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Use of Regional Scale Field and Modelling Methods to Evaluate Nearshore Groundwater Discharge to a Large Glacial Lake

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Abstract

Groundwater discharge may be an important pathway for delivering pollutants to large lakes, but this pathway is poorly understood in part because it is characterized by high spatial and temporal variability. Understanding the potential for groundwater discharge to deliver pollutants to lakes requires an evaluation of the magnitude and spatial variability of groundwater discharge to the lake, and the history of the discharging groundwater (e.g., groundwater recharge point, flow paths, and travel times). The first objective of this thesis was to evaluate and quantify the spatial variability of groundwater discharge to a large glacial lake, Lake Simcoe, Ontario, using the naturally occurring radon isotope tracer ($^{222}$Rn). Regional scale boat surveys were conducted along 80% of the Lake Simcoe shoreline using portable radon detection equipment. Groundwater discharge hotspot areas were identified based on spatial variability in lake water $^{222}$Rn concentrations, and regional hydrogeological features were linked to these hotspot areas to develop broadly applicable understanding of the observed spatial distribution of groundwater discharge. Key features included permeable nearshore surficial sediments, proximity to regional recharge features, and presence of tunnel channel deposits. The second objective of this thesis was to compare $^{222}$Rn-derived to model simulated estimates of groundwater discharge in two areas along the Lake Simcoe shoreline. This comparison built further confidence in the groundwater discharge estimates, and enabled the strengths and limitations of each method to be assessed. Particle tracking analysis was used to evaluate the history of groundwater discharging along the northwestern shoreline of Lake Simcoe, and the potential implications for lake water quality in this area. Results showed that groundwater discharging to the lake in the northern area is characterized by long flow paths and travel times, while groundwater discharge in the south originates in the nearshore areas with shorter travel times. The findings of this thesis provide broadly applicable knowledge needed to focus efforts aimed at managing non-point pollution sources to large glacial lakes including groundwater discharge.

Keywords

Groundwater discharge, $^{222}$Rn, Regional scale field methods, Numerical groundwater model, Hydrogeology, Large lakes, Lake Simcoe, Groundwater-surface water interactions
Co-Authorship Statement

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

Chapter 3: Hydrogeologic Controls on Groundwater Discharge to a Large Glacial Lake

Authors: Hayley Wallace, Clare Robinson

Contributions:

Hayley Wallace: Developed field deployment strategy, conducted field surveys, analyzed and interpreted field data, performed lab experiments, wrote draft of chapter

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

Chapter 4: Use of the Tracer $^{222}$Rn and Regional Scale Groundwater Models to Investigate Groundwater Inputs to a Large Glacial Lake

Authors: Hayley Wallace, Spencer Malott, E.J. Wexler, Clare Robinson

Contributions:

Hayley Wallace: Developed field deployment strategy, conducted field surveys, analyzed and interpreted field data, interpreted numerical simulation data, performed particle tracking simulation, wrote draft of chapter

Spencer Malott: Provided model support, aided in particle tracking simulation and analysis, aided in interpretation of model comparison results

E.J. Wexler: Provided model support, aided in interpretation of model comparison results

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

1 Introduction

1.1 Research Background

Over the past several decades, increases in urban and agricultural development and industrial activity has caused widespread deterioration of the water quality in lakes (International Institute for Sustainable Development, 2017; International Joint Commission, 2013; Palmer et al., 2011). For example, increased nutrient (i.e. nitrogen and phosphorous) and chloride loading has contributed to lake eutrophication and salinization respectively (Howard & Livingstone, 2000; Lewandowski et al., 2015). Although often a neglected component of the lake water budget, groundwater discharge can be an important transport pathway for the delivery of pollutants to lakes (Kazmierczak et al., 2016; Kidmose et al., 2015; Meinikmann et al., 2015; Tecklenburg & Blume, 2017). Groundwater can have elevated pollutant concentrations relative to receiving surface water, and as a result pollutant loading associated with groundwater discharge has been implicated in the deterioration of lake water quality and ecosystem health (Haack et al., 2005; Robinson, 2015; Roy & Malenica, 2013). Additionally, because lake water quality management initiatives have traditionally focused on reducing pollutant inputs from point sources and tributaries the relative importance of diffuse non-point source inputs, such as direct groundwater discharge, is increasing (Burnett et al., 2006; Lewandowski et al., 2015; Stets et al., 2010).

Land use activities, and associated pollutants, can directly affect groundwater quality in vulnerable aquifer systems (Eimers et al., 2005; Hansen et al., 2002; Kidmose et al., 2015). Although the link between groundwater quality and land use activities has been well established, the relationship between groundwater quality and subsequent lake water quality is more challenging to evaluate (Kornelsen & Coulibaly, 2014). Understanding the potential for groundwater discharge to deliver pollutants to lakes requires an evaluation of the magnitude and spatial variability of groundwater discharge to the lake, and the history
of the discharging groundwater; including land use in the recharge area, groundwater flow paths, and travel times (Hill, 1990; Smith & Swarzenski, 2012).

