulic gradient method together with temperature and stratigraphy data to characterize small-scale groundwater flow patterns into Shingobee Lake, USA. Meinikmann et al. (2013) more recently used water balance calculations to estimate the total groundwater nutrient inputs into Lake Arendsee in Northeastern Germany and evaluated the spatial variability of groundwater discharge using horizontal hydraulic gradient methods.

2.2.3 Heat as a tracer

Heat (temperature) can be used as a tracer for quantifying groundwater discharge to surface waters (Taniguchi, 2000, Taniguchi et al., 2003, Anderson, 2005, Schmidt et al., 2007, Rau et al., 2014). Heat tracer methods rely on differences in temperature between the groundwater and surface water with, for example, temperature depth profiles indicating gaining or losing conditions in a stream (Stonestrom and Constantz, 2003). While the groundwater temperature in Great Lakes Basin is relatively constant (7-12°C) year-round, the surface water temperature varies considerably (0-23°C) (Grannemann et al., 2000). Many methods are available that use heat as a tracer to estimate groundwater discharge rates including infrared thermal imagery, distributed temperature sensing (DTS), landsat thermal imaginary and vertical temperature profiling (Duarte et al., 2006, Anibas et al., 2009, Briggs et al., 2012, Lewandowski et al., 2013). Infrared thermal imagery, distributed temperature sensing (DTS) and landsat thermal imaginary methods have improved greatly in the last few years and these methods are now able
to rapidly map the temperature distributions for large areas (Duarte et al., 2006, Lowry et al., 2007, Briggs et al., 2012). These methods, however, generally require expensive equipment or advanced computational processing (Duarte et al., 2006, Rau et al., 2010, Lewandowski et al., 2013). In this research, groundwater discharge was estimated by using relatively inexpensive vertical temperature sticks to measure the vertical temperature profile below the sediment-water interface (Figure 2-4).

![Image](image-url)

**Figure 2-4:** Schematic of “temperature stick”: six thermocouples attached to a metal stake and a datalogger to measure vertical temperature profiles below the sediment-water interface. One extra thermocouple was used to measure the surface water temperature.

The vertical temperature profiling approach is based on the theory that groundwater flow influences the subsurface heat distribution as heat is transported both by heat conduction and advection (Bredehoeft and Papaopulos, 1965, Taniguchi et al., 2003, Burnett et al., 2006, Anderson, 2005). The governing equation for heat transport equation is given as (Domenico and Palciauskas, 1973, Domenico and Schwartz, 1998):

\[
\frac{\kappa_e}{\rho_\phi \phi} \nabla^2 T - \frac{\rho_w c_w}{\rho_\phi \phi} \nabla \cdot \left( T \frac{q_{gd}}{\phi} \right) = \frac{\partial T}{\partial t} 
\]

where \( T \) is temperature (°C) at any point at time \( t \) (s); \( \kappa_e \) is the thermal conductivity of solid-fluid matrix (J s\(^{-1}\) m\(^{-1}\) K\(^{-1}\)); \( \phi \) is the porosity of the sediment; \( q_{gd} \) is the vertical specific
groundwater flux (m s\(^{-1}\)); \(\rho_w\) and \(\rho_0\) are density of the fluid and the solid-fluid matrix (kg m\(^{-3}\)), respectively; \(c_w\) and \(c_0\) are specific heat of the fluid and the solid-fluid matrix (J kg\(^{-1}\) K\(^{-1}\)), respectively. Often eqn. (7) is simplified by assuming steady-state one-dimensional (vertical) groundwater flow through isotropic, homogenous saturated porous medium. Under these conditions, the equation is given as (Bredehoeft and Papaopulos, 1965):

\[
\frac{\partial^2 T}{\partial d^2} - \frac{\rho_w c_w q_{gd}}{\rho k_e} \frac{\partial T}{\partial d} = 0
\]  

(8)

where \(d\) is vertical depth beneath the sediment-water interface (cm). Measured vertical temperature profiles are often used together with the solution to eqn. (8) to estimate \(q_{gd}\) at a specific point. A type curve method was developed by Bredehoeft and Papaopulos (1965) to convert measured vertical temperature profiles to \(q_{gd}\).

Vertical temperature profiling beneath the sediment-water interface has been used extensively to calculate groundwater discharge to streams and small inland lakes (Kalbus et al., 2006, Stonestrom and Constantz, 2003). For example, Lapham (1989) used monthly and yearly temperature variations (25°C for stream water and 1.5°C for groundwater) to determine local groundwater fluxes to streams as well as the effective hydraulic conductivities of stream bed sediments. Schmidt et al. (2007) measured the temperature at a uniform depth along the Pine River River, Ontario and mapped the plan-view streambed temperature distribution to delineate the groundwater discharge zones. They calculated \(q_{gd}\) using the one-dimensional heat transport eqn. (8). Anibas et al. (2011) also used vertical temperature profiling to evaluate the temporal and spatial patterns of groundwater discharge into a river in Belgium. To analyze and interpret the field data, groundwater flow and heat transport modeling tools such as VS2DHI can be used to simulate measured temperature profiles and groundwater discharge conditions (Healy and Ronan, 1996).

Using heat as a tracer to estimate groundwater discharge is expanding in popularity as it is a relatively robust, quick and inexpensive measurement technique (Kalbus et al., 2006). Another advantage of using heat tracer approaches is that the thermal conductivity of sediment (\(k_e\)) varies over less orders of magnitude compared with \(K\) values and therefore it can be easier
to constrain this parameter value (Domenico and Schwartz, 1998, Anibas et al., 2011, Rau et al., 2014). However, using vertical temperature profiling to estimate $q_{gd}$ has its limitations. This approach requires that the temperature is sufficiently different between groundwater and surface water - this occurs only in specific seasons (e.g. in Southern Ontario groundwater is generally colder than surface water in summer, and warmer than surface water in winter). Moreover, this method only measures groundwater temperature as a point location and therefore only provides localized $q_{gd}$ estimates. Further, the vertical temperature profiling results are influenced by diurnal fluctuations in solar radiation, wind-induced waves and lake mixing - these factors can result in error in applying the steady-state heat transport equation to infer $q_{gd}$ (Rau et al., 2014). Therefore, vertical temperature surveys are constrained by the weather and should be conducted on cloudy days with calm water conditions.

2.2.4 Electrical Resistivity Tomography (ERT)

Electrical resistivity tomography (ERT) has proven to be a useful approach for providing insight into the spatial variability in groundwater discharge estimates (Kemna et al., 2002, Muchingami et al., 2012, Johnson et al., 2015). ERT surveys are used to determine spatial variability in the electrical resistivity (ER) of sediments where the ER ($\rho$, $\Omega$ m) is closely linked with the hydrogeological properties such as porosity, permeability and the fluid conductivity (Eqn. (9)) (Daily and Owen, 1991, Manheim et al., 2004, Swarzenski et al., 2006, Moore, 2010, Muchingami et al., 2012). In a saturated porous media, bulk ER ($\rho_b$) of the fluid is related to porosity ($\emptyset$), cementation factor of the sediment ($m$) and the fluid resistivity ($\rho_f$) by an empirical model called Archie’s Law (Archie, 1942):

$$\rho_b = \rho_f \emptyset^{-m}$$  \hspace{1cm} (9)

Electrical conductivity ($\sigma$, S m$^{-1}$) is the inverse of electrical resistivity. As electricity can be conducted by ionic transport through the saturated sediment, different types of aquifers have their own unique resistivity. For instance, clays generally show a low resistivity due to their large porosity (Johnson et al., 2015). As a result, ERT is often used to support geological mapping in complex environments such as marine and riverine environments. On the other
hand, groundwater is often more conductive than surface water as it has more dissolved salts and total dissolved solid (TDS). As such, ER has been used extensively for tracking groundwater, saline water as well as solute transport in aquifers (Kemna et al., 2002, Manheim et al., 2004, Befus et al., 2013, Befus et al., 2014).

When electrical current is injected into the ground through electrodes, the spatial distribution of electrical field is measured. There are two main ways to conduct ERT surveys for the purposes of better understanding groundwater-surface water interactions: one is the land-based static ERT surveys and the other is continuous offshore ERT surveys (Swarzenski and Izbicki, 2009). Studies including Daily et al. (1992) and Zarroca et al. (2011) describe methods for land-based ERT surveys, and these methods have been applied extensively to obtain high-resolution two-dimensional ERT images to understand subsurface geological properties (Griffiths and Barker, 1993). Land-based ERT surveys have also been widely used to identify the fresh groundwater and salt water interface in permeable marine coastal aquifers as salt water has a lower resistivity than freshwater (e.g. Hoefel and Evans, 2001, Swarzenski et al., 2006, Zarroca et al., 2011, Johnson et al., 2015).

Evans et al. (1999) used an ERT system towed behind a boat with cables dragged on the seafloor of Humboldt Bay, California to map the ER profiles. Snyder and Wightman (2002) conducted continuous ERT surveys in Ohio River to characterize the geological properties of the river bottom. Manheim et al. (2004) later conducted ERT surveys in the coastal bays of the Delmarva Peninsula, USA to identify fresh groundwater discharge and explain how the hydrogeology controls the groundwater discharge phenomena. Slater et al. (2010) combined an offshore ERT survey with fiber-optic distributed temperature sensing methods for interpreting the groundwater-surface water interactions and the timing of groundwater discharge to Columbia River, USA.

ERT survey is a rapid and efficient tool for improving understanding of coastal hydrogeology. They can provide detailed high-resolution information on the hydrogeological and geological properties of the subsurface environment including the lithology, saturation and porosity. This method is also generally less expensive, less susceptible to environmental noise, and has a
faster turn-around of information compared to other geophysical methods (Rucker et al., 2011). However, some factors limit the use of ERT methods. These include the long preparation time for establishing the electrode contact and also the complexity of using inversion modeling to convert ER plots to information on the hydrogeology and geology. Ground-truthing of the inferred information from the geophysical surveys is always required to guarantee the quality of the inversion models (Day-Lewis et al., 2005).

In the research presented in Chapter 3, resistivity cables were towed behind a boat to conduct offshore ERT surveys. These results were used together with offshore $^{222}$Rn survey measurements to better understand the offshore surficial geological controls associated with the identified direct groundwater discharge hotspot areas.

### 2.3 References


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Chapter 3

3 Multiple methods for characterizing groundwater discharge along a Lake Huron shoreline

3.1 Introduction

Nearshore water quality in the Laurentian Great Lakes is degraded due to pollutant inputs associated with urbanization, agriculture and industrial activities in the Great Lakes Basin (Environment Canada and U.S. Environmental Protection Agency, 2011). Water quality is continuing to degrade, for example in the western basin of Lake Erie where there has been an increasing proliferation of algal blooms in recent years (Rinta-Kanto et al., 2005, Bails et al., 2006). While water quality management efforts for the Great Lakes historically focused on identifying and controlling point pollution sources, it is now widely acknowledged that non-point pollution sources including groundwater discharge may be important contributors to pollutant loading (Hartmann, 1990, Mitsch and Wang, 2000, International Joint Commission, 2012). While it is generally thought that the magnitude of direct groundwater discharge to the Great Lakes is relatively small compared to tributary (river and stream) inputs, groundwater discharge may be an important pathway for delivering pollutants to lakes particularly in areas where pollutant concentrations are elevated in aquifers compared to receiving lake waters (Grannemann et al., 2000, Coon and Sheets, 2006, Kornelsen and Coulibaly, 2014). For instance, surficial unconfined aquifers adjacent to the Great Lakes have a high susceptibility to contamination due to, for example, activities and infrastructure associated with high populations residing in shoreline areas (e.g. septic systems, leaky sewers). Despite recognition of its potential importance, the magnitude of direct groundwater discharge and associated pollutant loading to the Great Lakes is not well understood.

A wide range of field methods have been developed, with many now routinely applied, to quantify groundwater discharge to marine coastal waters as well as to rivers and streams. Methods include naturally occurring isotopes (e.g. Radon-222 ($^{222}$Rn), radium isotopes,
uranium isotopes, carbon-14 (\(^{14}\)C), tritium (\(^{3}\)H)), seepage meters, hydraulic gradient methods, numerical groundwater modeling, electrical resistivity tomography (ERT) imaging, and heat tracer techniques (Shaw et al., 1990, Froehlich et al., 2005, Haack et al., 2005, Burnett et al., 2006, Kalbus et al., 2006, Dimova et al., 2013, Zarroca et al., 2014). Which method, or combination of methods, is the most appropriate to use depends on the specific groundwater-surface water interaction setting being studied (i.e. ocean vs. river, nearshore vs. offshore discharge, preferential vs. diffuse discharge). Most prior field studies have quantified direct groundwater discharge to the Great Lakes using local-scale measurement techniques such as seepage meters, nested piezometers, and groundwater monitoring wells (e.g. Cherkauer and Hensel, 1986, Cherkauer and Taylor, 1990, Harvey et al., 2000). As groundwater discharge can be highly spatially heterogeneous, field results using these techniques are not necessarily representative of groundwater discharge over a larger area (Cherkauer and McKereghan, 1991, Haack et al., 2005, Delin et al., 2007). At the regional-scale, water budget and numerical groundwater flow modelling have been used to estimate direct groundwater discharge to the Great Lakes (Bergstrom and Hanson, 1962, Sellinger, 1995, Grannemann et al., 2000, Feinstein et al., 2010). While these regional-scale methods provide valuable estimates of direct groundwater discharge, they generally require large amounts of input data, are often associated with large uncertainty, and are rarely validated with field data (Sellinger, 1995, Grannemann et al., 2000, Hoaglund et al., 2002). The suitability of the different available methods for quantifying direct groundwater discharge to large inland waters such as the Great Lakes, particularly at the regional-scale is not clear. Reliable methods to estimate of direct groundwater discharge as well as to identify spatial areas where groundwater discharge may be higher is needed to effectively manage direct groundwater inputs to the lakes. For instance, efforts to manage the contribution of groundwater and associated pollutant inputs to the lakes need to target areas where direct groundwater discharge is high and thus groundwater is more likely to deliver pollutants the surface waters.

This Chapter aims to evaluate the suitability of a combination of methods for quantifying direct groundwater discharge to nearshore waters of the Great Lakes. Based on a broad review of potential methods that may be suitable for estimating direct groundwater discharge to the lakes,
we specifically examine the suitability as well as compare estimates obtained using regional-
scale methods including $^{222}$Rn and electrical resistivity (ER) imaging, and local-scale methods
including nested mini-piezometers, groundwater monitoring wells and vertical temperature
profiling. These methods were applied along a 17 km stretch of permeable shoreline in
Nottawasaga Bay, Lake Huron (Figure 3-1).

3.2 Study site description

This study was conducted along 17 km of shoreline in eastern Nottawasaga Bay near the
Township of Tiny in Simcoe County. The study area extended from Mountain View Beach
(MVB) to Sand Castle Beach (Figure 3-1). Nottawasaga Bay is located in South Georgian Bay
which is a part of Lake Huron. The water quality and ecosystem health in Nottawasaga Bay is
threatened by anthropogenic stressors including high inputs of phosphorus (P) which lead to
nuisance and harmful algae blooms (Golder Associates Ltd., 2005a, Environment Canada,
2015b). Elevated nitrate (> 8 mg NO$_3$-N L$^{-1}$) concentrations in the groundwater have also been
observed in aquifers discharging to the lake in the study area (Golder Associates Ltd., 2005a).

Regionally, the study area is located within the Simcoe Uplands and Simcoe Lowland
physiographic regions (Chapman and Putnam, 1984). The physiography is characterized by
rolling till plains that slope steeply and gently depending on the location towards Georgian Bay
(Golder Associates Ltd., 2014). The Lowland area along Nottawasaga Bay is characterized as
a sand plain with largely coarse textured glaciolacustrine surficial sediments (sand and gravel,
as well as sand and silt) (Ontario Ministry of Northern Development and Mines, 2016). The
dominant land use along the coastal corridor adjacent to the shoreline (100-1000 m wide) is
residential with many summer cottages being converted to larger permanent dwellings in recent
years. Further landward and in the Upland regions the dominant land cover is rural pasture,
cropland and forest (Golder Associates Ltd., 2014). Regionally, the soils, including the soils in
the upland recharge area, are generally well drained with high potential for infiltration and thus
aquifer recharge (Golder Associates Ltd., 2014). The average precipitation in the area is around
1000 mm yr$^{-1}$ with the recharge rates estimated to range from 255 to 396 mm y$^{-1}$ (Golder
The regional hydrogeology consists of a series of continuous and discontinuous aquifer and confining units that are quite variable in composition and thickness depending on the location. Regional piezometric surface mapping indicates that Nottawasaga Bay is a major groundwater discharge feature in the region (Golder Associates Ltd., 2014). Geological mapping has identified a series of tunnel valley aquifers (sand and gravel) running north-south towards Nottawasaga Bay with intervalley areas containing up to five regional aquifers at varying depths (Golder Associates Ltd., 2014). A surficial unconfined (sand/gravel) aquifer unit is present across much of the Lowlands area with thickness up to 50 m thick but thinning towards Nottawasaga Bay. The surficial aquifer is generally exposed but in some areas it is capped by a surficial clay/silt confining unit. The surficial aquifer outcrops along the shoreline as sand/gravel beaches with drainage from this aquifer contributing to nearshore groundwater discharge to Nottawasaga Bay.