Groundwater discharge to lakes is often poorly characterized due to high spatial and temporal variability combined with limited tools available to adequately characterize this variability at a regional scale (Dimova et al., 2013; Mulligan & Charette, 2006; Russoniello et al., 2013). Groundwater discharge can enter the lake in-directly, through groundwater fed streams that flow into the lake, or directly through aquifer layers that intersect the lake bed (Kalbus et al., 2006). This thesis focuses on direct groundwater discharge to lakes. Spatial variability of direct groundwater discharge can be driven by heterogeneities in aquifer sediments, hydraulic gradient between groundwater and lake water levels, and the distribution and volume of recharge to aquifers discharging to the lake (Cherkauer & Hensel, 1986; Feinstein & Reeves, 2010; McBride & Pfannkuch, 1975; Schneider et al., 2005). Field methods that have been applied to quantify groundwater discharge include seepage meters, hydraulic gradient/piezometer measurements, thermal imaging, electrical resistivity tomography (ERT), and geochemical/isotopic tracer methods (Burnett et al., 2006; Dimova et al., 2015; Meinikmann et al., 2013; Santos et al., 2008; Ji et al., 2017). Mass balance calculations of the geochemical tracer Radon-222 ($^{222}$Rn), has been shown to be a suitable field method for the regional scale quantification of groundwater discharge (Burnett et al., 2001; Cable et al., 1996; Dulaiova et al., 2010). Numerical groundwater modelling has also been applied to characterize spatial and temporal groundwater flow patterns at a regional scale (Marchildon et al., 2016). Selecting an appropriate method or combination of methods to evaluate groundwater discharge depends on the characteristics of the study site and study objectives.

Although field methods may provide estimates of the magnitude of groundwater discharge, little attention is given to factors that control spatial variability of groundwater discharge (e.g. Burnett et al., 2002; Corbett et al., 1997; Dimova & Burnett, 2011; Santos et al., 2008). Similarly, field studies that evaluate groundwater discharge generally provides little insight into the history (i.e. flow paths and travel times) of discharging groundwater, and its potential implications for lake water quality. There is a need to quantify and characterize the spatial variability of groundwater discharge, to examine the relationship between the
observed spatial variability and the regional geologic environment, and to compare regional spatial groundwater discharge patterns determined using field and modelling tools. Assessment of geologic controls on the spatial variability of groundwater discharge to lakes in glacial environments is needed to develop broadly applicable and transferrable knowledge, which can be used to better target areas of high direct groundwater discharge for future monitoring efforts or management initiatives.

1.2 Research Objective

This thesis aims to develop understanding of the spatial distribution of groundwater discharge and its potential influence on the water quality of a large glacial lake, Lake Simcoe, Ontario. This thesis is divided into four objectives. The first objective is to quantify direct groundwater discharge, and identify groundwater discharge ‘hotspots’ (areas of elevated groundwater discharge relative to adjacent shoreline) in a large glacial lake using $^{222}$Rn as a tracer. The second objective is to evaluate hydrogeologic controls on the observed spatial variability of groundwater discharge. The third objective is to compare $^{222}$Rn-derived estimates of groundwater discharge to Lake Simcoe to groundwater discharge simulated using regional scale numerical groundwater models. Finally, the fourth objective is to evaluate the flow paths and travel times of groundwater discharging to the lake, and potential implications for the lake water quality. The findings of this research are broadly applicable to other large glacial lake settings.

1.3 Thesis Outline

This thesis is written in “Integrated Article Format.” A brief description of each chapter is presented below

Chapter 1: Introduces the research background and states the research objectives.

Chapter 2: Reviews relevant work related to regional scale quantification of groundwater discharge to lakes, with a focus on the use of $^{222}$Rn as a tracer to quantify groundwater discharge.
Chapter 3: Details field survey methods and data analysis to quantify direct nearshore groundwater discharge to Lake Simcoe. $^{222}\text{Rn}$ is used as a tracer to characterize spatial variability of groundwater discharge and the relationship between the regional hydrogeology and observed spatial variability is assessed.

Chapter 4: Details the comparison between $^{222}\text{Rn}$-derived and model simulated groundwater discharge for two shoreline areas of Lake Simcoe. Results demonstrate the value of comparing independent regional scale estimates of groundwater discharge and provide insight into the history of discharging groundwater and potential implications for lake water quality.

Chapter 5: Summarizes research findings and provides recommendations for future work.
1.4 References


Chapter 2

2 Literature Review

2.1 Groundwater Discharge to Lakes

Groundwater discharge can be an important pathway for the delivery of pollutants to lakes (Cherkauer et al., 1992; Dimova et al., 2013; Grannemann et al., 2000; International Joint Commission, 2013; Kornelsen & Coulibaly, 2014; Rosenberry et al., 2015). Groundwater can discharge to the lake indirectly, through groundwater fed tributaries that flow into the lake, or directly, through aquifer layers that are hydraulically connected to the lake (Kalbus et al., 2006). This review focuses on direct groundwater discharge to lakes. Although groundwater discharge is often a small and neglected component of the lake water budget, particularly for large lakes, pollutant concentrations can be elevated in groundwater relative to receiving lake water. As a result, groundwater discharge can be associated with high pollutant fluxes and cause deterioration of lake water quality and ecosystem health (Haack et al., 2005; Lewandowski et al., 2015; Smith & Swarzenski, 2012). For instance, high nutrient (i.e. nitrogen and phosphorous) loading have been shown to affect nutrient cycling and algal biomass in lakes (Naranjo et al., 2019). Sebestyen & Schneider (2004) found that elevated trace metal concentrations measured in nearshore groundwater were directly related to trace metal concentrations measured in aquatic plant tissue at several lake sites. Generally shallow, unconfined, aquifer layers are more susceptible to contamination. For example, in the Laurentian Great Lakes Basin, shallow aquifer layers have elevated nutrient and chloride concentrations in many areas (e.g. Cherkauer et al., 1992; Hill, 1990; Stotler et al., 2011). Although deep, regionally extensive, aquifer layers may provide a more productive source for municipal water supplies, shallow surficial aquifers can provide a higher portion of direct groundwater discharge to lakes in the nearshore area (Grannemann et al., 2000).

Evaluating groundwater discharge as a pathway for pollutant loading to lakes requires an understanding of (i) land use in the recharge area, where pollutants may enter the groundwater system, (ii) groundwater flow paths linking the recharge area to surface water, and (iii) geochemical transformations that may take place along these flow paths (Hill,
The relationship between land use and associated pollutants, and their subsequent impact on groundwater quality has been studied extensively (Boutt et al., 2001; Eimers et al., 2005; Hansen et al., 2002; Howard & Livingstone, 2000; Kidmose et al., 2015; Kornelsen & Coulibaly, 2014). High nutrient concentrations are often associated with agricultural activities as well as non-agricultural land use practices such as septic systems, leaky urban infrastructure, and landfills (Almasri, 2007; Nolan et al., 1997; Robertson et al., 1991). For example, Kidmose et al. (2015) found that nitrate concentrations in groundwater samples taken adjacent to a crop field were, on average, 70 times higher than samples taken in a forested area located approximately 500 m north. In urban areas elevated chloride concentrations are often associated with application of de-icing agents on roads (Boutt et al., 2001; Howard & Livingstone, 2000).

Evaluating groundwater history (i.e. subsurface flow paths and travel times) is more challenging. There is a general lack of understanding regarding the impact of groundwater discharge on lake water quality, despite studies demonstrating a relationship between groundwater pollutants and declines in ecosystem health (Haack et al., 2005; Kazmierczak et al., 2016; Lewandowski et al., 2015). This lack of understanding is due to the difficulty in quantifying groundwater discharge to surface waters; a diffuse pathway that is characterized by high spatial and temporal variability, and limited techniques available to adequately characterize this variability at a regional scale (Burnett et al., 2006; Dimova et al., 2015; Kornelsen & Coulibaly, 2014; Russoniello et al., 2013). Spatial variability, not only of pollutant concentrations in groundwater, but of the groundwater discharge itself can considerably impact the overall pollutant loading. In investigating the effects of spatial variability in both groundwater phosphorous (P) concentration and groundwater discharge to a small lake, Meinikmann et al., (2013) showed that estimated P loading to the lake increased from 327 kg yr\(^{-1}\) to 425 kg yr\(^{-1}\) when lake-bed temperature measurements were used as weighting factors to take spatial groundwater discharge patterns into account. Groundwater travel time is also an important consideration to determine the fate and impact of groundwater pollutants. In the Toronto area, Howard and Livingstone (2000) used modelling to show that conservative contaminants present in aquifers within a few kilometers of Lake Ontario will discharge to the lake over the next 50 years; suggesting a cumulative and legacy threat to lake water quality. The slow transport of nutrients,
particularly P, in shallow aquifers can also pose a legacy issue for receiving surface water and can buffer the impact of nutrient management strategies focused on tributaries and point sources of pollution (Jarvie et al., 2013; Martin et al., 2011; Sharpley et al., 2013).

There is a need for better understanding of the spatial variability, quantity, and history of groundwater discharging to lakes to develop effective lake water quality management actions. This thesis focuses on regional scale characterization of direct groundwater discharge to a large glacial lake, Lake Simcoe, Ontario.

2.2 Regional Scale Methods to Quantify Groundwater Discharge

A variety of local and regional scale methods are available to quantify groundwater discharge. These methods include seepage meters, hydraulic gradient/piezometer measurements, numerical groundwater models, water budget assessments, electrical resistivity tomography (ERT) imaging, thermal imaging, and geochemical/isotopic tracer methods (Dimova et al., 2015; Ji et al., 2017; Kidmose et al., 2015; Meinikmann et al., 2013; Mulligan & Charette, 2006; Santos et al., 2008). Selecting an appropriate method, or combination of methods, depends on the site characteristics and the objective of the study (Burnett et al., 2006). For example, seepage meter and hydraulic gradient/piezometer methods provide information from spot measurements and it is therefore more appropriate to apply these methods on a local scale (i.e. 1-100 m) (Dimova et al., 2013; Lambert & Burnett, 2003). Alternatively, geochemical tracer and numerical groundwater modelling methods are more suitable for regional scale (i.e. 1-100 km) measurements, and characterization of regional scale spatial variability (Bugna et al., 1996; Burnett & Dulaiova, 2003; Kornelsen & Coulibaly, 2014; Santos et al., 2008). In this thesis, the magnitude and spatial variability of direct groundwater discharge to a large glacial lake is evaluated at a regional-scale using the geochemical tracer Radon-222 ($^{222}$Rn). $^{222}$Rn-derived groundwater discharge estimates are (i) used to determine how regional hydrogeologic features may influence the observed spatial variability in groundwater discharge, and (ii) compared with independent groundwater discharge estimates derived from numerical groundwater models. The following sections provide a review of the use
of $^{222}$Rn and numerical groundwater modelling to estimate groundwater discharge to surface waters, with a focus on discharge to large lakes.