While there are no major tributaries discharging to Nottawasaga Bay along the studied shoreline, Lafontaine Creek with a mean flow rate of around 3000 m$^3$ d$^{-1}$ discharges approximately 400 m south of Wahnekewaning Beach (WB, shown in Figure 3-1) (Golder Associates Ltd., 2013). The water level in Nottawasaga Bay is relatively stable but it does fluctuate seasonally (typically highest after spring run-off) and the shoreline is exposed to wind waves generated by high offshore winds.

Six beaches in the study area were selected for detailed local-scale groundwater discharge measurements: Mountain View Beach (MVB, 44°40’20.74”N, 79°58’57.50”W), Balm Beach (BB, 44°40’59.86”N, 79°59’40.88”W), beach at the end of Concession Road 12 (C12, 44°42’46.93”N, 80°01’25.23”W), Wahnekewaning Beach (WB, 44°43’23.06”N, 80°02’00.63”W), beach at the end of Concession Road 15 (C15, 44°44’00.70”N, 80°04’40.00”W) and Lafontaine Beach (LB, 44°44’29.81”N, 80°05’22.63”W). The nearshore surficial aquifer from LB to C12 consists of mainly sands and gravel with a clay or silty sand layer of varying thickness located at depths ranging from 0.5 - 25 m below the surface (Golder Associates Ltd., 2005b). From BB to MVB, the nearshore surficial aquifer is comprised of mainly fine sands (Golder Associates Ltd., 2005b).
Figure 3-1: Map of the Lake Huron and Georgian Bay showing the location of the study area in Nottawasaga Bay near the Township of Tiny (red box) and track of regional-scale offshore $^{222}\text{Rn}$ and ERT survey (red line). Map on left is reproduced from Northeast Michigan Lake Huron Watershed (2014). The six locations where local-scale groundwater discharge measurements were conducted are also shown (white dots).

3.3 Methods

3.3.1 Regional-scale measurements

3.3.1.1 Offshore $^{222}\text{Rn}$ surveys

A regional-scale offshore $^{222}\text{Rn}$ survey was conducted on June 24th - 25th, 2014 to evaluate direct groundwater discharge and identify shoreline areas with potentially higher groundwater inputs (herein called groundwater discharge hotspots; Figure 3-1). The survey was performed by continuously sampling for $^{222}\text{Rn}$ from a boat travelling along the shoreline as close to the shore as possible (typically 50 – 200 m offshore). $^{222}\text{Rn}$ has been used extensively over the last two decades to quantify groundwater discharge into marine coastal waters, small lakes as well as rivers (Cable et al., 1996, Corbett et al., 1997, Burnett and Dulaiova, 2003, Burnett et al.,
2006, Somashekar and Ravikumar, 2010, Dimova et al., 2013, Ono et al., 2013, Dimova et al., 2015). $^{222}\text{Rn}$ is a suitable tracer to quantify groundwater discharge because $^{222}\text{Rn}$ is a conservative gas with a short half-life ($t_{1/2} = 3.8$ days) and typically has significantly higher concentrations in groundwater than in surface water (Swarzenski et al., 2003, Burnett and Dulaiova, 2006).

Continuous $^{222}\text{Rn}$ measurements were performed using an automated radon monitoring system (Figure 3-2). Lake water was pumped continuously to the system using a submersible pump (Little Giant, Franklin Electric Co., Inc.) installed at a depth of approximately 0.5 m below the lake water surface as a boat travelled at a maximum speed of 3 km h$^{-1}$. Pumped water was delivered to an air-water exchanger where radon was distributed between the flowing water and air in a closed air loop. The closed air loop passes through commercial radon-in-air monitoring units (RAD7, Durridge Co., Inc.) to determine the $^{222}\text{Rn}$ concentrations. Due to time taken for $^{222}\text{Rn}$ to equilibrate between water and air in the air-water exchanger, long sampling integration times are required particularly when surface water concentrations are low. As low $^{222}\text{Rn}$ concentrations were measured in the lake water in the study area we used a sampling integration time of 15 min and three RAD7 units were connected in parallel. Dulaiova et al. (2005) showed that running multiple RAD7 units in parallel increases the system response and subsequently reduces the measurement uncertainties. The measurement accuracy for RAD7 units are ±5% (DURRIDGE Company Inc., 2015). The uncertainties reported alongside the $^{222}\text{Rn}$ concentrations are the standard deviations following Poisson statistics (square root of the counts from RAD7) (DURRIDGE Company Inc., 2015). It is recommended that the $^{222}\text{Rn}$ uncertainties are below ±20% when using $^{222}\text{Rn}$ as a tracer to quantify groundwater discharge (Dulaiova et al., 2005). The boat position and thus sampling locations were recorded by a handheld GPS (GeoExplorer 7, Trimble Navigation Ltd.).
Figure 3-2: Schematic of $^{222}\text{Rn}$ measurement system with three RAD7 units connected in parallel (modified from Dulaiova et al., 2005).

A high spatial resolution survey was conducted in an area identified from the regional-scale survey as a potential groundwater discharge hotspot. The high-resolution survey was conducted on June 26$^{th}$, 2014 and covered a 4 km stretch of shoreline. For the high-resolution survey, continuous measurements were obtained along 3-4 alongshore transects located at increasing distance offshore. The first shore-parallel transect was run as close to the shoreline as possible (100 - 300 m offshore) with each transect located approximately 200 m further offshore.

A steady state $^{222}\text{Rn}$ mass balance model was applied to estimate the groundwater discharge rates using $^{222}\text{Rn}$ data from the regional-scale and high-resolution surveys (Cable et al., 1996, Burnett et al., 2001, Dulaiova et al., 2010, Schmidt et al., 2010, Smith and Swarzenski, 2012). Each $^{222}\text{Rn}$ measurement was assumed to represent the $^{222}\text{Rn}$ concentration in a well-mixed coastal box, the volume ($V$, m$^3$) of which was determined based on the distances between adjacent measurement points and the water depth. The mass balance model was applied to each $^{222}\text{Rn}$ measurement. The model assumes the steady state conditions whereby the $^{222}\text{Rn}$ inputs to the box equal the $^{222}\text{Rn}$ outputs (see Section 2.2.1). For our study area, the main $^{222}\text{Rn}$ input to the box is thought to be groundwater discharge ($J_{gw}$, dpm m$^{-2}$ d$^{-1}$) and the main losses of $^{222}\text{Rn}$ are radioactive decay ($J_{\text{decay}}$, dpm m$^{-2}$ d$^{-1}$) and evasion to the atmosphere ($J_{\text{atm}}$, dpm m$^{-2}$ d$^{-1}$). $^{222}\text{Rn}$ production from $^{226}\text{Ra}$ ($J_{\text{prod}}$, dpm m$^{-2}$ d$^{-1}$) can be an important $^{222}\text{Rn}$ input in marine coastal waters, but in freshwater environments $^{226}\text{Ra}$ concentrations are low and
therefore this input may be neglected in mass balance calculations (Moore, 1996, Dulaiova and Burnett, 2008). Diffusion from sediments ($J_{\text{diff}}$, dpm $\text{m}^{-2} \text{d}^{-1}$) can also deliver $^{222}\text{Rn}$ to the coastal water box in some settings, but quantification of diffusive fluxes from similar bottom sediments in nearby Lake Simcoe, indicated that input of $^{222}\text{Rn}$ from sediment diffusion is likely very small. This is consistent with other studies that have shown diffusion of $^{222}\text{Rn}$ from sediments in coastal areas with permeable sediments is often much lower than $^{222}\text{Rn}$ inputs from groundwater discharge (Burnett and Dulaiova, 2003, Smith and Swarzenski, 2012). Finally, $^{222}\text{Rn}$ losses due to offshore mixing processes ($J_{\text{mix}}$, dpm $\text{m}^{-2} \text{d}^{-1}$) are thought to be minor and not considered in the mass balance calculations as Nottawasaga Bay is non-tidal and the nearshore waters were extremely calm over the survey periods.

Following the approach of Dulaiova et al. (2010), all $^{222}\text{Rn}$ measurements from the regional-scale and high-resolution surveys were converted to a total groundwater discharge rates ($Q_{\text{TGD}}$, m$^3$ d$^{-1}$) by:

$$Q_{\text{TGD}} = \frac{C_{cw} \times V}{\tau \times C_{gw}}$$

(1)

where $C_{cw}$ is the $^{222}\text{Rn}$ concentration in the surface water corrected for non-groundwater discharge losses, $C_{gw}$ is the $^{222}\text{Rn}$ concentration in groundwater end-member (dpm m$^{-3}$) and $\tau$ is the flushing time of water in the coastal water box (d). It was assumed that $\tau$ was equal to the $^{222}\text{Rn}$ mean life ($\tau = \frac{1}{\lambda_{\text{Rn}}} = 5.53$ d, where $\lambda_{\text{Rn}}$ is the $^{222}\text{Rn}$ decay rate). For each $^{222}\text{Rn}$ measurement $V$ was calculated based on the average water depth ($z$), length of shoreline between subsequent sampling points and the distance between sampling points in the offshore direction. For the regional-scale survey, the distance between the shoreline and a sampling point was used as the distance offshore in the calculation. For comparison with other methods, $Q_{\text{TGD}}$ measured in regional-scale survey was divided by the length of shoreline between subsequent sampling points to calculate the groundwater discharge per unit width of shoreline ($Q_{\text{gw}}$, m$^3$ m$^{-1}$ d$^{-1}$). For high-resolution survey, $Q_{\text{TGD}}$ was divided by the representative lake bed area (based on distances between adjacent measurement points) to calculate the specific groundwater flux or groundwater discharge per unit area of lakebed ($q_{gw}$, m d$^{-1}$). $q_{gw}$ were then
integrated by distance offshore to estimate $Q_{gd}$ at discrete shoreline locations for comparison with $Q_{gd}$ estimated from the regional-scale survey.

Groundwater samples were collected from shallow temporary groundwater wells installed at six beaches to measure $^{222}$Rn concentrations in the groundwater end-member ($C_{gw}$, see Figure 3-1). Wells were located 10 - 100 m landward of the shoreline. Samples were collected using a peristaltic pump and stored in airtight glass vials (40 ml or 250 ml according to expected $^{222}$Rn concentrations). Care was taken when sampling and filling the vials to avoid entrainment of air bubbles. $^{222}$Rn concentrations in groundwater were determined using a RAD H2O (Durridge Co., Inc.) connected to a RAD7 unit.

$C_{cw}$ was determined for all surface water sampling points by correcting the measured $^{222}$Rn concentration in the surface water ($C_w$, dpm m$^{-3}$) for the following $^{222}$Rn losses:

1) $^{222}$Rn loss to the atmosphere ($J_{atm}$) was calculated by (Burnett and Dulaiova, 2003):

$$J_{atm} = k(C_w - \alpha C_{air})$$

where $k$ is the gas-transfer coefficient (m d$^{-1}$), $C_{air}$ is the $^{222}$Rn concentration in the atmosphere (dpm m$^{-3}$) and $\alpha$ is a partitioning coefficient of $^{222}$Rn between water and air (dimensionless). $k$ was calculated using the empirical equation (MacIntyre et al., 1995):

$$k(600) = 0.45 \times u_{10}^{1.6} \times (Sc/600)^{-b}$$

where $Sc$ is the Schmidt number or the ratio of kinematic viscosity ($v$) to molecular diffusivity ($D_m$) (i.e. $Sc = v/D_m$) and $u_{10}$ is the wind speed at a 10 m height above the water surface (m d$^{-1}$); $b = 0.5$ for $u_{10} > 3.6$ m s$^{-1}$ or $b = 0.667$ for $u_{10} < 3.6$ m s$^{-1}$ (MacIntyre et al., 1995, Dulaiova and Burnett, 2006). The wind speed was not measured on the boat but rather wind speed data was obtained from the Environment of Canada weather station at Collingwood which is located approximately 30 km away from the study site (Environment Canada, 2015a). Sensitivity analyses were conducted to evaluate the sensitivity of the results to wind speed. $\alpha$ is temperature-dependent and was calculated by (Burnett and Dulaiova, 2003):
\[
\alpha = 0.105 + 0.405\exp(-0.05027T)
\]

where \(T\) is the temperature of the water (°C) measured in the air-water exchanger by a EL-USB-TC temperature datalogger (Lascar Electronics Inc.).

2) \(^{222}\text{Rn}\) loss due to decay \((J_{\text{decay}})\) was calculated by:

\[
J_{\text{decay}} = z\lambda_{\text{Rn}} C_{\text{Rn}}
\]

3.3.1.2 Offshore ERT survey

A continuous two-dimensional ERT survey was conducted simultaneously with the regional-scale \(^{222}\text{Rn}\) survey on June 24\(^{th}\) – 25\(^{th}\), 2014 to provide insight into the offshore surficial geology and thus assist in interpretation of the \(^{222}\text{Rn}\) survey results. Measured ER is a function of properties of the subsurface geologic materials (e.g., porosity, sediment density) as well as the electrical conductivity of the water (Rein et al., 2004). In freshwater environments, such as the Great Lakes, variations in the water electrical conductivity are limited compared with marine environments, and therefore anomalies in the ER data are mainly associated with geologic variability. The ERT survey was conducted by continuously profiling along the 17 km stretch of shoreline using a floated towed SuperSting R8 resistivity meter (Advanced Geosciences, Inc). The array towed behind the boat consisted of a 30 m floating cable with electrodes positioned with 3 m spacing. Similar to Befus et al. (2014), the SuperSting R8 system continuously controlled current injection to create resistivity mapping using a translating dipole-dipole array. The ERT dataset was inverted to generate two-dimensional distributions of ER below the lake bed using the Res2DInv software program with the absolute inversion errors less than 2% (Loke, 2010). Details of the ERT survey set up and inversion method are provided in Befus et al. (2014).
3.3.2 Local-scale measurements

Local-scale groundwater discharge measurements were conducted from June 2013 – July 2014 at six beach sites along the study shoreline (see Figure 3-1) to compare local-scale estimates with the $^{222}\text{Rn}$ survey results.

3.3.2.1 Horizontal hydraulic gradient measurements

A temporary shore-normal transect of shallow groundwater monitoring wells was installed at all six beach sites between June 2013 - July 2014 to measure the horizontal hydraulic gradient perpendicular to the shoreline. Transects consisted of 4 - 6 wells and extended from the shoreline to up to 100 m onshore. The groundwater wells were installed manually using a hand auger. The specific locations of the installed wells and the measurement date for all sites are provided in Figure 3-9. Groundwater levels were measured using a portable electronic water tape with an accuracy of ± 3 mm (Heron Instruments Inc.) and the locations and elevations of all wells were surveyed using a total station (GTS-239W, Topcon Positioning System Inc.). The lake water level was measured using a stilling piezometer - transparent polycarbonate tube with measuring tape attached to the outside (Cartwright and Nielsen, 2001).

The hydraulic gradient at each site was used to estimate $Q_{gd}$ by applying Darcy’s Law:

$$Q_{gd} = -K \frac{dh}{dl} D$$  \hspace{1cm} (6)

where $D$ is the depth of the unconfined aquifer at the transect location (m); $h$ is the measured head at each location and $L$ is the distance between groundwater wells (m). $D$ for all sites was determined from Ontario Ministry of the Environment Well Records (Ontario Ministry of Environment, 2015). Grain size analysis was performed on sediment samples collected from the base of each well during installation and for each well location $K$ was estimated using the Hazen method (Hazen, 1911) (see Appendix 1). The horizontal hydraulic gradient $\frac{dh}{dl}$ used in Eqn. (6) was calculated as difference between the measured groundwater heads at the furthest landward well and the shoreline. Along permeable straight shorelines (i.e. no headlands) the
main groundwater flow direction is perpendicular to the shoreline, and therefore horizontal hydraulic gradient measurements were aligned perpendicular to the shoreline.

3.3.2.2 Vertical hydraulic gradient measurements

Local-scale spatial variability of groundwater discharge was evaluated and $Q_{gd}$ was estimated using 4-5 nested mini-piezometers installed in a shore-normal transect extending from the shoreline to 5 m offshore. This measurement technique was used only at MVB and BB sites as the nested mini-piezometers could not be manually installed at other sites due to the presence of thin gravel and/or clay layers. Nested mini-piezometers provided the vertical differential head between two known depths directly below the sediment-water interface. The nested mini-piezometers were made from flexible 50 mm diameter tubing attached to a rigid PVC pipe for installation purposes and screened at the base. Mini-piezometers had openings spaced at 0.2 m depth intervals below the sediment-water interface with piezometer tubes attached to a differential manometer to measure the vertical differential head. An oil-water manometer system was used to magnify the differential head and thus allow for more accurate measurement (Gibbes et al., 2007). $K$ was estimated for each nested mini-piezometer location using the Hazen method, and specific flux ($q_{gd}$, m d$^{-1}$) was calculated for each location by applying Darcy’s Law. $q_{gd}$ estimates were then integrated by distance offshore to estimate $Q_{gd}$ (Burnett et al., 2006).

3.3.2.3 Vertical temperature profile measurements

Vertical temperature profiling below the sediment-water interface was also used to estimate $Q_{gd}$ at BB (June 25th, 2014). Vertical temperature profiles were obtained by a “temperature stick” method (Anibas et al., 2011). The temperature sticks were inserted into the sediment to reach a depth of 80 cm and temperature was measured using thermocouples with a resolution of ± 0.1 °C. This method was only applied at one site due to challenges with equipment installation in gravel and clay sediment. Using temperature as a tracer to quantify groundwater discharge requires a large temperature difference between surface water and groundwater. As a result, early spring and summer are the ideal seasons for using temperature as a tracer to
evaluate groundwater discharge to the Great Lakes. A schematic and description of the
temperature stick measurement approach is provided in Section 2.2.3 (Figure 2-4).
Temperature measurements were made at four discrete locations along a shore-normal transect
that extended from the shoreline to 15 m offshore at BB. As analysis of the temperature profile
data was based on an assumption of steady, 1-D (vertical) groundwater flow, temperature sticks
were not installed within 2 m of the shoreline due to the more complex groundwater flow
patterns in this area. At each measurement location, the temperature stick was installed for at
least 10 minutes to ensure stable temperature readings. The temperature survey was conducted
on a cloudy day to avoid any potential heating of the surface water over the measurement
period.

Analysis of vertical temperature profiles was based on solution of the steady-state one-
dimensional heat conduction-advection equation (Taniguchi et al., 2003):

\[
\frac{\partial^2 T}{\partial d^2} - \frac{\rho_w c_w q_{gd}}{\phi \kappa_e} \frac{\partial T}{\partial d} = 0
\]  

(7)

where \(\kappa_e\) is thermal conductivity of the sediment (J s\(^{-1}\) m\(^{-1}\) K\(^{-1}\)); \(T\) is groundwater temperature
(\(^{\circ}\)C); \(d\) is depth below the sediment-water interface (cm); \(\phi\) is the porosity of the sediment; \(\rho_w\)
is density of the groundwater (g cm\(^{-3}\)) and \(c_w\) is the specific heat capacity of the groundwater
(J kg\(^{-1}\) K\(^{-1}\)). We used the \(\rho_w = 1.0\) g cm\(^{-3}\) and \(c_w = 4181\) J kg\(^{-1}\) K\(^{-1}\) for groundwater. We used
the \(\phi = 0.25-0.35\) from Golder Associates Ltd. (2005a). KD-2 Pro Thermal Properties Analyser
(Labcell Ltd.) was used to measure \(\kappa_e\) for sediment collected at all locations that the
temperature stick was installed. \(\kappa_e\) was determined to be between 1.30 and 1.63 J s\(^{-1}\) m\(^{-1}\) K\(^{-1}\) at
BB. Use of \(q_{gd}\) in Eqn. 7 assumes all groundwater discharge across the interface is in the vertical
direction.

A type curve method that is based on solution of eqn. (7) with boundary conditions \(T = T_0\) at
\(z = 0\) and \(T = T_L\) at \(d = L\) was used to estimate \(q_{gd}\) (Bredehoeft and Papaopoulos, 1965).
Solution of (7) is:

\[
\frac{T - T_0}{T_L - T_0} = \frac{\exp(\beta d/L) - 1}{\exp(\beta) - 1}
\]

(8)
where \( L \) is length of vertical distance between \( T_0 \) and \( T_L \) (m) and \( \beta \) is defined as:

\[
\beta = \rho_w c_w q_{gd} \frac{L}{\vartheta_{K_e}} \quad \text{(9)}
\]

Using the type curve method, the normalized temperature \( \frac{T-L}{T_L-T_0} \) vs. vertical distance \( (d/L) \) was first plotted and the observed non-dimensional temperature profiles were matched with type curves to obtain \( \beta \). \( q_{gd} \) was then calculated as:

\[
q_{gd} = \frac{\vartheta_{K_e} \beta}{\rho_w c_w L} \quad \text{(10)}
\]

The calculated \( q_{gd} \) for all measurement locations was integrated by distance offshore to estimate \( Q_{gd} \) (Taniguchi et al., 2002, Burnett et al., 2006).

### 3.4 Results and discussion

#### 3.4.1 Regional-scale \( ^{222} \text{Rn} \) and ERT survey

Results from the regional-scale survey reveal large spatial variability in the \( ^{222} \text{Rn} \) concentrations and ER profiles along the surveyed shoreline (Figure 3-3). The average \( ^{222} \text{Rn} \) concentration in the nearshore water was \( 0.42 \pm 0.15 \text{ dpm L}^{-1} \). Overall, \( ^{222} \text{Rn} \) concentrations in the nearshore lake water are relatively low compared to other surface waters (e.g. \( \sim 8 \text{ dpm L}^{-1} \) in streams discharging into Lake Simcoe, \( > 10 \text{ dpm L}^{-1} \) for marine coastal waters in Gulf of Mexico (Smith, 2012)). Prior studies have shown that \( ^{222} \text{Rn} \) can still be used to quantify groundwater discharge in low \( ^{222} \text{Rn} \) surface waters (e.g. Burnett et al., 2008). The highest \( ^{222} \text{Rn} \) concentration in the surface water was observed 4 km from the northern extent of the surveyed shoreline near LB (\( 1.71 \pm 0.35 \text{ dpm L}^{-1} \)) and C15 (\( 1.35 \pm 0.31 \text{ dpm L}^{-1} \)). The higher \( ^{222} \text{Rn} \) concentrations in these areas compared to the rest of the shoreline suggests that these areas may be potential groundwater discharge hotspots. High \( ^{222} \text{Rn} \) concentration in the surface water was also found 4 km further south near WB (\( 1.43 \pm 0.31 \text{ dpm L}^{-1} \)) - this may be due to the indirect groundwater discharge to Lafontaine Creek.
$^{222}\text{Rn}$ concentrations in the groundwater at the six beach sites were found to be 2-4 orders of magnitude higher than concentrations in the surface water. $^{222}\text{Rn}$ concentrations ranged from 77 ± 22 dpm L$^{-1}$ (C12) to 212 ± 51 dpm L$^{-1}$ (LB) with the variability attributed to different sediments in the nearshore aquifer across the surveyed shoreline (Figure 3-1, Table A2-2). $^{222}\text{Rn}$ groundwater concentrations were not elevated in areas where the surface water $^{222}\text{Rn}$ concentrations were elevated suggesting that the observed spatial variability in the surface water was not due to variability in the groundwater end-member concentrations. It is noted that only one groundwater sample was collected at each beach site. As $^{222}\text{Rn}$ concentrations can be highly heterogeneous in groundwater (Mullinger et al., 2007), it is recommended that multiple groundwater samples be collected and/or sediment equilibrium experiments be set up in the future to better characterize the groundwater end-member.

Figure 3-4 shows the estimated $Q_{gd}$ along the shoreline calculated using Eqn. (1). The spatial variability in $Q_{gd}$ are consistent with the variability in $^{222}\text{Rn}$ surface water concentrations with high $Q_{gd}$ estimated near LB (4.54 ± 1.10 m$^3$ m$^{-1}$ d$^{-1}$) and C15 (5.51 ± 1.33 m$^3$ m$^{-1}$ d$^{-1}$). Reasons for the higher estimated $Q_{gd}$ around LB and C15 are provided in Section 3.4.4. Mass balance calculations were not performed for $^{222}\text{Rn}$ measurements located within 500 m of the mouth of Lafontaine Creek. Input parameter values and results for key components in the mass balance calculations (Eqn. (1)) are provided in Table A2-2.
Figure 3-3: Inverted ER profiles (left) and the distribution of $^{222}\text{Rn}$ concentrations (right) from the regional-scale survey. The white dots represent the beach sites where local-scale groundwater discharge measurements were conducted. The red box covers the area where the high spatial resolution $^{222}\text{Rn}$ survey was completed.
Figure 3-4: Calculated $Q_{gd}$ along the shoreline surveyed in the regional-scale survey.

The blue boxes with error bars show the calculated $Q_{gd}$ values. The white dots represent the beach sites where local-scale groundwater discharge measurements were conducted.

The ERT survey results show anomaly higher ER values around the shoreline area where high $^{222}$Rn concentrations were observed (i.e. LB and C15) compared to the remainder of the surveyed shoreline. The higher ER around and just north of LB and C15 may be attributed to different offshore subsurface geologic conditions such as higher permeability (i.e. sand and gravel sediments will generally have a higher resistivity than clay and silt bottom sediments). This interpretation needs to be confirmed by additional field ground-truthing.

3.4.2 High resolution $^{222}$Rn survey

$^{222}$Rn concentrations in the high-resolution survey conducted from Belle-eau-Clair Beach to Cove Beach ranged from $0.14 \pm 0.10$ to $3.30 \pm 0.48$ dpm L$^{-1}$ with an average concentration of $1.10 \pm 0.26$ dpm L$^{-1}$ (Figure 3-5). Consistent with the regional-scale survey (Figure 3-3), the highest $^{222}$Rn concentrations were observed near LB and C15. Figure 3-6 shows the $^{222}$Rn
concentration versus distance offshore for four offshore transects (T1-T4). Importantly, $^{222}\text{Rn}$ concentrations were found to be highest near the shoreline with concentrations decreasing exponentially offshore ($R^2 = 0.82-0.93$). This result indicates that the high surface water $^{222}\text{Rn}$ concentrations are likely due to nearshore groundwater discharge.

Figure 3-5: $^{222}\text{Rn}$ concentrations measured in the high-resolution survey. The white dots represent the beach sites where local-scale groundwater discharge measurements were conducted. $^{222}\text{Rn}$ concentrations vs. distance offshore are shown in Figure 3-6 for the four shore-normal transects indicated (black lines, T1-T4).
Figure 3-6: $^{222}\text{Rn}$ concentrations from the high-resolution $^{222}\text{Rn}$ survey as a function of distance offshore. The error bars show the $2\sigma$ uncertainty for the measured $^{222}\text{Rn}$ concentrations. Locations of the four shore-normal transects, T1-T4, are shown in Figure 3-5.

Figure 3-7 shows $q_{gd}$ calculated for all measurement points in the high-resolution survey. A summary of input parameters and key components in the $^{222}\text{Rn}$ mass balance calculation are summarized in Table A2-1. Similar to the spatial variability in $^{222}\text{Rn}$ concentrations, $q_{gd}$ was highest ($(3.85 \pm 0.93) \times 10^{-2}$ m d$^{-1}$) close to the shoreline near C15 and LB. In contrast to the $^{222}\text{Rn}$ concentrations, $q_{gd}$ did not consistently decrease offshore and this was due to consideration of the increasing water depth offshore in the mass balance calculations. $q_{gd}$ along transects perpendicular to the shorelines at C15 and LB (see Figure 3-7) were converted to $Q_{gd}$. The estimated $Q_{gd}$ are $7.38 \pm 1.78$ m$^3$ m$^{-1}$ d$^{-1}$ and $6.17 \pm 1.49$ m$^3$ m$^{-1}$ d$^{-1}$ for C15 and LB, respectively. The estimated $Q_{gd}$ for high-resolution survey are higher than that for regional-scale survey ($5.51 \pm 1.33$ m$^3$ m$^{-1}$ d$^{-1}$ and $4.54 \pm 1.10$ m$^3$ m$^{-1}$ d$^{-1}$ for C15 and LB, respectively). This discrepancy is because for the regional-scale survey data, $Q_{gd}$ only included groundwater discharge close to the shore, whereas $q_{gd}$ further offshore, although low, was considered for in the $Q_{gd}$ calculation using the high-resolution data.
Evasion of $^{222}\text{Rn}$ to the atmosphere ($J_{\text{atm}}$) represents the largest loss of $^{222}\text{Rn}$ from the nearshore water. This flux accounted for approximately 70-80% of the total $^{222}\text{Rn}$ losses for calculations done using the regional-scale and high-resolution $^{222}\text{Rn}$ survey data (see Table A2-1 and A2-2). Additional calculations were performed taking the wind speed values ($u_{10}$) as the average wind speed from the previous 12 hours, 2 days and 5 days. Variations in these input parameter values led to variation in $Q_{\text{gd}}$ up to 20% (see Appendix 3). The overall spatial variability in $Q_{\text{gd}}$ along the shoreline however remained consistent regardless of these input parameter values.

Figure 3-7: $q_{\text{gd}}$ calculated from high-resolution $^{222}\text{Rn}$ survey data. The colored dots represent the sampling points with the color indicating the calculated $q_{\text{gd}}$ value. The numbers adjacent to the dots represent the sampling point number with data provided in Table A2-1. The white dots show the beach sites where local-scale groundwater discharge measurements were conducted. The white lines are the transects where $Q_{\text{gd}}$ were calculated for C15 and LB.
3.4.3 Local-scale groundwater discharge measurements

The horizontal hydraulic gradient measurements indicated lakeward groundwater flow at all six beach sites (Figure 3-8). The hydraulic gradients ranged from 0.009 – 0.021 with gradients highest for LB and MVB. When the hydraulic gradients were multiplied by estimated K and D for each site, $Q_{gd}$ was found to be highest for LB (12.4 – 29.4 m$^3$ m$^{-1}$ d$^{-1}$) compared to all other sites, with higher $Q_{gd}$ also found for C15 (0.6 – 2.2 m$^3$ m$^{-1}$ d$^{-1}$) and MVB (1.2 – 5.9 m$^3$ m$^{-1}$ d$^{-1}$; Table 3-1). The range in $Q_{gd}$ values at each site is due to the range in K determined from sediment samples collected at each groundwater well location. While the groundwater level data for the six sites were collected at different times between June 2013 – July 2014, continuous two-year data of groundwater levels at BB and MVB show the horizontal hydraulic gradients at these sites were relatively constant and therefore the overall observed trends between the sites should be consistent (i.e. gradients ranged from 0.011-0.016 at BB and 0.020-0.026 at MVB over the two-year monitoring period).

Specific flux ($q_{gd}$) as a function of distance offshore for BB and MVB, calculated from the vertical hydraulic gradient data obtained from nested-piezometers, are shown in Figure 3-9. The highest $q_{gd}$ was observed near the shoreline at both BB (0.31 m d$^{-1}$) and MVB (0.23 m d$^{-1}$) with $q_{gd}$ decreasing offshore and becoming negligible at a distance of only ~7 m offshore. Integrating $q_{gd}$ with the distance offshore provides estimates of $Q_{gd}$ for BB and MVB to be 1.2 m$^3$ m$^{-1}$ d$^{-1}$ and 1.7 m$^3$ m$^{-1}$ d$^{-1}$, respectively.
Figure 3-8: Groundwater level (blue line) and sand surface profiles (black line) at the six study sites. The filled blue squares represent the locations of the wells. The horizontal hydraulic gradient \( \frac{dh}{dl} \) was calculated from the groundwater level measurements at each site.
Figure 3-9: Specific flux ($q_{gd}$) as a function of distance offshore for Balm Beach (BB) and Mountain View Beach (MVB). Data was obtained at field sites in June 25th, 2014 using vertical nested mini-piezometers.

The vertical temperature measurements were conducted at BB in June 2014 when the surface water temperature was high (around 18°C) compared to the groundwater temperature (around 8°C at a depth of 80 cm below the sediment-water interface). Figure 3-10 (a) shows the vertical temperature profiles at all sampling locations along the transect at BB. All vertical temperature profiles are convex with temperature decreasing with depth. This indicates that the groundwater flow below the sediment-water interface was upward (Taniguchi et al., 2003). The temperature decrease was greatest (4-6°C) in the first 20 cm below the sediment-water interface and became more stable with depth for all measurement locations except 1A. For measurement locations 1A, the temperature only decreased slightly (less than 0.5 °C) over the first 20 cm below the sediment-water interface. This is may be due to surface water infiltrating across the sediment-water interface near the shoreline due to wave action. Figure 3-10 (b) shows the estimated $q_{gd}$ as a function of distance offshore calculated from the vertical temperature profiles. The highest $q_{gd}$ was observed 5 m from the shoreline (0.23 m d$^{-1}$) with $q_{gd}$ decreasing with distance offshore. $q_{gd}$ at 1A was lower than further offshore, likely due to infiltration of surface water close to the
shoreline causing a non-steady vertical temperature distribution. The $Q_{gd}$ was calculated to be $0.97 \pm 0.10 \text{ m}^3 \text{ m}^{-1} \text{ d}^{-1}$.