### 2.2.1 $^{222}$Rn as a Tracer for Groundwater Discharge

$^{222}$Rn has been widely used as a geochemical tracer for groundwater discharge in marine, riverine, and lake environments (Burnett et al., 2006; Burnett & Dulaiova, 2003; Cable et al., 1996; Dimova et al., 2013; Ji et al., 2017; Mullinger et al., 2007; Ono et al., 2013; Peterson et al., 2007). In general, a suitable natural tracer for groundwater discharge should be (i) conservative, (ii) have elevated concentrations in groundwater, (iii) if it is radioactive, decay at a time scale that is comparable to relevant coastal processes, and (iv) be relatively straightforward to measure (Burnett et al., 2001; Cable et al., 1996; Tuccimei et al., 2005). $^{222}$Rn is a naturally occurring isotope of radium ($^{226}$Ra); a daughter product of the $^{238}$U decay chain. It is a chemically and biologically inert gas, with concentrations that are typically 3-4 order of magnitude higher in groundwater than receiving surface water (Corbett et al., 1997). $^{222}$Rn gas is emitted by the decay of $^{226}$Ra from the rock or soil matrix, and dissolves into groundwater as it travels through the aquifer (Klug et al., 2007). Elevated $^{222}$Rn concentrations in groundwater have been measured in almost all aquifer material, including glacial sediments (Je & Eyles, 1998; Mulligan & Charette, 2006; Schmidt & Schubert, 2007). The half-life of $^{222}$Rn is 3.82 d, which is a suitable length of time for studying nearshore processes (Burnett et al., 2001). Delivery of $^{222}$Rn to surface waters occurs predominately by discharging groundwater, and to a lesser extent, through diffusion from lake bed sediments (Burnett & Dulaiova, 2003). The use of $^{222}$Rn, as well as other geochemical tracers, is ideal for characterizing large scale variability in groundwater discharge, because the tracer signal is integrated in the water column, which effectively smooths out smaller scale heterogeneities (Burnett et al., 2006; Swarzenski, 2007).

Measurements of $^{222}$Rn concentrations in groundwater and surface water can be taken using the portable RAD7 unit (Durridge Co.). This is a commercially available alpha spectrometer that measures $^{222}$Rn based on the activity of its daughter products- primarily $^{218}$Po (Kluge et al., 2007). $^{222}$Rn concentrations in discrete water samples can be measured using the RAD H$_2$O system (Durridge Co.) with a single RAD7 detector (e.g. Dimova et
For regional scale assessment of \(^{222}\text{Rn}\) concentrations in surface waters, a continuous monitoring system was developed by Burnett et al. (2001). Measurements are taken by continuously pumping water to an air-water exchanger (RAD AQUA; Durridge Co.), where \(^{222}\text{Rn}\) concentrations between air and water reach equilibrium within a closed air loop system, and are measured by the RAD7 detector (Burnett & Dulaiova, 2003; Lane-Smith et al., 2002; Durridge Co.). Many studies have used this continuous sampling method for investigation of both spatial and temporal variability in \(^{222}\text{Rn}\) concentrations (e.g. Burnett et al., 2008; Dulaiova et al., 2010; Peterson et al., 2007; Smith & Swarzenski, 2012). Dulaiova et al. (2005) also showed that multiple RAD7 units can be connected in parallel to increase sensitivity and resolution of measurements.

\(^{222}\text{Rn}\) measurements can be used to estimate groundwater discharge by applying a steady state mass balance model for a well-mixed surface water volume (shown in Figure 2.1; Burnett et al., 2001; Burnett & Dulaiova, 2003; Cable et al., 1996; Dulaiova et al., 2010; Schmidt et al., 2010; Smith & Swarzenski, 2012). This mass balance model, which considers the various sources and sinks of \(^{222}\text{Rn}\) within the water volume, is given as:

\[
0 = J_{gw} + J_{diff} - J_{atm} - J_{mix} + J_{prod} - J_{decay}
\]  

(2-1)

where \(J_{GW}\) (dpm m\(^{-2}\) d\(^{-1}\)) is the \(^{222}\text{Rn}\) input from groundwater discharge, \(J_{diff}\) (dpm m\(^{-2}\) d\(^{-1}\)) is the diffusion of \(^{222}\text{Rn}\) from lake bed sediment, \(J_{prod}\) (dpm m\(^{-2}\) d\(^{-1}\)) is the production of \(^{222}\text{Rn}\) from \(^{226}\text{Ra}\) decay, \(J_{atm}\) (dpm m\(^{-2}\) d\(^{-1}\)) is the evasion of \(^{222}\text{Rn}\) to the atmosphere, \(J_{decay}\) (dpm m\(^{-2}\) d\(^{-1}\)) is the decay of \(^{222}\text{Rn}\), and \(J_{mix}\) (dpm m\(^{-2}\) d\(^{-1}\)) is the offshore mixing with low \(^{222}\text{Rn}\) waters. While \(J_{prod}\) can be a significant source term for coastal ocean sites, studies have shown that \(^{226}\text{Ra}\) concentrations are very low in freshwater, and this production can be neglected in lake settings (Dulaiova & Burnett, 2008; Moore, 1996). In environments where the advective flux of groundwater is the primary source of \(^{222}\text{Rn}\) in the water column, \(J_{diff}\) can also be neglected (Dimova et al., 2013; Dulaiova et al., 2010; Ji et al., 2017).
Figure 2.1: Conceptual diagram of $^{222}\text{Rn}$ mass balance model for estimating groundwater discharge (modified from Burnett & Dulaiova, 2003). Sources of $^{222}\text{Rn}$ in the lake water volume include groundwater discharge ($J_{gw}$), diffusion of $^{222}\text{Rn}$ from lake bed sediments ($J_{diff}$), and production of $^{222}\text{Rn}$ from $^{226}\text{Ra}$ decay ($J_{prod}$). Losses of $^{222}\text{Rn}$ in the lake water volume include atmospheric evasion ($J_{atm}$), in-situ decay of $^{222}\text{Rn}$ ($J_{decay}$), and mixing with offshore waters ($J_{mix}$).

The earliest application of the $^{222}\text{Rn}$ mass balance concept was by Ellins et al. (1990) who considered $^{222}\text{Rn}$ losses due to atmospheric evasion and radioactive decay to estimate groundwater inputs to a stream. Cable et al. (1996) later applied a linked benthic exchange-horizontal transport model, together with surface water $^{222}\text{Rn}$ measurements, to identify areas of high groundwater discharge to a coastal area along the Gulf of Mexico. This benthic exchange-horizontal transport model was also applied by Corbett et al. (1997) to a small lake site in south Carolina with results showing that groundwater inputs to the lake accounted for a significant portion (10-33%) of the inputs in the lake water budget. In applying the steady state mass balance model (Equation 2-1), Burnett & Dulaiova (2003) presented a method for estimating specific groundwater flux ($q_{gd}$; m d$^{-1}$): 

$$q_{gd} = \frac{J_{gw}}{C_{gw}}$$  \hspace{1cm} (2-2)

where $C_{gw}$ (dpm m$^{-3}$) is the $^{222}\text{Rn}$ concentration of the groundwater endmember (Schmidt et al., 2010; Smith & Swarzenski, 2012). Studies have applied this method to evaluate total
groundwater flux into surface waters, and to investigate temporal variability at a single location using continuous $^{222}\text{Rn}$ time-series measurements (e.g. Burnett & Dulaiova, 2006; Dimova et al., 2013; Lambert & Burnett, 2003; Mulligan & Charette, 2006; Schmidt et al., 2010). For example, Burnett et al. (2008) used continuous $^{222}\text{Rn}$ measurements at a location along a coastal embayment in Brazil to show that specific groundwater flux can be driven by tidal fluctuations. The $^{222}\text{Rn}$ mass balance approach has also been applied to characterize spatial variability of groundwater discharge using continuous offshore $^{222}\text{Rn}$ surveys (Dimova et al., 2013; Dulaiova et al., 2010; Ji et al., 2017; Ono et al., 2013). An equation that has been used to estimate groundwater discharge ($Q$; m$^3$ d$^{-1}$) from continuous spatial $^{222}\text{Rn}$ measurements is (Dulaiova et al., 2010):

$$Q = \frac{C_{sw}V}{C_{gw} \tau} \quad (2-3)$$

where $C_{sw}$(dpm m$^{-3}$) is the measured surface water $^{222}\text{Rn}$ concentration corrected for sources and losses, $V$ (m$^3$) is the surface water volume over which the mass balance is applied, and $\tau$ (d) is the flushing rate. Dulaiova et al. (2010) conducted along-shore $^{222}\text{Rn}$ boat surveys at two coastal sites using 3-RAD7 units in parallel, while simultaneously measuring nitrate concentrations to identify locations of elevated groundwater discharge and their potential to be sources of non-point source pollution. More recently, Dimova et al. (2013) applied this method to evaluate total nearshore groundwater discharge to several small lakes in Florida. Studies have shown that, in general, $^{222}\text{Rn}$-derived estimates of groundwater discharge agree well with independent estimates from seepage meter, hydraulic gradient, and hydrologic model estimates (Burnett et al., 2006; Burnett & Dulaiova, 2003; Dimova & Burnett, 2011; Mulligan & Charette, 2006; Tuccimei et al., 2005). Reported uncertainties in groundwater discharge estimates are propagated errors from the source and sink terms in the $^{222}\text{Rn}$ mass balance model (Dimova et al., 2013).

Although $^{222}\text{Rn}$ has been widely applied as a tracer to quantify groundwater discharge, the mass balance method has uncertainties and limitations with its application; namely the $^{222}\text{Rn}$ losses due to atmospheric evaporation ($J_{\text{atm}}$), offshore mixing ($J_{\text{mix}}$), and quantification of the groundwater endmember $^{222}\text{Rn}$ concentration ($C_{gw}$) (Burnett et al., 2007). The method
provided in Macintyre et al. (1995), and adapted by Burnett & Dulaiova (2003) and Dulaiova et al. (2010) is often used to evaluate $J_{\text{atm}}$ (dpm m$^{-2}$d$^{-1}$):

$$J_{\text{atm}} = k(C_{\text{WC}} - \alpha C_{\text{air}})$$  \hspace{1cm} (2-4)