![Figure 3-10](image_url)

**Figure 3-10**: (a) Vertical temperature profiles measured at BB in June 25th, 2014 and (b) specific flux ($q_{gd}$) estimated from vertical temperature profiles as a function of distance offshore at BB in June 2014.

### 3.4.4 Comparison of methods

$Q_{gd}$ determined at the six beach sites using the different methods are summarized in Table 3-1. The spatial variability in $Q_{gd}$ along the shoreline was consistent between the $^{222}\text{Rn}$ and horizontal hydraulic gradient methods (i.e. high $Q_{gd}$ near LB and C15 compared to the other sites), thus providing confidence in these shoreline areas being identified as groundwater discharge hotspots. $Q_{gd}$ estimated from the horizontal hydraulic gradient measurements, however, was generally higher compared to $Q_{gd}$ estimated from the other methods particularly for LB, BB and MVB. Higher $Q_{gd}$ estimated from the horizontal hydraulic gradient measurements was unexpected because this method only quantifies the terrestrial (inland) groundwater discharge while the other methods also include in their estimates water that is recirculating across the sediment-water interface (i.e. due to wave action). Potential reasons for discrepancies in the estimated $Q_{gd}$ are described below.
The comparatively higher $Q_{gd}$ estimated for LB and C15 compared to the other sites may be associated with groundwater discharge from a tunnel valley aquifer system (sand and gravel) that runs north-south and outcrops at Nottawasaga Bay around this location (Golder Associates Ltd., 2014, Appendix 5). Physiographical and hydrogeological characteristics of this shoreline area also support the higher observed $Q_{gd}$. For instance, the topography landward of the shoreline is steep (rises 20 m over 500 m distance) and the surficial unconfined aquifer is thick (15 - 20 m) and highly permeable (mainly medium sands and rounded gravel) (Golder Associates Ltd., 2005a, Ontario Ministry of Environment, 2015).

The horizontal hydraulic gradient measurements provide lower estimates of $Q_{gd}$ at WB (0.6 - 1.9 m$^3$ m$^{-1}$ d$^{-1}$) compared to other sites - this is not consistent with the high $^{222}$Rn concentrations observed offshore from WB (Figure 3-3). Lafontaine Creek discharges to the south of WB and the high $^{222}$Rn concentrations observed offshore may be attributed to elevated $^{222}$Rn concentrations in the creek discharge rather than direct groundwater discharge. $^{222}$Rn concentrations in nearby creeks in Simcoe County have been found to range from 6.5 - 15.7 dpm L$^{-1}$ which is considerably higher than $^{222}$Rn concentrations in the nearshore lake water (average value = 1.14 ± 0.21 dpm L$^{-1}$). Furthermore, as the Lafontaine Creek flows behind and parallel with WB, the creek may incise the surficial aquifer and capture the groundwater flowing towards the lake - this would lead to the smaller observed hydraulic gradient and lower direct groundwater discharge.

The high-resolution $^{222}$Rn survey data (Figure 3-5), vertical hydraulic gradient data (Figure 3-9) and vertical temperature profile data (Figure 3-10) indicate groundwater discharge is highest near the shoreline with discharge decreasing offshore. This is to be expected for discharge from a permeable unconfined aquifer. Aside from the larger uncertainty in $Q_{gd}$ estimated from the horizontal hydraulic gradient measurements, $Q_{gd}$ estimated for BB (0.4 - 1.9 m$^3$ m$^{-1}$ d$^{-1}$; 0.6 - 2.7 m$^3$ m$^{-1}$ d$^{-1}$) and MVB (1.3-2.0 m$^3$ m$^{-1}$ d$^{-1}$; 1.2 - 5.9 m$^3$ m$^{-1}$ d$^{-1}$) are comparable for the vertical and horizontal gradient techniques, respectively. The spatial offshore variability in $q_{gd}$ measured at BB was different between the vertical hydraulic gradient and temperature methods although these measurements were performed on the same day (Figure 3-9 c.f. Figure 3-10).
The discrepancy is attributed to variations in temperature, wind speed and wave conditions over the measurement period.

While the combination of techniques adopted were suitable for quantifying nearshore groundwater discharge to the Great Lakes at different scales, there are limitations of these techniques and \( Q_{gd} \) estimates could be further refined. For instance, large uncertainties in laboratory-determined \( K \) values as well the depth of the nearshore aquifers \( (D) \) introduced large uncertainty in \( Q_{gd} \) for the hydraulic gradient measurement approaches. It is recommended that pumping tests be conducted at each site to more accurately determine \( K \) and thus minimize this uncertainty. Further, the difficulty in installing mini-nested piezometers and temperature sticks at sites with layers of clay or gravel at shallow depths in the nearshore, limited our ability to use the vertical hydraulic gradient and vertical temperature profiling methods. Simplifications in the \(^{222}\text{Rn} \) mass balance calculations (i.e. neglecting offshore mixing, using wind speed from weather station located 30 km from the boat, and poor characterization of heterogeneities in the \(^{222}\text{Rn} \) groundwater end-member concentrations) may also have caused uncertainties in \( Q_{gd} \) calculated from the \(^{222}\text{Rn} \) survey data.

Table 3-1: Estimated groundwater discharge \((Q_{gd}, \text{m}^3\text{m}^{-1}\text{d}^{-1})\) determined by four measurement techniques at six locations along the shoreline of Nottawasaga Bay. “/” means no data is available.

<table>
<thead>
<tr>
<th></th>
<th>C12</th>
<th>WB</th>
<th>C15</th>
<th>LB</th>
<th>BB</th>
<th>MVB</th>
</tr>
</thead>
<tbody>
<tr>
<td>Vertical hydraulic gradient</td>
<td>/</td>
<td>/</td>
<td>/</td>
<td>/</td>
<td>0.4 -1.9</td>
<td>1.3 - 2.0</td>
</tr>
<tr>
<td>Horizontal hydraulic gradient</td>
<td>1.4 -2.3</td>
<td>0.6 - 1.9</td>
<td>0.6 - 2.2</td>
<td>12.4 - 29.4</td>
<td>0.6 -2.7</td>
<td>1.2 - 5.9</td>
</tr>
<tr>
<td>Vertical temperature profiles</td>
<td>/</td>
<td>/</td>
<td>/</td>
<td>/</td>
<td>0.9 - 1.1</td>
<td>/</td>
</tr>
<tr>
<td>Regional-scale (^{222}\text{Rn} ) survey</td>
<td>0.7 -1.3</td>
<td>/</td>
<td>4.2 - 6.8</td>
<td>3.4 - 5.6</td>
<td>0.9 - 1.4</td>
<td>0.4 -0.6</td>
</tr>
</tbody>
</table>
3.5 Conclusions

Although prior studies have evaluated direct groundwater discharge to the Great Lakes, the magnitude of direct groundwater discharge remains poorly understood. Moreover, there are limited tools and techniques available for quantifying groundwater discharge to large inland water such as the Great Lakes, particularly at the regional-scale. The ability to quantify regional-scale groundwater discharge is a first critical step to better managing this potentially important non-point pollution source. This research evaluated four different methods ($^{222}$Rn, vertical temperature profiling, horizontal and vertical hydraulic gradient measurements) for quantifying groundwater discharge along a 17 km stretch of shoreline in Nottawasaga Bay. Applying multiple methods reduces the limitations and uncertainties of using a single method and also provides characterization of groundwater discharge at multiple scales. An offshore $^{222}$Rn survey was able to successfully identify groundwater discharge hotspots along the surveyed shoreline (near LB and C15) with the measured spatial variability in $Q_{gd}$ along the shoreline confirmed by other measurement techniques. Anomaly high $^{222}$Rn concentrations around WB were attributed to tributary inflow (Lafontaine Creek) to the lake, highlighting the need to identify all external $^{222}$Rn inputs in evaluating regional-scale $^{222}$Rn data. The ERT survey results were able to show that changes in the subsurface surficial lithology may be associated with the higher observed $^{222}$Rn concentrations and thus higher $Q_{gd}$ around LB and C15. The high-resolution $^{222}$Rn survey as well as the vertical temperature profiling and vertical hydraulic gradient measurement showed that groundwater discharge is highest near the shoreline with discharge decreasing offshore. This indicates that direct groundwater discharge quantified is from the surficial unconfined aquifer rather than discharge from confined aquifers that may incise the lake bed further offshore. Uncertainties in using and applying all measurements techniques remain and these uncertainties should be addressed in the future.
3.6 References


Chapter 4

4 Using $^{222}\text{Rn}$ as a tracer to quantify groundwater discharge along the shoreline of Lake Simcoe

4.1 Introduction

Lake Simcoe is the largest lake in southern Ontario, aside from the Great Lakes, with an area of 722 km$^2$. Lake Simcoe is of immense economic, recreational and ecological value, with fisheries and tourism alone estimated to be worth about 200 million dollars annually (Marchildon et al., 2015, Lake Simcoe Region Conservation Authority, 2016, Environment Canada, 2015b). The water quality in Lake Simcoe is impaired due to urban and agricultural activities in the watershed that have changed the natural landscape and vegetation, and contributed to increased pollutants inputs to the lake (Ontario Ministry of the Environment, 2009). The main pollutants that impair water quality in Lake Simcoe are nutrients, particularly phosphorus [P], fecal pollutants (pathogens, $E. \text{coli}$), and contaminants such as chloride, heavy metals, organic chemicals and sediments (Gewurtz et al., 2011, Palmer et al., 2011, North et al., 2013, Gudimov et al., 2015, Oni et al., 2015, Environment Canada, 2015b). There is also increasing concern regarding emerging contaminants such as pharmaceuticals and personal care products (Ontario Ministry of the Environment, 2009). Excessive P loading has been the largest cause of the water quality impairment in Lake Simcoe since the 1970s, resulting in ecosystem changes including the reduction of the cold-water fish populations (e.g. lake trout, lake whitefish and lake herring) and the excessive growth of invasive macrophytes and algae (Evans et al., 1996, Winter et al., 2007, Lake Simcoe Science Advisory Committee, 2008, Gewurtz et al., 2011, Environment Canada, 2015b, Gudimov et al., 2015).

Efforts to quantify and manage pollutant inputs to Lake Simcoe have focused mainly on sources such as water pollution control plants, inputs from tributaries, urban storm water runoff, runoff from agricultural areas and atmospheric deposition (Eimers et al., 2005, Winter et al., 2007, Ontario Ministry of the Environment, 2009). Groundwater discharge may also be
an important pathway for delivering pollutants into Lake Simcoe but this pathway is not well understood with few studies conducted to evaluate groundwater inputs into Lake Simcoe. While groundwater inputs are not thought to be a major component in the overall water balance for the lake (Winter et al., 2007), groundwater may be enriched with pollutants relative to the receiving lake water due to sources including septic systems, leaky sewers and agricultural activities (e.g., fertilizer and manure land application). It has been estimated that septic systems alone may contribute approximately 4.4 tonnes of P to Lake Simcoe annually, representing 6% of the total P loading to the lake (Ontario Ministry of the Environment, 2009). Winter et al. (2007) considered only septic systems located within a 100 m band around the lake and estimated the P loading to be slightly lower (3.87 tonnes per year). These estimates are based on many assumptions due to large uncertainty regarding the groundwater pathway as well as P mobility along its subsurface discharge pathway from the septic tile beds to the lake. Roy and Malenica (2013) more recently measured concentrations of nutrients (P and nitrogen [N]) and toxic contaminants in shallow groundwater below the shoreline in Kempenfelt Bay, Lake Simcoe (see map in Figure 4-1). High concentrations of pollutants including P, nitrate, chloride, chlorinated solvents and petroleum compounds were frequently detected suggesting that urban groundwater may be an important contributor of pollutants to the lake.

To target water quality management plans, there is a need not only to quantify total pollutant loadings associated with the groundwater pathway but also to identify spatial areas where groundwater discharge is substantial (herein called groundwater discharge hotspots) and thus where groundwater pollutants may deteriorate surface water quality. As a requirement of the Lake Simcoe Protection Plan (LSPP), assessments of ecologically significant groundwater recharge areas (ESGRAs) have been conducted for subwatersheds in the Lake Simcoe Basin (e.g. Marchildon et al., 2015). Numerical groundwater models of all subwatersheds have been developed for these ESGRA assessments with water budget calculations estimating the total groundwater discharge to the lake from each subwatershed (Table 4-1). While these estimates provide some indications of potential spatial variability in groundwater discharge around the lake, these estimates have not been validated, and moreover field data is currently not available for this validation. To our knowledge, the only field work that has attempted to identify
groundwater discharge areas was a seismic survey conducted by Lewis et al. (2007) which found deep submarine hollows at the bottom of Kempenfelt Bay. These hollows may act as offshore groundwater discharge points with North et al. (2013) proposing that discharge associated with these hollows may be contributing to lower O\textsubscript{2} concentrations in Kempenfelt Bay with implications for coldwater fish habitats.

**Table 4-1: Total groundwater discharge from subwatersheds to Lake Simcoe estimated using numerical groundwater models developed for ESGRA assessments. Groundwater discharge per m of shoreline (Q\textsubscript{gd}) is calculated based on the estimated shoreline length for each subwatershed.**

<table>
<thead>
<tr>
<th>Subwatershed</th>
<th>Total groundwater discharge (m\textsuperscript{3} d\textsuperscript{-1})</th>
<th>Shoreline length (m)</th>
<th>Q\textsubscript{gd} (m\textsuperscript{3} m\textsuperscript{-1} d\textsuperscript{-1})</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Maskinonge River</td>
<td>864</td>
<td>750</td>
<td>1.15</td>
<td>(GENIVAR Inc., 2013)</td>
</tr>
<tr>
<td>Georgina Creeks</td>
<td>11,731</td>
<td>24,000</td>
<td>0.49</td>
<td>(GENIVAR Inc., 2013)</td>
</tr>
<tr>
<td>Innisfil Creeks</td>
<td>20,510</td>
<td>38,000</td>
<td>0.54</td>
<td>(AquaResource Inc., 2013b)</td>
</tr>
<tr>
<td>Barrie Creeks</td>
<td>9,600</td>
<td>9,500</td>
<td>1.01</td>
<td>(AquaResource Inc., 2013a)</td>
</tr>
<tr>
<td>Lovers Creek</td>
<td>-778</td>
<td>2,500</td>
<td>-0.31</td>
<td>(Earthfx Inc., 2012)</td>
</tr>
<tr>
<td>Hewitts Creek</td>
<td></td>
<td>150</td>
<td>-5.18</td>
<td>(Earthfx Inc., 2012)</td>
</tr>
<tr>
<td>Hawkestone Creek</td>
<td>2,304</td>
<td>23,000</td>
<td>0.10</td>
<td>(Earthfx Inc., 2013)</td>
</tr>
<tr>
<td>Oro Creeks South</td>
<td>15,602</td>
<td>17,500</td>
<td>0.89</td>
<td>(Earthfx Inc., 2012)</td>
</tr>
</tbody>
</table>

The objective of this Chapter is to evaluate the use of the natural tracer $^{222}$Rn as a regional-scale assessment tool for identifying groundwater discharge hotspots along the shoreline of Lake Simcoe. Results are presented from $^{222}$Rn boat surveys conducted in the western and southern parts of Lake Simcoe as well as from high spatial resolution surveys conducted in areas identified as potential groundwater discharge hotspots (Figure 4-1). To reduce uncertainty in the $^{222}$Rn measurement technique and subsequent estimates of groundwater
discharge, factors contributing to observed temporal variability in $^{222}$Rn concentrations in the lake water are also assessed.

4.2 Study site description

The Lake Simcoe shoreline for which regional-scale $^{222}$Rn boat surveys were conducted are shown in Figure 4-1. Surveys were conducted in Kempenfelt Bay extending northward to Hawkestone Region, as well as in the Innisfil and Georgina Areas. Lake Simcoe is typically divided into three parts - Kempenfelt Bay, Cook’s Bay and the main lake (Figure 4-2). The watershed area of Lake Simcoe is around 3400 km$^2$ and contains a population of approximately 0.4 million people (Lake Simcoe Region Conservation Authority, 2016). Barrie is located at the western end of Kempenfelt Bay and is the largest city in the watershed. Lake Simcoe is linked to Georgian Bay (Lake Huron) by Lake Couchiching to the north, and Lake Ontario by Trent-Seven waterway to the south (Roy and Malenica, 2013). The average depth of Lake Simcoe is 15 m with depths reaching up to 42 m near the mouth of Kempenfelt Bay (AquaResource Inc., 2013a).