where $C_{\text{WC}}$ (dpm m$^{-3}$) and $C_{\text{air}}$ (dpm m$^{-3}$) are the $^{222}\text{Rn}$ concentrations measured in surface water and ambient air respectively. The gas transfer coefficient, $k$ (m d$^{-1}$), and the Ostwald’s solubility coefficient, $\alpha$ (dimensionless), are calculated by (Dimova et al., 2013; Macintyre et al., 1995):

$$k(600) = 0.45 \times 1.6 \times \left( \frac{S_{c}}{600} \right)^{-b}$$  \hspace{1cm} (2-5)

$$\alpha = 0.105 + 0.405 \exp(-0.05027T)$$  \hspace{1cm} (2-6)

where $u_{10}$ (m s$^{-1}$) is the wind speed at 10 m above ground, and $\alpha$ is dependent on the temperature at the air-water interface (°C). There is some evidence to suggest that these equations, which are based on principles of diffusion across the air-water interface and wind-dependent gas transfer, may not provide the correct rate of atmospheric evasion under extreme conditions (i.e. high winds causing waves, tidal surges, precipitation). For example, Burnett et al. (2007) observed significant drops in $^{222}\text{Rn}$ inventory during tropical storm ‘Alberto’ in the Gulf of Mexico, where the method for calculating $J_{\text{atm}}$ underestimated the $^{222}\text{Rn}$ losses by 40-80% when wind speeds reached up to ~20 m s$^{-1}$ with a maximum storm surge of ~0.5 m. In a sheltered harbor environment, Burnett & Dulaiova (2006) also reported decreases in calculated groundwater flux that correspond to periods of high winds (~10 m s$^{-1}$). In non-tidal environments, where losses from atmospheric evasion may represent a large portion of the total $^{222}\text{Rn}$ losses (~19-90%), an underestimation in $J_{\text{atm}}$ can significantly modify the calculated groundwater discharge rate (Santos et al., 2008).

Similarly, there is uncertainty in calculating the $^{222}\text{Rn}$ loss due to offshore mixing ($J_{\text{mix}}$; dpm m$^{-2}$d$^{-1}$) in different environments. $J_{\text{mix}}$ can be a significant loss term in tidal coastal environments, with Burnett et al. (2007) concluding that mixing loss can represent 60-97% of total $^{222}\text{Rn}$ losses in these environments. However, in non-tidal (i.e. lake, non-tidal
lagoon) environments, \( J_{\text{mix}} \) does not represent a significant loss and is often neglected (e.g. Corbett et al., 1997; Schmidt & Schubert, 2007; Tuccimei et al., 2005). For example, Santos et al. (2008) estimated horizontal offshore mixing in a non-tidal environment only accounted for 18-20% of total losses. \( J_{\text{mix}} \) can be estimated using the method developed by Moore (2000) and adapted by Smith & Swarzenski (2012), by first using offshore transect measurements of conservative short lived radium isotopes \(^{223}\text{Ra} \) and \(^{224}\text{Ra} \) to calculate the horizontal eddy diffusion coefficient \( K_h \) (m\(^2\) d\(^{-1}\)):

\[
\ln(C_x) = \ln(C_0) - \frac{x}{\sqrt{\lambda/K_h}} \quad \text{(2-7)}
\]

where \( C_x \) and \( C_0 \) (dpm m\(^{-3}\)) are the \(^{223}\text{Ra} \) and \(^{224}\text{Ra} \) concentrations, \( x \) (m) is the distance offshore, and \( \lambda \) is the decay constant (d\(^{-1}\)). The slope of the line, once concentrations are plotted against offshore distance, is equal to \( \sqrt{\lambda/K_h} \), and \( K_h \) can be solved for each transect.

The \( J_{\text{mix}} \) is then calculated by:

\[
J_{\text{mix}} = -K_h \left( \frac{C_{S+1} - C_{S-1}}{2\Delta x} \right) \times \frac{z_{\text{mix}}}{x} \quad \text{(2-8)}
\]

where, \( \frac{C_{S+1} - C_{S-1}}{2\Delta x} \) (dpm m\(^{-3}\) m\(^{-1}\)) is the offshore \(^{222}\text{Rn} \) gradient, and \( z_{\text{mix}} \) is the thickness of the mixed layer. However, if \(^{223}\text{Ra} \) and \(^{224}\text{Ra} \) concentrations are not measured, equation 2-7 cannot be directly applied to solve for \( K_h \) using offshore \(^{222}\text{Rn} \) measurements because \(^{222}\text{Rn} \) is not conservative in surface water (i.e. losses due to atmospheric evasion and decay). The offshore transect of \(^{222}\text{Rn} \) concentrations must first be corrected for these losses. An iterative method to account for the losses was presented by Santos et al. (2008).

Offshore gradients of measured \(^{222}\text{Rn} \) (non-conservative tracer) and conductivity (conservative tracer) can be used in a steady state advection diffusion equation to estimate the horizontal eddy diffusion coefficient \( K_h \) (m\(^2\) d\(^{-1}\)):

\[
K_h \frac{\partial^2 C}{\partial x^2} - \omega \frac{\partial C}{\partial x} - \lambda C = 0 \quad \text{(2-9)}
\]

where \( \omega \) (m d\(^{-1}\)) is the horizontal surface advection, and \( C \) is the concentration of the tracer. To calculate \( K_h \), the measured offshore \(^{222}\text{Rn} \) concentrations are corrected for \( J_{\text{decay}} \) and \( J_{\text{atm}} \) over the travel time to the measurement location, and an iterative approach is used;
whereby the corrected $^{222}$Rn are substituted back into equation (2-9) to estimate a new $K_h$ and $\omega$, until the values converge. The final $K_h$ value is used to calculate $J_{\text{mix}}$ (Burnett et al., 2008; Moore, 2000; Santos et al., 2008). The uncertainty associated with the $J_{\text{mix}}$ term comes from its sensitivity to parameters assigned in the equations; for example the $z_{\text{mix}}$ term in equation 2-8 (Burnett et al., 2007). The impact of this uncertainty on the final groundwater discharge values, however, depends on the relative magnitude of the mixing loss within the mass balance. For example, in a coastal environment Lambert & Burnett (2003) found that doubling $J_{\text{mix}}$ increased the groundwater discharge by 25%, because $J_{\text{mix}}$ represented a large component in the mass balance in this environment. Conversely, Santos et al. (2008) found that in a non-tidal environment a sensitivity analysis of the $K_h$ value derived from the iterative method only changed the estimated groundwater discharge by 4%. It should also be noted that the mass balance method for calculating loss of $^{222}$Rn due to mixing only considers mixing in the offshore direction, and neglects the effect of alongshore currents.

The most significant source of uncertainty in applying the $^{222}$Rn mass balance method to estimate groundwater discharge is in quantifying the $^{222}$Rn concentration of the groundwater endmember, $C_{gw}$ (dpm m$^{-3}$) (Burnett et al., 2007). Measured groundwater $^{222}$Rn concentrations exhibit high spatial variability due to natural geologic heterogeneities, and as such determining a representative concentration can be challenging (Dimova & Burnett, 2011). Dimova et al. (2013) showed that, in some cases, uncertainties in groundwater discharge estimates can be higher than 50% due to uncertainties associated with assigning an endmember concentration. It is not uncommon for studies to report a range of measured groundwater $^{222}$Rn concentrations spanning 1-2 orders of magnitude (Burnett et al., 2006; Cable et al., 1996; Ellins et al., 1990; Moore, 1996; Schmidt & Schubert, 2007). For example, Mulligan & Charette (2006) measured groundwater $^{222}$Rn concentrations ranging from 190 dpm L$^{-1}$ to 5900 dpm L$^{-1}$ at small (~400m) coastal aquifer study site. To reduce variability, and ensure a more representative estimate, studies have used sediment equilibration experiments to estimate groundwater endmember $^{222}$Rn concentration (Burnett & Dulaiova, 2003; Burnett et al., 2008; Santos et al., 2008; Tuccimei et al., 2005). For this approach, a known volume of nearshore lake-bed or coastal
sediment is collected, and allowed to equilibrate with overlying surface water in an airtight container for at least 20 d. This is sufficient time to ensure that equilibrium is reached between $^{222}\text{Rn}$ and sediment-bound $^{226}\text{Ra}$. By combining the measured equilibrium $^{222}\text{Rn}$ activity in the surface water, and physical properties of the sediment (i.e. bulk density, porosity), the pore water $^{222}\text{Rn}$ concentration can be estimated (Chanyotha et al., 2014; Corbett et al., 1998). Following a recommendation by Burnett et al. (2007), some studies have used a combination of groundwater sampling and sediment equilibration methods to estimate a representative groundwater endmember $^{222}\text{Rn}$ concentration (e.g. Dimova et al., 2013; Schmidt et al., 2010; Schmidt & Schubert, 2007).

In applying $^{222}\text{Rn}$ as a tracer to quantify groundwater discharge to a large glacial lake, key challenges include consideration of the effects of weather conditions (i.e. wind, waves, and precipitation), offshore mixing loss, and spatial variability of groundwater $^{222}\text{Rn}$ concentrations. This thesis addresses these limitations in applying $^{222}\text{Rn}$ data to estimate groundwater discharge rates.

### 2.2.2 Regional Scale Groundwater Modelling

While many studies have shown that $^{222}\text{Rn}$ is a suitable tracer for evaluation of groundwater discharge at a regional scale, these studies do not consider the history of the discharging groundwater (i.e. groundwater flow paths, aquifer residence times, and recharge areas) in discussing the potential impacts on lake water quality. Regional scale numerical groundwater models can be used to evaluate groundwater history including characterizing spatial and temporal groundwater flow patterns, and informing management decisions. Regional scale groundwater models are often developed and applied for a variety of uses, including delineation of drinking water protection zones, evaluating the impacts of climate change, and evaluating the effects of land use management changes and well pumping (e.g. Buxton et al., 1991; Hassan et al., 2014; Niswonger et al., 2014; Rock & Kupfersberger, 2002; Woolfenden & Nishikawa, 2014).