**North and West side of Kempenfelt Bay:** This survey area covered the Barrie Creeks subwatershed, Oro Creeks South subwatershed and Hawkestone Creek subwatershed (Figure 4-2). The physiography on the north side of Kempenfelt Bay belongs to Simcoe Uplands area which is dominated by rolling till plains as well as broad erosional valleys which contain sand and clay plains (AquaResource Inc., 2013a). The elevation of the water surface in Kempenfelt Bay is around 220 masl, while the Oro Moraine formed an east-west ridge to the north of the study area which can reaches up a higher elevation of to 375 masl (AquaResource Inc., 2013a). The surface water as well as the groundwater drain into Lake Simcoe following the topography. The aquifers are generally composed of till and fine-grained sediment with stratified sand and gravels (AquaResource Inc., 2013a). The thickness of the aquifers ranges from 10 to 30 m in most areas. According to the ESGRA assessment for the Oro Creeks South and Hawkestone Creeks subwatershed, a tunnel valley filled with coarse-textured stratified deposits incised Oro Moraine along the north of Hawkestone Creeks subwatershed which may change the groundwater flow conditions and cause the high groundwater discharge (Earthfx Inc., 2013).
Further, the hydrogeological properties in Barrie Creeks Watersheds is mainly controlled by the aquifers which contain ice contact deposits, kame moraines, coarse-gained sediments with relatively high hydraulic conductivity (AquaResource Inc., 2013a). Creeks including Dyment’s Creek, Bunker’s Creek and Kidd’s Creek discharge into Lake Simcoe in the west end of Kempenfelt Bay, but there are no additional creeks in the north side of Kempenfelt Bay (AquaResource Inc., 2013a). Many creeks including Shelswell Creek, Oro Creek and Hawkestone Creek discharge into Lake Simcoe through Oro Creeks South and Hawkestone Creek subwatershed (Earthfx Inc., 2013). According to the numerical modelling results in the ESGRA assessment (Table 4-1), Oro Creeks South Subwatershed and Barrie Creeks Subwatershed are tend to be potential high groundwater discharge areas.

South side of Kempenfelt Bay and Innisfil Area: This survey area covers Lovers Creeks subwatershed, Hewitts Creeks subwatershed and Innisfil Creeks subwatershed (Figure 4-2). The elevation of Innisfil Heights to the south of Kempenfelt Bay reaches 300 masl with the land elevation dropping to around 220 masl at Lake Simcoe. The Peterborough Drumlin Field characterized as a drumlinized till plain dominates most of this study area but the nearshore area is mainly dominated by sand plains which are characteristic of the Simcoe Lowlands physiographic region (Chapman and Putnam, 1984). A series of small streams flow into the south side of Kempenfelt Bay including Hewitts Creek, Innisfill Creek and Lovers Creek. These creeks have high base flows and therefore indirect groundwater discharge may play an important role here (AquaResource Inc., 2013b). However, as the base flow along the eastern shoreline of Innisfil Area was observed quite little, groundwater discharge is not a significant contributor of the flow in the creeks (AquaResource Inc., 2013b). The ESGRA assessment (Table 4-1) also provides a relatively low predicted groundwater \( Q_{gd} \) for Innisfil Creeks Subwatershed.

Georgina Area: This survey area includes the western part of the Maskinonge subwatershed and Georgina Creeks subwatershed. The north part of the Georgina Creeks subwatershed belongs to the Simcoe Lowlands physiographic region and has lower elevation (220-280 masl) compared to the south part of the subwatershed (320-340 masl). The Oak Ridges Moraine in the south of the study area leads the surface water as well as the groundwater flow northward.
into Lake Simcoe (GENIVAR Inc., 2013). The physiography belongs to the Simcoe Lowland region which is comprised of mostly rolling sandy hills with clay plains in the Georgina Creeks subwatershed (GENIVAR Inc., 2013). The aquifers mainly consist of sand, gravel, diamicton and organic deposits with thickness of around 15 m in the northern part of Georgina Creeks subwatershed (GENIVAR Inc., 2013). Based on the ESGRA assessment results (Table 4-1), Georgina Creek Subwatershed has a relatively low $Q_{gd}$. In our study area, only a small area around Keswick Beach belongs to Maskinonge River subwatershed. Maskinonge River has a relatively high base flow index (BFI, ratio of base flow to total stream flow) of 0.532.

Figure 4-1: Map of the survey areas in Lake Simcoe including the track of regional-scale $^{222}\text{Rn}$ boat surveys (red line), locations for groundwater endmember sampling (white and yellow dots), and locations where local-scale groundwater discharge measurements and groundwater endmember sampling were conducted (yellow dots).
Figure 4-2: Map of subwatershed areas and stream network of Lake Simcoe. The red dots represent the creeks that were sampled for $^{222}$Rn concentrations. Figure modified from The Louis Berger Group Inc. (2010).
4.3 Methods

4.3.1 $^{222}$Rn boat surveys

Regional-scale $^{222}$Rn boat surveys were conducted along 80 km of shoreline as shown in Figure 4-1 from June - September 2015 to identify potential groundwater discharge hotspots. In-lake $^{222}$Rn concentrations were found to exhibit considerable variability between survey days and therefore multiple $^{222}$Rn surveys were conducted along some shoreline areas to evaluate factors (e.g. precipitation, wind speed) that may affect this temporal variability.

The $^{222}$Rn boat surveys were performed by continuously sampling and analyzing $^{222}$Rn as a boat travelled at a maximum speed of 3 km hr$^{-1}$ along the shoreline as close to the shore as possible (typically 50 – 200 m offshore). Lake water was pumped continuously to a $^{222}$Rn detection system using a submersible pump (Little Giant, Franklin Electric Co., Inc.) installed at a depth of approximately 0.5 m below the lake water surface. Lake water was delivered to an air-water exchanger where $^{222}$Rn equilibrated between the flowing water and air in a closed air loop. $^{222}$Rn concentrations were measured as air in the closed air loop passed through commercial radon-in-air monitoring units (RAD 7, Durridge Co., Inc.). Five RAD 7 detectors were connected in parallel and a sampling integration time of 15 min was used to improve accuracy of the $^{222}$Rn measurements due to the low $^{222}$Rn concentrations in the lake. The boat position and thus sampling locations were recorded by a handheld GPS (GeoExplorer 7, Trimble Navigation Ltd.). Conductivity and pH were also sampled continuously during the $^{222}$Rn boat surveys using a YSI logger (YSI 610, YSI Inc.) towed alongside the boat. This data is provided in Table A6-1, A-2, A6-3 and A6-4, Appendix 6 but not discussed in this Chapter as correlations between $^{222}$Rn, conductivity and pH were poor and provided limited insight for identification of groundwater discharge hotspots. Regional-scale $^{222}$Rn surveys were conducted on June 9th and 11th in Kempenfelt Bay, July 6th and 8th on the north side of Kempenfelt Bay towards Hawkestone Creek, July 8th and August 12th in the Innisfil Area, and August 13th and September 23th in the Georgina Area.
Two high spatial resolution $^{222}$Rn boat surveys were conducted in areas where higher in-lake $^{222}$Rn concentrations were observed during the regional-scale surveys and thought to be attributed to direct groundwater discharge into the lake. These areas were near Johnson’s Beach (JB) and Jackson’s Point Beach (JPB) (see Figure 4-1). A high spatial resolution $^{222}$Rn survey was also conducted near Keswick Beach where high in-lake $^{222}$Rn concentrations are thought to be attributed to indirect groundwater discharge into the Maskinonge River (and subsequent discharge to the lake). The high-resolution survey near JB was conducted on July 10th, covering a 4 km stretch of shoreline. The high-resolution surveys near Keswick Beach and JPB were conducted on September 24th and 25th, respectively, along 4 km of shoreline. For the high-resolution surveys, continuous $^{222}$Rn measurements were obtained along three alongshore transects with the first alongshore transect run as close to the shoreline as possible (100 - 300 m offshore) with each transect located approximately 200 m further offshore.

### 4.3.2 $^{222}$Rn mass balance calculations

A steady state $^{222}$Rn mass balance model was applied to estimate the groundwater discharge into the lake using $^{222}$Rn data from the regional-scale and high-resolution surveys (Cable et al., 1996, Burnett et al., 2001, Dulaiova et al., 2010, Schmidt et al., 2010, Smith and Swarzenski, 2012). The mass balance model considers all the sources and losses of $^{222}$Rn concentration in a well-mixed coastal box, the volume ($V$, $m^3$) of which was determined based on the distances between adjacent sampling points and the water depth ($z$, m). The model assumes steady state conditions whereby $^{222}$Rn inputs to the box equal the $^{222}$Rn outputs (see Section 2.2.1). For our study area, the main $^{222}$Rn inputs to the box are groundwater discharge ($J_{gw}$, dpm m$^{-2}$ d$^{-1}$), sediment diffusion ($J_{diff}$, dpm m$^{-2}$ d$^{-1}$) and input from tributary discharge, and the main losses of $^{222}$Rn are radioactive decay ($J_{decay}$, dpm m$^{-2}$ d$^{-1}$) and evasion to the atmosphere ($J_{atm}$, dpm m$^{-2}$ d$^{-1}$). $^{222}$Rn production from $^{226}$Ra can represent an important $^{222}$Rn input in marine environments, but $^{226}$Ra concentrations are low in freshwater environments and therefore this input can be neglected (Moore, 1996, Dulaiova and Burnett, 2008). As Lake Simcoe is non-tidal and the lake was calm over the survey periods, $^{222}$Rn loss due to offshore mixing ($J_{mix}$, dpm m$^{-2}$ d$^{-1}$) is thought to be minor and not considered in our mass balance calculations (Santos,
Tributary discharge into the lake, particularly from groundwater-fed streams with high base flow indices, may also contribute $^{222}$Rn to the nearshore lake water. $^{222}$Rn concentrations were measured in six creeks discharging to Lake Simcoe within the study area to evaluate this important input of $^{222}$Rn to the lake. Creeks sampled include Dyment’s Creek, Whiskey Creek, Lovers Creek, Shelswell Creek, Hawkestone Creek and Georgina Creek (Figure 4-2). $^{222}$Rn concentrations in each creek was measured by continuously pumping water from the middle of the creek to our RAD7 detection system (five RAD7 units connected in parallel) for at least 2 hours. Although $^{222}$Rn concentrations in creeks were measured in the field to understand this $^{222}$Rn source to the lake, this $^{222}$Rn input was not accounted for in the mass balance calculations as we did not simultaneously measure the creek discharges. Mass balance calculations were not performed for $^{222}$Rn measurements located within 500 m of a creek mouth.

Following the approach of Dulaiova et al. (2010), all in-lake $^{222}$Rn measurements from the regional-scale and high-resolution surveys except for those within 500 m of a creek mouth were converted to a total groundwater discharge rate ($Q_{TD}$, m$^3$ d$^{-1}$) by:

$$Q_{TD} = \frac{C_{cw} \times V}{\tau \times C_{gw}}$$

where $C_{cw}$ is the in-lake $^{222}$Rn concentration corrected for non-groundwater discharge losses, $C_{gw}$ is the $^{222}$Rn concentration in the groundwater endmember (dpm m$^{-3}$) and $\tau$ is the flushing time of water in the coastal water box (d). It was assumed that $\tau$ was equal to the $^{222}$Rn mean life ($\tau = 1/\lambda_{Rn} = 5.53$ d, where $\lambda_{Rn}$ is the $^{222}$Rn decay rate). For each $^{222}$Rn measurement, $V$ was calculated based on the average water depth, length of shoreline between subsequent sampling points and the distance between sampling points in the offshore direction. For the regional-scale survey data, the distance between the shoreline and a sampling point was considered as the offshore distance. For comparison with $Q_{td}$ estimates for each subwatersheds by ESGRA (see Table 4-1) as well as local-scale groundwater discharge estimates determined from measurement of the horizontal hydraulic gradient near the shore (see Section 4.3.3), $Q_{TD}$ measured in regional-scale survey was divided by the length of shoreline between subsequent sampling points to calculate the groundwater discharge per unit width of shoreline ($Q_{gd}$, m$^3$ m$^{-1}$ d$^{-1}$). For high-resolution survey, $Q_{TD}$ was divided by the representative lake bed area (based
on distances between adjacent measurement points) to calculate the specific groundwater flux or groundwater discharge per unit area of lakebed ($q_{gd}$, m d$^{-1}$). $q_{gd}$ were then integrated by distance offshore to estimate $Q_{gd}$ at discrete shoreline locations for comparison with $Q_{gd}$ estimated from the regional-scale survey. It is important to note that the $^{222}$Rn concentration may be impacted by the indirect groundwater discharge from Maskinonge River, thus $Q_{gd}$ were not calculated for the high-resolution survey near Keswick Beach.

$^{222}$Rn groundwater endmember samples were collected at fourteen beaches in the study areas to evaluate $^{222}$Rn concentrations in the groundwater endmember and spatial variability in these endmember concentrations. The beach sites include Oro Beach (OB), Johnson’s Beach (JB), Centennial Beach (CB), Minet’s Point Beach (MPB), Wilkin’s Beach (WKB), Willow Beach (WLB), Jackson’s Point Beach (JPB), Bayview Memorial Park Beach (BMB), Heritage Park Beach (HPB), Tollendal Beach (TB), beach at the end of Lockhart Road (LHB), beach at the end of 10th Road (10B), Innisfil Park Beach (IPB) and Paradise Beach (PB). The locations of all beaches are shown in Figure 4-1 with their coordinate locations provided in Table A5-1, Appendix 5. Sampling was conducted by installing temporary groundwater wells located 10 - 100 m landward of the shoreline. Groundwater samples were collected using a peristaltic pump and stored in airtight glass vials (40 ml or 250 ml). $^{222}$Rn concentrations in groundwater were determined using a RAD H$_2$O (Durridge Co., Inc.) connected to a RAD7 unit.

$C_{cw}$ was determined for all in-lake $^{222}$Rn sampling points (except those within 500 m of a creek mouth) by correcting the measured $^{222}$Rn concentration for $^{222}$Rn loss by atmospheric evasion, decay and $^{222}$Rn input by sediment diffusion. Following Burnett and Dulaiova (2003) and Dulaiova and Burnett (2006), $^{222}$Rn loss due to atmospheric evasion, ($J_{atm}$) was calculated by:

$$J_{atm} = k(C_{cw} - \alpha C_{air})$$  \hspace{1cm} (2)

where $k$ is the gas-transfer coefficient (m h$^{-1}$); $C_{air}$ is the $^{222}$Rn concentrations in the air (dpm L$^{-1}$) and $\alpha$ is the partitioning coefficient of $^{222}$Rn between water and air (dimensionless). $k$ was calculated by an empirical equation (MacIntyre et al., 1995):

$$k(600) = 0.45 \times u_{10}^{1.6} \times (Sc/600)^{-0.5}$$  \hspace{1cm} (3)
where $Sc$ is the Schmidt number and $u_{10}$ is the wind speed at a 10 m height above the water surface (km h$^{-1}$). The wind speed was obtained from the nearest Environment Canada weather station at Barrie (Environment Canada, 2015a). $\alpha$ was calculated by (Burnett and Dulaiova, 2003):

$$\alpha = 0.105 + 0.405 \exp(-0.05027T)$$  \hspace{1cm} (4)

where $T$ is the temperature of the water ($^\circ$C).

$^{222}$Rn loss due to decay ($J_{\text{decay}}$) was calculated by:

$$J_{\text{decay}} = z \lambda_{Rn} C_{cw}$$  \hspace{1cm} (5)

Sediment equilibration laboratory experiments were conducted to determine $^{222}$Rn inputs to the coastal water box from sediment diffusion ($J_{\text{diff}}$) (Corbett et al., 1998, Chanyotha et al., 2014). Here, sediment samples were collected from the bottom of the lake near the shoreline of JB, MPB, OB and JPB. Samples were sealed in a glass bottle with 250 mL of surface water from the collection site for more than 21 days to allow for equilibration before measuring the $^{222}$Rn that had diffused from the sediment to the water. $J_{\text{diff}}$ from the sediment was calculated by (Martens et al., 1980):

$$J_{\text{diff}} = (\lambda_{Rn} D_s)^{1/2}(C_{eq} - C_{cw})$$  \hspace{1cm} (6)

where $C_{eq}$ is the equilibrium concentration of $^{222}$Rn released from sediment (dpm L$^{-1}$) and $C_{cw}$ was the measured $^{222}$Rn concentration in the surface water where the sediment was collected (dpm L$^{-1}$). $D_s$ is the effective wet bulk sediment diffusion coefficient in sediments and was estimated by (Ullman and Aller, 1982):

$$-\log \frac{D_s}{\phi} = \left(\frac{980}{T}\right) + 1.59$$  \hspace{1cm} (7)

where $\phi$ is the porosity of the sediment samples and here $T$ is the overlying surface water temperature measured in the field (K).
4.3.3 Hydraulic gradient measurements

For comparison with the results from the $^{222}$Rn boat surveys, local groundwater discharge measurements were also performed from July - October 2015 at six beaches in the study area (shown in Figure 4-1). At each beach, a temporary transect of 4 - 6 groundwater monitoring wells were installed perpendicular to the shoreline to determine the horizontal hydraulic gradient. It was difficult to find suitable beaches for local-scale groundwater measurements as a large amount of shoreline in the study area is private land, and the presence of clay, silt sand and organic matters along much shoreline prevented manual installation of temporary monitoring wells. As sites where groundwater wells could be installed, groundwater levels were measured using a portable electronic water tape with an accuracy of ±3 mm (Heron Instruments Inc.) and the lake water level was measured using a stilling piezometer - transparent polycarbonate tube with measuring tape attached to the outside (Cartwright and Nielsen, 2001). The locations and elevations of all wells were surveyed using a total station (GTS-239W, Topcon Positioning System Inc.). $Q_{gd}$ was estimated from the measured groundwater levels at each site by applying Darcy’s Law:

$$Q_{gd} = -K \frac{dh}{dl}D$$  \hspace{1cm} (8)

where $D$ is the depth of the unconfined aquifer at the transect location (m); $h$ is the measured groundwater level at each location and $L$ is the distance between groundwater wells (m). $D$ for each site was determined from Ontario Ministry of the Environment Well Records (Ontario Ministry of Environment, 2015). $K$ was estimated for all well locations by performing grain size analysis on sediment samples collected from the base of all well during installation and applying the Hazen method (Hazen, 1911) (see Appendix 1). The horizontal hydraulic gradient $\left(\frac{dh}{dl}\right)$ used in Eqn. (8) was calculated as difference between the measured groundwater level at the furthest landward well and the water level at the shoreline.
4.4 Results and discussion

4.4.1 Regional-scale $^{222}$Rn survey results

In-lake $^{222}$Rn concentrations measured during the regional-scale boat surveys are shown in Figures 4-3, 4-4 and 4-5. $^{222}$Rn concentrations in the nearshore lake water in Kempenfelt Bay ranged from $0.65 \pm 0.07$ dpm L$^{-1}$ to $3.57 \pm 0.49$ dpm L$^{-1}$ with an average of $1.63 \pm 0.43$ dpm L$^{-1}$ for surveys conducted on June 9th and 11th (Figure 4-3). The highest in-lake $^{222}$Rn concentrations were observed near JB ($3.57 \pm 0.49$ dpm L$^{-1}$) and TB ($3.30 \pm 0.57$ dpm L$^{-1}$) during the survey completed on June 9th, 2015 (Figure 4-3 (a)). High $^{222}$Rn concentrations around TB are thought to be due to indirect groundwater discharge to Lovers Creek and Hewitts Creek. These creeks discharge to the lake near this location and are coldwater groundwater-fed streams with sustained summer base flows (Earthfx Inc., 2012). $^{222}$Rn concentration in Lovers Creek was measured to be $9.11 \pm 0.10$ dpm L$^{-1}$ (Table 4-2) which is nearly an order of magnitude higher than in-lake $^{222}$Rn concentrations. This result suggests that indirect groundwater discharge to these creeks may impact not only the stream water quality but also the water quality in the receiving lake. There are no creeks flow into the lake near JB suggesting that this area may be potential hotspot for direct groundwater discharge to the lake. While the $^{222}$Rn concentrations were lower near JB during the survey conducted on July 6th, 2015 (from $0.64 \pm 0.11$ dpm L$^{-1}$ to $2.03 \pm 0.21$ dpm L$^{-1}$ with an average of $1.00 \pm 0.17$ dpm L$^{-1}$), the highest $^{222}$Rn concentration on this survey day was still observed near JB ($2.03 \pm 0.21$ dpm L$^{-1}$, Figure 4-3 (a)). This survey extended north along the shoreline that represents the lake boundary for the Oro Creek South and Hawkestone subwatersheds. $^{222}$Rn concentration were relatively low along the shoreline of these subwatersheds despite the ESGRA water budget assessments estimating relatively high $Q_{gd}$ ($0.89 \text{ m}^3 \text{ m}^{-1} \text{ d}^{-1}$) to the lake from the Oro Creek South subwatershed (see Table 4-1) (Earthfx Inc., 2013).
Table 4-2: $^{222}\text{Rn}$ concentration and base flow index (BFI) measured in creeks discharging in Lake Simcoe in the study area. The error shown for the $^{222}\text{Rn}$ concentrations is the 2σ uncertainty. “/” means no data is available.

<table>
<thead>
<tr>
<th>Creek</th>
<th>Subwatershed</th>
<th>$^{222}\text{Rn}$ Concentrations (dpm L$^{-1}$)</th>
<th>Base flow index (BFI)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Dyment’s Creek</td>
<td>Barrie Creeks</td>
<td>$8.67 \pm 0.06$</td>
<td>0.66</td>
</tr>
<tr>
<td>Whiskey Creek</td>
<td>Barrie Creeks</td>
<td>$6.52 \pm 0.26$</td>
<td>0.66</td>
</tr>
<tr>
<td>Lovers Creek</td>
<td>Lovers Creeks</td>
<td>$9.11 \pm 0.10$</td>
<td>0.59</td>
</tr>
<tr>
<td>Shelswell Creek</td>
<td>Oro Creeks South</td>
<td>$10.14 \pm 0.15$</td>
<td>0.33</td>
</tr>
<tr>
<td>Hawkestone Creek</td>
<td>Hawkestone Creeks</td>
<td>$15.68 \pm 0.31$</td>
<td>0.44</td>
</tr>
<tr>
<td>Georgina Creek</td>
<td>Georgina Creeks</td>
<td>$7.89 \pm 0.09$</td>
<td>/</td>
</tr>
</tbody>
</table>

Figure 4-4 shows the in-lake $^{222}\text{Rn}$ concentrations from the boat surveys conducted along the shoreline in the Innisfil Area and western shore of Cooks Bay on July 8th and August 12th 2015. The in-lake $^{222}\text{Rn}$ concentrations were low along this shoreline on July 8th (from $0.66 \pm 0.06$ dpm L$^{-1}$ to $1.32 \pm 0.35$ dpm L$^{-1}$ with an average of $0.96 \pm 0.17$ dpm L$^{-1}$) and as well as on August 12th (from $0.55 \pm 0.12$ dpm L$^{-1}$ to $2.79 \pm 0.31$ dpm L$^{-1}$ with an average of $0.87 \pm 0.12$ dpm L$^{-1}$). Although high $^{222}\text{Rn}$ concentration was observed at one sampling location 2 km south of Big Bay Point ($2.79 \pm 0.31$ dpm L$^{-1}$) compared to all other measurements on August 12th, the $^{222}\text{Rn}$ concentration was not elevated here during the July 8th survey ($0.66 \pm 0.15$ dpm L$^{-1}$).

During our sampling, we found a small stream close to the high $^{222}\text{Rn}$ data point discharge into the lake which may contain high $^{222}\text{Rn}$ concentration. The low observed in-lake $^{222}\text{Rn}$ concentrations along this shoreline is consistent with the relatively low $Q_{gd}$ estimated from the ESGRA water budget calculations ($0.54$ m$^3$ m$^{-1}$ d$^{-1}$, Table 4-1).

The in-lake $^{222}\text{Rn}$ concentrations along the eastern shore of Cooks Bay and around the Georgina Area ranged from $0.67 \pm 0.14$ dpm L$^{-1}$ to $2.46 \pm 0.06$ dpm L$^{-1}$ with an average of $1.26 \pm 0.31$ dpm L$^{-1}$ on August 13th, and ranged from $0.65 \pm 0.05$ dpm L$^{-1}$ to $1.63 \pm 0.30$ dpm L$^{-1}$ with an average of $1.11 \pm 0.31$ dpm L$^{-1}$ on September 23rd (shown in Figure 4-5). The highest $^{222}\text{Rn}$ concentrations were observed around 2 km north of Keswick Beach ($2.46 \pm 0.06$ dpm L$^{-1}$) and approximately 4 km west of JPB ($2.22 \pm 0.17$ dpm L$^{-1}$) on August 13th (Figure 4-5 (a)). Higher $^{222}\text{Rn}$ concentrations compared to other measurement locations were also observed in these areas on September 23rd (Figure 4-5 (b)). The high in-lake $^{222}\text{Rn}$ concentrations north of
Keswick Beach are attributed to indirect groundwater discharge to Maskinonge River which flows into the lake near this location. BFI for Maskinonge River is relatively high (0.532) supporting the potential role of indirect groundwater discharge (GENIVAR Inc., 2013, Chu, 2011). High in-lake $^{222}$Rn concentrations west of JPB may be due to high direct groundwater discharge as there are no creeks discharging to the lake around this area.
Figure 4-3: In-lake $^{222}\text{Rn}$ concentrations from regional-scale survey conducted along the shore of Kempenfelt Bay on (a) June 9$^{th}$ (yellow numbers) and 11$^{th}$ (white numbers) 2015, and (b) July 6$^{th}$ (white) and 8$^{th}$ (yellow), 2015. The filled circles represent the measurement locations and their colour indicate the $^{222}\text{Rn}$ concentration range. The red box labeled (A) in (a) indicates the area where a high-resolution survey was completed on July 10$^{th}$, 2015.
Figure 4-4: In-lake $^{222}$Rn concentrations from regional-scale survey conducted along the shore of Innisfil Area on (a) July 8th and (b) August 12th, 2015. The filled circles represent the measurement locations with their colour indicating the $^{222}$Rn concentration range.
Figure 4-5: In-lake $^{222}\text{Rn}$ concentrations from regional-scale survey conducted along the shore of Georgina Area on (a) August 13$^{th}$, 2015 and (b) September 23$^{rd}$, 2015. The filled circles represent the measurement locations with their colour indicating the $^{222}\text{Rn}$ concentration range. The red box (B) and (C) in (a) indicates the area where high-resolution surveys were completed on September 24$^{th}$ and 25$^{th}$, 2015, respectively.
Groundwater samples for $^{222}$Rn were collected from fourteen beaches in the study areas. Figure 4-6 provides the locations of the beaches and a summary of the measured $^{222}$Rn groundwater concentrations. High spatial variability was observed between beach sites with highest $^{222}$Rn groundwater concentrations observed at LHB ($748 \pm 40$ dpm L$^{-1}$), BMB ($383 \pm 53$ dpm L$^{-1}$), and JB ($217 \pm 20$ dpm L$^{-1}$) (all data is provided in Table A5-1, Appendix 5). Considering all measurements, the average $^{222}$Rn groundwater concentration was $132 \pm 19$ dpm L$^{-1}$ which was nearly 100 times higher than the in-lake concentrations. At a given site, $^{222}$Rn concentrations were generally higher in the shallow groundwater close to the shoreline but additional sampling is required to confirm this and determine an appropriate endmember concentration to use for mass balance calculations. To evaluate temporal variability in the $^{222}$Rn groundwater end-member concentrations, multiple sampling on different dates was conducted at JB, CB, MPB and WKB. $^{222}$Rn groundwater concentrations in CB and WKB were quite consistent between the sampling events. However, although the groundwater end-member samples were collected only 1-3 weeks apart, $^{222}$Rn concentrations in the groundwater varied considerably at JB and MPB (Table A5-1, Appendix 5). It is unclear if this variability is due to high spatial heterogeneity of temporal variability. Sediment equilibration experiments conducted using sediment collected near the shoreline at JB and MPB found the sediment equilibrium concentrations ($C_{eq}$) to be around 2.75 dpm L$^{-1}$. These concentrations were used in the mass balance calculations (Eqn. 6).

High $^{222}$Rn concentrations were found in the six creeks sampled in the study area (Table 4-2). The high $^{222}$Rn concentrations are likely caused by groundwater discharge to the streams, which is consistent with these stream being coldwater groundwater-fed streams (AquaResource Inc., 2013a, Earthfx Inc., 2013, GENIVAR Inc., 2013). BFI are generally high for the creeks in study subwatersheds indicating groundwater input is important in these creeks (Table 4-1). For instance, in the Georgina Area, it is estimated only 7% of the groundwater directly discharges into Lake Simcoe, while approximately 72% of the groundwater discharge is to streams and wetlands (GENIVAR Inc., 2013).
Groundwater discharge rates ($Q_{gd}$) calculated using Eqn. (1) for all in-lake $^{222}$Rn measurements, except for those within 500 m of the creek mouth, are provided in Figures 4-7, 4-8 and 4-9. The input parameters for the calculations are provided in Table A6-1, A6-2 and A6-3 in Appendix 6. In contrast to the distribution of $^{222}$Rn concentrations, highest $Q_{gd}$ was found 3 km east of Shanty Bay (sampling point 4 and 5 in Figure 4-7 (a)). While high $^{222}$Rn concentrations were not observed here, the high $Q_{gd}$ is due to the relatively low groundwater end-member value for OB (Table A5-1, Appendix 5). High $Q_{gd}$ were also estimated at discrete locations along the south shore of Kempenfelt Bay (sampling point 15-18 in Figure 4-7 (a)). Here, the high $Q_{gd}$ was consistent with the high $^{222}$Rn concentrations measured along this shoreline. $Q_{gd}$ calculated for sampling points along the shorelines in the Innisfil Area and Georgina Area were all relatively low with slightly higher $Q_{gd}$ estimated near JPB and the north of Keswick Beach.

**Figure 4-6:** Groundwater endmember sampling locations (yellow dots) with the range of measured $^{222}$Rn concentrations (dpm L$^{-1}$) shown in brackets. All data is provided in Table A5-1, Appendix 5.
Evasion of $^{222}\text{Rn}$ to the atmosphere ($J_{\text{atm}}$) represented the largest loss of $^{222}\text{Rn}$ from the nearshore water column and thus in the mass balance model (Table A6-1, A6-2, A6-3 and A6-4, Appendix 6). The atmospheric evasion typically accounted for 50-60% of the total $^{222}\text{Rn}$ losses with this loss affected by wind speed and water temperature (Eqn. (2-4)). Inaccurate estimation of this loss term may lead to some uncertainty in the estimated $Q_{gd}$. Furthermore, determining an appropriate value for $C_{gw}$ remains a challenge for accurate calculation of $Q_{gd}$ from $^{222}\text{Rn}$ survey data due to large spatial variability in $^{222}\text{Rn}$ groundwater concentrations along the shoreline. Burnett et al. (2007) suggested that the groundwater samples collected from shallow monitoring wells near the shore are most representative of the groundwater endmember. In our calculation, we used concentrations for groundwater samples collected from the most nearshore monitoring wells at a given site as groundwater end-members values. These samples generally had higher $^{222}\text{Rn}$ concentrations compared to groundwater samples collected further onshore. Estimated $J_{\text{diff}}$ were less than 10 dpm$^{-2}$ m d$^{-1}$ at all locations which represents only a minor component in the mass balance calculation (less than 3% of losses).
Figure 4-7: In-lake $Q_{gd}$ from regional-scale survey conducted along the shore of Kempenfelt Bay on (a) June 9\textsuperscript{th} (yellow) and 11\textsuperscript{th} (white) 2015, and (b) July 6\textsuperscript{th} (white) and 8\textsuperscript{th} (yellow), 2015. The filled circles represent the measurement locations and their colour indicate the $Q_{gd}$ range. All data is provided in Table A6-1, Appendix 6.
Figure 4-8: In-lake $Q_{ed}$ from regional-scale survey conducted along the shore of Innisfil Area on (a) July 8th and (b) August 12th, 2015. The filled circles represent the measurement locations with their colour indicates the $Q_{ed}$ range. All data is provided in Table A6-2, Appendix 6.
Figure 4-9: In-lake $Q_{gd}$ from regional-scale survey conducted along the shore of Georgina Area on (a) August 13$^{th}$, 2015 and (b) September 23$^{rd}$, 2015. The filled circles represent the measurement locations with their colour indicating the $Q_{gd}$. All data is provided in Table A6-3, Appendix 6.
4.4.2 High-resolution $^{222}$Rn survey results