The US Geological Survey (USGS) groundwater modelling program MODFLOW has been used extensively to investigate and characterize groundwater flow systems (Batelaan et al., 2003; Boutt et al., 2001; Meriano & Eyles, 2003; Modica & Buxton, 1998).
MODFLOW simulates unconfined and confined groundwater flow by numerically solving the three-dimensional groundwater flow equation, for each cell in the model domain (Matula et al., 2014). Once the hydrostratigraphy of the regional aquifer-aquitard system, corresponding properties (i.e. storage coefficient, hydraulic conductivity), and boundary conditions are assigned, the model is able to simulate groundwater levels, groundwater flow between shallow and deep aquifer layers, and groundwater discharge across model boundaries including surface water features. For example, Boutt et al. (2001) estimated the total direct groundwater discharge to Grand Traverse Bay, Lake Michigan, to be 7% of the total water budget input, by summing the MODFLOW simulated groundwater fluxes across the lake boundary. More recently, integrated groundwater-surface water modelling has been conducted using programs such as GSFLOW; which combines PRMS (Precipitation Runoff Modelling Software) to simulate recharge, with MODFLOW to simulate groundwater flow (Huntington & Niswonger, 2012; Markstrom et al., 2008; Woolfenden & Nishikawa, 2014). Precipitation, climate conditions, soil type, land surface topography and land use are incorporated into the PRMS model to produce inputs for the MODFLOW groundwater model including groundwater recharge and runoff (Leavesley et al., 1983).

Groundwater models have also been used to simulate contaminant transport including evaluating the potential for delivery of contaminants to surface water via groundwater discharge (e.g. Boutt et al., 2001; Howard & Livingstone, 2000; Meriano & Eyles, 2003). For example, Cherkauer et al. (1992) used groundwater flow and contaminant transport modelling to show that agricultural and urban land uses in the Green Bay area, Lake Michigan, may contribute 58% and 50% of chloride and nitrate loading, respectively, from the study area and only 38% of water volume. More recently, Kidmose et al. (2015) compared field measured nitrate concentrations with a groundwater flow and transport model to estimate nitrate loading to a lake in Denmark. They found that groundwater from a crop field area accounted for 23% of total groundwater discharge to the lake but 96% of the nitrate loading. Knowledge of aquifer residence times (i.e. groundwater travel time from recharge to discharge) can also be combined with contaminant transport modelling to simulate long-term effects on water quality. Modica & Buxton (1998) showed that under the current land use conditions in their study area, 89% of groundwater discharging to a
stream is enriched with nitrogen while 11% is considered ‘clean’ having travel times that predate agricultural land use. Furthermore, to evaluate the effects of land use management changes, they showed that if land application of nitrogen sources stopped it could take up to 10 years for the percentage of nitrogen-enriched groundwater discharge to be reduced to 45%. To evaluate the specific groundwater flow paths that connect areas of recharge and discharge, particle tracking analysis can also be conducted (Batelaan et al., 2003; Marchildon et al., 2016; Matula et al., 2014; Modica et al., 1998; Rock & Kupfersberger, 2002).

Particle tracking can be performed using programs including the USGS program MODPATH (Pollock, 2012). MODPATH uses assigned aquifer properties (i.e. porosity and hydraulic conductivity), and cell-to-cell flow velocities generated from a MODFLOW simulation to calculate flow paths for virtual ‘particles’ placed in areas of interest within the model domain. Particles can be tracked three dimensionally, forward in time to their discharge point, or backward in time to their recharge point (Batelaan et al., 2003; Marchildon et al., 2016; Matula et al., 2014). Howard & Livingstone (2000) used reverse particle tracking from the Lake Ontario shoreline to determine travel times from known sources of groundwater contamination (i.e. landfills, underground storage tanks, and agricultural areas) to the lake.

Although regional scale groundwater flow models can provide insight into flow pathways and residence times, there are several limitations associated with their application. The accuracy of a numerical groundwater model and particle tracking depends on the accuracy and level of detail incorporated into the hydrogeologic characterization (Buxton et al., 1991). The accuracy of the model is also a function of the model scale and resolution, and depending on the scale of the model it may not incorporate smaller scale heterogeneities (Modica & Buxton, 1998). Given these limitations, field validation of model results is recommended (Feinstein & Reeves, 2010; Grannemann et al., 2000; Kornelsen & Coulibaly, 2014). For example, Batelaan et al., (2003) validated simulated groundwater discharge patterns by mapping the growth of phreatophytes. Results showed 79% agreement between groundwater discharge and plant locations, which increased confidence in model results.
In Ontario, a large number of regional-scale groundwater models have been developed for drinking water source protection and water resource management initiatives (shown in Figure 2.2). Under the Clean Water Act, passed by the Ontario government in 2006, watershed scale models were developed for Source Water Protection Planning for regions containing existing and future drinking water sources (Holysh & Gerber, 2014). In addition, regional scale models have been developed for areas of special interest for water quality protection (Kornelsen & Coulibaly, 2014; Sharpe et al., 2004). Under the Lake Simcoe Protection Plan (MOECC, 2009) integrated groundwater-surface water models were developed for subwatershed areas within the Lake Simcoe Basin as part of Tier 2 Water Budget Analysis and Water Quantity Stress Assessment, and Ecologically Significant Groundwater Recharge Area (ESGRA) Assessment studies (AquaResource Inc., 2013; Earthfx Inc., 2012, 2013; Marchildon et al., 2016). Although the objective of these models was not to evaluate the quantity and spatial distribution of groundwater discharge to surface water bodies, it may be possible to apply them for this purpose provided they incorporate sufficiently characterized recharge and hydrogeologic information. Many of the models developed