Results from the high-resolution surveys conducted around JB (red box A in Figures 4-3 (a)), Keswick Beach (red box B in Figure 4-5 (a)) and JPB (red box B in Figure 4-5 (a)) are shown in Figure 4-10. The survey results show in all three areas $^{222}$Rn concentrations were highest near the shoreline with concentrations generally decreasing offshore. High $^{222}$Rn concentrations were again found higher near JB ($2.09 \pm 0.41$ dpm L$^{-1}$) and JPB ($2.21 \pm 0.26$ dpm L$^{-1}$) on these high-resolution survey dates. The decreasing offshore trend in concentrations indicates that $^{222}$Rn in the areas is delivered to the lake via nearshore groundwater discharge rather offshore groundwater discharge that may occur where the lake bed intercepts deeper confined aquifer units. In Figure 4-10 (b), we can still observe high $^{222}$Rn concentration ($2.24 \pm 0.27$ dpm L$^{-1}$) in the sampling point which is far from the shoreline but close to the river mouth. This high value may be contributed from the flow of Maskinonge River with indirect groundwater discharge.
Figure 4-10: In-lake $^{222}$Rn concentrations measured in high-resolution surveys conducted in (a) Area A on July 10th, (b) Area B on September 24th and (c) Area C on September 25th. The filled circles represent the measurement locations with their colour indicating the $^{222}$Rn concentration range. All data is provided in Table A6-4, Appendix 6.
The $^{222}$Rn mass balance model was applied to estimate direct specific groundwater flux ($q_{gd}$) in high-resolution survey areas A and C (Figure 4-11). The mass balance model was not applied in area B where the high in-lake $^{222}$Rn concentrations are thought to be due to indirect groundwater discharge. Consistent with the $^{222}$Rn data, the highest $q_{gd}$ was found near the shoreline with $q_{gd}$ decreasing offshore. For Area A, $q_{gd}$ near JB (sampling point 1-2) was the highest ($4.38 \pm 0.97$ m d$^{-1}$). For Area C, the highest $q_{gd}$ ($2.92 \pm 0.24$ m d$^{-1}$) were measured at the sampling point to the east of JPB (sampling point 1-4). Sampling points 1-2, 2-2 and 3-2 in Figure 4-11 (a) and sampling points 1-2, 2-2 and 3-2 in Figure 4-11 (b) were used to calculate $Q_{gd}$ near JB and JPB, respectively, for comparison with $Q_{gd}$ estimated from the regional-scale survey data. The estimated $Q_{gd}$ are $2.92 \pm 0.65$ m$^3$ m$^{-1}$ d$^{-1}$ and $1.97 \pm 0.18$ m$^3$ m$^{-1}$ d$^{-1}$ for JB and JPB, respectively. $Q_{gd}$ estimated along these transects using the high-resolution data ($2.92 \pm 0.65$ m$^3$ m$^{-1}$ d$^{-1}$ and $1.97 \pm 0.18$ m$^3$ m$^{-1}$ d$^{-1}$) were slightly higher than that calculated from regional-scale survey ($2.45 \pm 0.54$ m$^3$ m$^{-1}$ d$^{-1}$ in June 9th for JB, and $1.26 \pm 0.15$ m$^3$ m$^{-1}$ d$^{-1}$ in September 23rd for JPB, respectively). This difference is because, $Q_{gd}$ estimated using the regional-scale survey only included groundwater discharge close to the shore, whereas $Q_{gd}$ calculated using the high-resolution data also included the lower groundwater discharge occurring further offshore.
Figure 4-11: In-lake $q_{gd}$ measured in high-resolution surveys conducted in (a) Area A on July 10th and (b) Area C on September 25th. The filled circles represent the measurement locations with their colour indicating the $q_{gd}$ range.

4.4.3 Hydraulic gradient measurements

Table 4-2 provides the measured horizontal hydraulic gradients and input parameters used for calculating $Q_{gd}$ at six beach sites in the study area (locations shown in Figure 4-1). The measured groundwater levels indicate that groundwater was flowing into Lake Simcoe
from beach sites at the measurement times. Repeated measurements were performed at JB, MPB and WKB to evaluate temporal variability showed only small differences in the hydraulic gradient (~0.001). The range of $Q_{gd}$ calculated for each site was large due to large variation in $K$ determined from sediment samples collected from individual groundwater wells at each beach (Table 4-3). The highest $Q_{gd}$ was estimated at OB ($2.93 - 6.31 \text{ m}^3 \text{ m}^{-1} \text{ d}^{-1}$), relatively high $Q_{gd}$ estimated from regional-scale $^{222}\text{Rn}$ survey were also measured in the nearshore water around 2 km north of OB (Figure 4-7(b)). The ESRGA assessment also shows relatively high $Q_{gd}$ from the Oro Creeks South subwatershed (Table 4-1). The thickness of surficial aquifer near JB (8.5 m) was larger than the other beaches, but as $K$ was quite small, estimated $Q_{gd}$ ($0.86 - 1.26 \text{ m}^3 \text{ m}^{-1} \text{ d}^{-1}$) in May 2015 was also small compared to other sites. This is also in contrast to the relatively high in-lake $^{222}\text{Rn}$ concentrations measured near JB. To reduce uncertainties in $K$, it is recommended that pumping tests or slug tests be conducted at the sites. Furthermore, there is also a need to reduce uncertainty in the depth of the surficial aquifer at all beach sites. The groundwater discharge patterns were quite complex for MPB and WKB as there are many coldwater creeks in the subwatersheds (see Figure 4-2) which may increase the $^{222}\text{Rn}$ concentrations in the lake water and the creek may incise the surficial aquifer and capture the groundwater flowing towards the lake leading to smaller horizontal hydraulic gradient near the shore and lower direct groundwater discharge. $Q_{gd}$ calculated using $^{222}\text{Rn}$ and hydraulic gradient measurement methods as well as the results from ESGRA assessment at WLB and JPB match well with each other (Table 4-1 and 4-3, Figure 4-9).
Table 4-3: Input parameter values and results for estimation of $Q_{gd}$ based on groundwater level measurements at six beach sites (see Figure 4-1 for locations).

<table>
<thead>
<tr>
<th>Beach name</th>
<th>Measurement date</th>
<th>$\frac{dh}{dL}$</th>
<th>$D$ (m)</th>
<th>$K$ (m d$^{-1}$)</th>
<th>$Q_{gd}$ (m$^3$ m$^{-1}$ d$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>OB</td>
<td>28/07/2015</td>
<td>0.015</td>
<td>2.44</td>
<td>80 - 173</td>
<td>2.93 - 6.31</td>
</tr>
<tr>
<td>JB</td>
<td>28/05/2015</td>
<td>0.008</td>
<td>8.53</td>
<td>13 - 19</td>
<td>0.86 – 1.26</td>
</tr>
<tr>
<td></td>
<td>16/06/2015</td>
<td>0.007</td>
<td></td>
<td></td>
<td>0.75 – 1.11</td>
</tr>
<tr>
<td>MPB</td>
<td>08/06/2015</td>
<td>0.004</td>
<td>4.57</td>
<td>58 - 84</td>
<td>1.05 – 1.54</td>
</tr>
<tr>
<td></td>
<td>15/06/2015</td>
<td>0.003</td>
<td></td>
<td></td>
<td>0.79 – 1.16</td>
</tr>
<tr>
<td>WKB</td>
<td>10/06/2015</td>
<td>0.019</td>
<td>0.91</td>
<td>28 - 103</td>
<td>0.48 – 1.80</td>
</tr>
<tr>
<td></td>
<td>18/06/2015</td>
<td>0.020</td>
<td></td>
<td></td>
<td>0.51 – 1.89</td>
</tr>
<tr>
<td>WLB</td>
<td>04/10/2015</td>
<td>0.013</td>
<td>0.61</td>
<td>65 - 91</td>
<td>0.51 – 0.72</td>
</tr>
<tr>
<td>JPB</td>
<td>04/10/2015</td>
<td>0.010</td>
<td>1.22</td>
<td>30 - 93</td>
<td>0.37 – 1.13</td>
</tr>
</tbody>
</table>

4.4.4 Factors affecting temporal variability of in-lake $^{222}$Rn concentrations

$^{222}$Rn concentrations along the surveyed shoreline were observed to vary temporally between sampling dates. The $^{222}$Rn inventory in the lake and thus in-lake $^{222}$Rn concentrations are influenced by various $^{222}$Rn inputs and outputs which vary over time. Loss of $^{222}$Rn to the atmosphere represented the large loss term in the mass balance calculations, see Section 4.3.2) and therefore variability in this output flux ($J_{atm}$) is expected to have contributed to the observed temporal variability. $J_{atm}$ is mainly governed by the wind speed and water temperature (see Eqn. (2-4)). It is possible the in-lake $^{222}$Rn concentrations are also affected by precipitation and associated creek discharge which may dilute nearshore $^{222}$Rn concentrations. Alternatively, the offshore distance of the measurement location on a given day may also influence the measured $^{222}$Rn concentration (concentrations decrease with offshore distance, see Section 4.4.2).

$^{222}$Rn concentrations near JB and near Shanty Bay measured on different days are shown in Figure 4-12 together with the wind speed, precipitation, water temperature and offshore distance for the measurement. With the preliminary data available, it seems temporal variability in $^{222}$Rn concentrations may be due to varying wind speed and, to a lesser extent,
precipitation. Correlations between the 12-hour wind speed and precipitation are shown in Figure 4-13 with wind speed showing an inverse relationship with the in-lake $^{222}\text{Rn}$ concentrations. These results are consistent with Burnett and Dulaiova (2006) who observed $^{222}\text{Rn}$ inventories in coastal waters of Donnalucata, Sicily to decrease in response to high winds (10 m s$^{-1}$). $^{222}\text{Rn}$ inventories were also found to considerably decrease during a storm during a case study in Dor Beach, Israel (Burnett et al., 2007). Additional higher frequency measurements at select sites are required to confirm the cause of the temporal variability in $^{222}\text{Rn}$ in-lake concentrations. Understanding the temporal variability is essential for being able to select sampling days that will provide optimum conditions for the survey as well as to compare regional-scale $^{222}\text{Rn}$ survey data from different survey dates.
Figure 4-12: $^{222}$Rn concentrations at JB (a) and Shanty Bay (b) on different sampling days compared with precipitation, wind speed and distance offshore.
Figure 4-13: Relationship between $^{222}\text{Rn}$ concentrations and (a) wind speed (average of last 12 hrs) and (b) precipitation (average of last 12 hrs) at JB on different sampling days during June and July, 2015.

4.5 Conclusions

Groundwater discharge may be an important pathway for delivering pollutants to Lake Simcoe, however, this pathway is poorly quantified. Further, there are limited field methods currently available to quantify groundwater discharge into large inland lakes. The objective of this research was to evaluate the use of $^{222}\text{Rn}$ as a tracer for evaluating regional-scale groundwater discharge and to use this tool to identify hotspots for direct groundwater discharge into Lake Simcoe. Regional-scale $^{222}\text{Rn}$ surveys were conducted along 80 km of shoreline in the west and eastern parts of Lake Simcoe around Kempenfelt Bay, Innisfil Area and Georgina Area. High in-lake $^{222}\text{Rn}$ concentrations were observed near JB, JPB, TB and Keswick Beach. High in-lake $^{222}\text{Rn}$ concentrations near JB and JPB are thought to be due to direct nearshore groundwater discharge, however, high $^{222}\text{Rn}$ concentrations near TB and Keswick Beach are likely associated with indirect groundwater discharge into creeks that enter the lake near these locations. The high in-lake $^{222}\text{Rn}$ concentrations at the latter locations reveal that indirect groundwater discharge affects the water quality in the receiving lake. Direct groundwater discharge rates ($Q_{gd}$) calculated using all $^{222}\text{Rn}$ survey
data, except for sampling points located within 500 m of creek mouths, indicated potential direct groundwater discharge hotspots. Selection of appropriate groundwater endmember $^{222}\text{Rn}$ concentrations along the surveyed shoreline introduced considerable uncertainty in the $Q_{gd}$ estimates and it is recommended further field work is conducted to address this uncertainty. Finally, in-lake $^{222}\text{Rn}$ concentrations were found to vary temporally between survey days. While preliminary analysis suggests that this variability is due to varying wind speed and, to a less extent, precipitation, additional high temporal resolution $^{222}\text{Rn}$ in-lake sampling is required to confirm this and determine a quantitative relationship that can account for the temporal variability in $^{222}\text{Rn}$ concentrations.
4.6 Reference


Chapter 5

5 Summary and recommendations

5.1 Summary

Groundwater discharge may be an important pathway for delivering pollutants including nutrients, metals, organic contaminants and chloride into large inland lakes such as the Great Lakes and Lake Simcoe. This pathway, however, is poorly understood. In this thesis, field work was conducted along 17 km of shoreline of Nottawasaga Bay and 80 km of shorelines of Lake Simcoe using multiple groundwater discharge field methods including $^{222}\text{Rn}$ boat surveys, electrical resistivity tomography (ERT) surveys, vertical temperature profiling, and vertical and horizontal hydraulic gradients measurements. Through the field work and data analyses, this thesis aimed to address three distinct research objectives.

The first objective focused on the assessment of suitable field techniques for quantifying groundwater discharge into large inland lakes at different spatial scales (regional- and local-scale). The naturally-occurring tracer $^{222}\text{Rn}$ was found to be suitable for identifying, at the regional-scale, shoreline areas with potential higher groundwater discharge. A steady-state $^{222}\text{Rn}$ mass balance model was applied in all sampling points (except for those near the creek mouth) to estimate groundwater discharge rates per m of shoreline. An ERT survey conducted simultaneously with the regional-scale $^{222}\text{Rn}$ survey in Nottawasaga Bay provided insight into surficial geological variability that may be associated with the observed groundwater discharge hotspots. High spatial resolution $^{222}\text{Rn}$ surveys as well as vertical temperature profiling and vertical gradient measurements at specific shoreline locations in the surveyed area indicated that highest groundwater discharge rates are found close to the shoreline with discharge decreasing offshore. This result suggests that the groundwater discharge quantified is from the surficial aquifer rather than deeper confined aquifers that may intercept the lakebed further offshore. Groundwater discharge rates calculated from different field methods showed similar trends with discrepancies between methods associated with uncertainties with input parameters such as hydraulic conductivity.
and surficial aquifer depth. The combination of field methods adopted was useful for understanding the groundwater discharge patterns at the regional- as well as the local-scale.

The second objective was to evaluate the spatial pattern and quantity of groundwater discharge along shorelines of large inland lakes in Southern Ontario and link observed discharge groundwater patterns to hydrogeological characteristics of the nearshore area. $^{222}$Rn concentrations and groundwater discharge rates along the survey shoreline exhibited high variability likely due to the varying hydrogeological conditions along the shoreline. Hot spots for direct groundwater discharge were identified near Lafontaine Beach and the beach at the end of Concession Road 15 in Nottawasaga Bay, as well as near Johnson’s Beach and Jackson’s Point Beach in Lake Simcoe. Indirect groundwater discharge hot spots were also found near Lafontaine Creek in Nottawasaga Bay, as well as Lovers Creek and Maskinonge Creek in Lake Simcoe where coldwater streams discharging into Lake Simcoe. Groundwater discharge hotspots were normally found in locations where topography landward of the shoreline is steep and the unconfined saturated aquifer is thick and highly permeable (mainly medium sands and rounded gravel). Large groundwater discharge around Lafontaine Beach and the beach at the end of Concession Road 15 may be attribute to a tunnel valley aquifer system (sand and gravel) that outcrops at Nottawasaga Bay around this location.

The third objective was to evaluate the factors contributing to temporal variability of $^{222}$Rn concentrations in the lake water to reduce uncertainties in this measurement techniques and enable regional-scale $^{222}$Rn surveys from different days to be compared. Preliminary analysis of in-lake $^{222}$Rn concentrations measured at the same locations on different days in Lake Simcoe indicated that wind speed and, to a less extent, precipitation influences the in-lake $^{222}$Rn concentrations. Steady state mass balance calculations indicate that the main loss of $^{222}$Rn from the lake water column was atmospheric evasion, and the wind speed considerably affects this loss term.
5.2 Recommendations

Chapter 3 of this thesis evaluated different groundwater discharge methods along a 17 km stretch of shoreline in Nottawasaga Bay. Key recommendations for improving the field techniques and groundwater discharge ($Q_{gd}$) estimates are as follows:

- The large uncertainties in laboratory-determined $K$ values as well as the difficulties in determining the $D$ for nearshore surficial aquifers introduced large uncertainty in $Q_{gd}$ for the hydraulic gradient measurement approaches. It is recommended that pumping tests or slug tests be conducted at each beach site to more accurately determine the aquifer hydraulic conductivity as required to constrain $Q_{gd}$ estimates using the hydraulic gradient measurement approaches.

- Difficulty in manually installing equipment at beach sites with layers of clay or gravel at shallow depths near the shoreline limited use of the vertical hydraulic gradient and vertical temperature profiling methods at all beach sites. While these methods can provide valuable information on local-scale groundwater discharge patterns, the techniques used require improvement so they can be used along shorelines with gravel and cobble sediment.

- Assumptions adopted for the $^{222}$Rn mass balance calculations (i.e. neglecting offshore mixing, production from $^{226}$Ra and sediment diffusion) may also have caused uncertainties in $Q_{gd}$ calculated from the $^{222}$Rn survey data. It is recommended that additional data including $^{226}$Ra in the water and sediment, and $^{222}$Rn sediment diffusion is collected to confirm the validity of these assumption for the studied shorelines.

In Chapter 4 of this thesis $^{222}$Rn was successfully applied as a tracer to identify two hotspots for direct groundwater discharge as well as two hotspots for indirect groundwater discharge (i.e. groundwater discharge into a creek with subsequent discharge to the lake). Recommendations for improving evaluation of groundwater discharge to Lake Simcoe include:
• $^{222}$Rn concentrations in groundwater end-member varied considerably and this concentration needs to be better constrained for the $^{222}$Rn mass balance calculations. Additional sediment equilibrium experiments with sediment collected from the surface of the lake bed near the shore is recommended to determine an appropriate groundwater end-member.

• A better understanding of temporal variabilities of the in-lake $^{222}$Rn concentrations including the influence of wind speed and precipitation is required to reduce uncertainties in applying the steady-state mass balance model to estimate $Q_{gd}$, as well as better compare regional-scale $^{222}$Rn surveys from different days. It is recommended to conduct a time series stationary monitoring to quantify the factors controlling the temporal variabilities of the in-lake $^{222}$Rn concentrations.

• Additional analysis is required to better understand the link between the spatial distribution of $^{222}$Rn concentrations in the lake water as well as groundwater discharge rates relative to the varying hydrogeological conditions around Lake Simcoe.

• Evaluation of the groundwater-lake interactions including geochemical cycling in nearshore area that have been identified as discharge hot spots is required to better understand the contribution of groundwater discharge to the water quality issues in Lake Simcoe.
Appendices

Appendix 1: Estimation of hydraulic conductivity

Calculation of the groundwater discharge rate \( Q_{gd} \) is based on Darcy’s Law which requires knowledge of the aquifer hydraulic conductivity \( K \) (Eqn. (6) in Chapter 2). In this study, \( K \) for the different beach sites was determined by grain size analysis and application of Hazen method (Eqn. (A1-1)) (Hazen, 1911). Sediment samples were collected from the bottom of each monitoring well upon installation. Sediment samples were dried in an oven at 110 °C for about 10 hours. Approximately 900 g dry sediment passed through 8 different sized sieves placed on a vibration machine for 15 min. Grain size number used in our analysis followed the ASTM (American Society for Testing and Materials) Standards (ASTM International, 2013):

<table>
<thead>
<tr>
<th>ASTM Grain Size Number</th>
<th>Grain Diameter (mm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>4</td>
<td>4.760</td>
</tr>
<tr>
<td>10</td>
<td>2.000</td>
</tr>
<tr>
<td>20</td>
<td>0.850</td>
</tr>
<tr>
<td>40</td>
<td>0.420</td>
</tr>
<tr>
<td>60</td>
<td>0.250</td>
</tr>
<tr>
<td>100</td>
<td>0.150</td>
</tr>
<tr>
<td>140</td>
<td>0.106</td>
</tr>
<tr>
<td>200</td>
<td>0.075</td>
</tr>
</tbody>
</table>

The sediment in each sieve was weighed to calculate a cumulative weight percent and the grain size distribution curve. The grain size distribution curve can be used to calculate \( K \):

\[
K = C d_{10}^2
\]  
(A1-1)
where $d_{10}$ is the effective grain size (mm) which is the diameter of the final 10% of sediment that passes through the sieves; $C$ is a coefficient that factors in the sorting characteristics of the sediment, the value depends on how sorted the sediment is.

Grain size distribution curves were determined for each sample to estimate $d_{10}$. Figure A1-1 provides an example of the sediment collected from BB.

Figure A1-1: Grain size distribution graph for the sediment samples collected at BB. The red line represents the effective grain size of the final 10% of sediment that passes through the sieves ($d_{10}$).
Appendix 2: $^{222}$Rn survey results in Nottawasaga Bay

Table A2-1: Summary of input parameter values and flux components in the $^{222}$Rn mass balance calculations for high-resolution survey. The numbers for the sampling points are shown in Figure 3-8. The uncertainty for the $^{222}$Rn concentrations represent the 2σ uncertainty.

<table>
<thead>
<tr>
<th>No.</th>
<th>$C_{gw}$ (dpm L$^{-1}$)</th>
<th>$C_w$ (dpm L$^{-1}$)</th>
<th>$z$ (m)</th>
<th>$J_{atm}$ (dpm m$^2$ d$^{-1}$)</th>
<th>$J_{decay}$ (dpm m$^2$ d$^{-1}$)</th>
<th>$q_{gd}$ (10$^{-2}$ m d$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>199 ± 49</td>
<td>0.14 ± 0.10</td>
<td>2</td>
<td>179</td>
<td>50</td>
<td>0.14 ± 0.03</td>
</tr>
<tr>
<td>2</td>
<td>199 ± 49</td>
<td>0.85 ± 0.24</td>
<td>1</td>
<td>1153</td>
<td>153</td>
<td>0.73 ± 0.18</td>
</tr>
<tr>
<td>3</td>
<td>199 ± 49</td>
<td>0.92 ± 0.25</td>
<td>1</td>
<td>1255</td>
<td>166</td>
<td>0.79 ± 0.26</td>
</tr>
<tr>
<td>4</td>
<td>199 ± 49</td>
<td>3.12 ± 0.47</td>
<td>3</td>
<td>4294</td>
<td>1695</td>
<td>3.85 ± 0.93</td>
</tr>
<tr>
<td>5</td>
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<td>1571</td>
<td>3.56 ± 0.86</td>
</tr>
<tr>
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<td>598</td>
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<tr>
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<td>2536</td>
<td>334</td>
<td>1.60 ± 0.39</td>
</tr>
<tr>
<td>8</td>
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<td>1</td>
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<td>180</td>
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</tr>
<tr>
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</tr>
<tr>
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<td>2327</td>
<td>307</td>
<td>1.47 ± 0.35</td>
</tr>
<tr>
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<td>1.34 ± 0.31</td>
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<td>1829</td>
<td>483</td>
<td>1.40 ± 0.34</td>
</tr>
<tr>
<td>12</td>
<td>199 ± 49</td>
<td>1.47 ± 0.32</td>
<td>3</td>
<td>2020</td>
<td>800</td>
<td>1.81 ± 0.44</td>
</tr>
<tr>
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</tr>
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<td>436</td>
<td>1.26 ± 0.30</td>
</tr>
<tr>
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</tr>
<tr>
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<td>309</td>
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</tr>
<tr>
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<td>668</td>
<td>358</td>
<td>0.69 ± 0.17</td>
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<tr>
<td>19</td>
<td>212 ± 51</td>
<td>0.99 ± 0.26</td>
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<td>1351</td>
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<td>1.21 ± 0.29</td>
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<tr>
<td>20</td>
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<td>956</td>
<td>382</td>
<td>0.86 ± 0.21</td>
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<td>21</td>
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<td>763</td>
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<td>0.69 ± 0.17</td>
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<tr>
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<td>573</td>
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<tr>
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<td>475</td>
<td>256</td>
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</tr>
<tr>
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<td>477</td>
<td>322</td>
<td>0.56 ± 0.14</td>
</tr>
<tr>
<td>25</td>
<td>212 ± 51</td>
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<td>386</td>
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<td>279</td>
<td>230</td>
<td>0.37 ± 0.09</td>
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<td>665</td>
<td>534</td>
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</table>
Table A2-2: Summary of input parameter values and flux components in the $^{222}$Rn mass balance calculations for regional-scale survey. The uncertainty for the $^{222}$Rn concentrations represent the $2\sigma$ uncertainty.

<table>
<thead>
<tr>
<th>Sampling sites</th>
<th>$C_{gw}$ (dpm L$^{-1}$)</th>
<th>$C_w$ (dpm L$^{-1}$)</th>
<th>$z$ (m)</th>
<th>$I_{atm}$ (dpm m$^2$ d$^{-1}$)</th>
<th>$I_{decay}$ (dpm m$^2$ d$^{-1}$)</th>
<th>$Q_{gd}$ (m$^3$ m$^{-1}$ d$^{-1}$)</th>
</tr>
</thead>
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<tr>
<td>C12</td>
<td>77 ± 22</td>
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<td>3</td>
<td>453</td>
<td>183</td>
<td>0.99 ± 0.29</td>
</tr>
<tr>
<td>WB</td>
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<td>0.61 ± 0.20</td>
<td>3</td>
<td>831</td>
<td>333</td>
<td>1.01 ± 0.24</td>
</tr>
<tr>
<td>C15</td>
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<td>930</td>
<td>5.51 ± 1.33</td>
</tr>
<tr>
<td>LB</td>
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<td>1.71 ± 0.35</td>
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<td>1683</td>
<td>668</td>
<td>4.54 ± 1.10</td>
</tr>
<tr>
<td>BB</td>
<td>115 ± 22</td>
<td>0.84 ± 0.24</td>
<td>3</td>
<td>1144</td>
<td>455</td>
<td>1.16 ± 0.28</td>
</tr>
<tr>
<td>MVB</td>
<td>178 ± 30</td>
<td>0.57 ± 0.19</td>
<td>1</td>
<td>771</td>
<td>103</td>
<td>0.51 ± 0.07</td>
</tr>
</tbody>
</table>
Appendix 3: Sensitivity analysis for $^{222}\text{Rn}$ atmosphere evasion

A sensitivity analysis was performed to evaluate the input values used for wind speed ($u_{10}$) in the calculation of $^{222}\text{Rn}$ atmospheric evasion ($J_{\text{atm}}$).

Average 24-hr wind speed prior to the survey period was used to calculate $J_{\text{atm}}$ for the results shown in Chapter 3. This wind-speed value was 8.17 km h$^{-1}$. Additional calculations were performed using the average wind speed for 12-hr (7.46 km h$^{-1}$), 2 d (7.13 km h$^{-1}$) and 5 d (7.21 km h$^{-1}$) prior to the survey period. The additional calculations were performed for three sampling locations with high (location 4, 5.11 ± 1.23 m$^3$ m$^{-1}$ d$^{-1}$), medium (location 14, 1.99 ± 0.48 m$^3$ m$^{-1}$ d$^{-1}$) and low (location 1, 0.15 ± 0.04 m$^3$ m$^{-1}$ d$^{-1}$) calculated $Q_{gd}$ (see Figure 3-8). The results of the sensitivity analyses are provided in Table A3-1. $k_1$, $J_{\text{atm}1}$ and $Q_{gd1}$ represents the situation under $u_{10} = 7.46$ km h$^{-1}$; $k_2$, $J_{\text{atm}2}$, and $Q_{gd2}$ represents the situation under $u_{10} = 7.13$ km h$^{-1}$; $k_3$, $J_{\text{atm}3}$, and $Q_{gd3}$ represents the situation under $u_{10} = 7.21$ km h$^{-1}$. The variations in $u_{10}$ resulted in a change in $Q_{gd}$ of up to 10%, 17% and 16% for the three sampling points, respectively.

**Table A3-1: The sensitivity analysis results for different $u_{10}$**

<table>
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<th>Sampling site</th>
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<th>$k_1$</th>
<th>$k_2$</th>
<th>$k_3$</th>
<th>$J_{\text{atm}}$</th>
<th>$J_{\text{atm}1}$</th>
<th>$J_{\text{atm}2}$</th>
<th>$J_{\text{atm}3}$</th>
<th>$Q_{gd}$</th>
<th>$Q_{gd1}$</th>
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<td>1.10</td>
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<td>4.60 ± 1.10</td>
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<td></td>
<td></td>
<td>1646</td>
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<td>1312</td>
<td>1324</td>
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<td>1.79 ± 0.43</td>
<td>1.65 ± 0.40</td>
<td>1.67 ± 0.40</td>
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<td>0.15 ± 0.04</td>
<td>0.14 ± 0.04</td>
<td>0.13 ± 0.03</td>
<td>0.13 ± 0.03</td>
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Appendix 4: Upland till landforms and tunnel channel aquifers

Figure A4-1: Upland till landforms and tunnel channel aquifers in Midland and Penetanguishene Area (Golder Associates Ltd., 2014). The black dots show the locations where local-scale groundwater discharge measurements were conducted.
Appendix 5: $^{222}$Rn concentrations in groundwater endmembers along Lake Simcoe

Table A5-1: $^{222}$Rn concentrations in groundwater endmembers for all of the sampling points

<table>
<thead>
<tr>
<th>Sampling sites</th>
<th>Latitude (N)</th>
<th>Longitude (W)</th>
<th>Number of wells</th>
<th>$C_{gw}$ (dpm L$^{-1}$)</th>
<th>Measurement Date</th>
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<td></td>
<td></td>
<td></td>
<td>Min</td>
<td>Max</td>
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<td>63 ± 5</td>
<td>90 ± 6</td>
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<td>79°39'27.61&quot;</td>
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<td></td>
<td></td>
<td></td>
<td></td>
<td>3 83 ± 21</td>
<td>135 ± 39</td>
</tr>
<tr>
<td>CB</td>
<td>44°22'45.34&quot;</td>
<td>79°41'20.17&quot;</td>
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<td>67 ± 21</td>
<td>70 ± 19</td>
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<tr>
<td></td>
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<td></td>
<td>3 50 ± 6</td>
<td>93 ± 17</td>
</tr>
<tr>
<td>MPB</td>
<td>44°22'34.24&quot;</td>
<td>79°40'06.45&quot;</td>
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<td>93 ± 15</td>
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## Appendix 6: $^{222}$Rn survey results in Lake Simcoe

Table A6-1 (Cont.): Summary of input parameter values and flux components in the $^{222}$Rn mass balance calculations for regional-scale survey in Kempenfelt Bay. The uncertainty for the $^{222}$Rn concentrations represent the 2σ uncertainty. “/” means no data is available.

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<th>$T$ (°C)</th>
<th>Cond. (µs cm$^{-1}$)</th>
<th>pH</th>
<th>$C_{gw}$ (dpm L$^{-1}$)</th>
<th>$u_{10}$ (km h$^{-1}$)</th>
<th>$z$ (m)</th>
<th>$I_{atm}$ (dpm m$^{-2}$ d$^{-1}$)</th>
<th>$I_{decay}$ (dpm m$^{-2}$ d$^{-1}$)</th>
<th>$I_{diff}$ (dpm m$^{-2}$ d$^{-1}$)</th>
<th>$Q_{gd}$ (m$^{-3}$ m$^{-1}$ d$^{-1}$)</th>
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**Sampling date: 6/09/2015**

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Table A6-2 (Cont.): Summary of input parameter values and flux components in the $^{222}$Rn mass balance calculations for regional-scale survey in Innisfil Area. The uncertainty for the $^{222}$Rn concentrations represent the 2σ uncertainty. “/” means no data is available.

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| 11 | 0.66 ± 0.07 | 23.04 | 409 | 8.4 | 462 ± 39 | 17.58 | 1 | 2536 | 119 | 4.58 | 0.90 ± 0.08 |
| 12 | 0.68 ± 0.14 | 22.47 | 408 | 8.4 | 462 ± 39 | 17.58 | 1 | 2604 | 122 | 4.54 | 0.56 ± 0.05 |
| 13 | 0.80 ± 0.08 | 22.49 | 420 | 8.4 | 462 ± 39 | 17.58 | 1 | 3083 | 145 | 4.27 | 0.90 ± 0.07 |
| 14 | 0.65 ± 0.08 | 22.6 | 419 | 8.46 | 462 ± 39 | 18.83 | 1 | 2790 | 118 | 4.60 | 0.51 ± 0.04 |
| 15 | 0.65 ± 0.05 | / | / | / | 462 ± 39 | 18.83 | 1 | 2769 | 117 | 4.61 | 0.40 ± 0.03 |
Table A6-3 (Cont.): Summary of input parameter values and flux components in the \(^{222}\)Rn mass balance calculations for regional-scale survey in Georgina Area. The uncertainty for the \(^{222}\)Rn concentrations represent the 2σ uncertainty. “/” means no data is available.

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Sampling date: 8/13/2015

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<th>(u_{10}) (km h(^{-1}))</th>
<th>(z) (m)</th>
<th>(I_{atm}) (dpm m(^2) d(^{-1}))</th>
<th>(I_{decay}) (dpm m(^2) d(^{-1}))</th>
<th>(I_{diff}) (dpm m(^2) d(^{-1}))</th>
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Table A6-4 (Cont.): Summary of input parameter values and flux components in the $^{222}$Rn mass balance calculations for high resolution survey near JB (Area A) and JPB (Area C). The uncertainty for the $^{222}$Rn concentrations represent the 2σ uncertainty. “/” means no data is available.

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Sampling location: JB (Area A); Sampling date: 7/10/2015

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References


Curriculum Vitae

Name: Tao Ji

Post-secondary Education and Degrees:
- China University of Geosciences, Wuhan, Hubei Province, China
  2005-2009 BESc.
- East China Normal University, Shanghai, China
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- The University of Western Ontario, London, Ontario, Canada
  2013-2016 MSc.

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- Research Assistant
  The University of Western Ontario
  2013-2016
- Teaching Assistant
  The University of Western Ontario
  2013-2015

Publications: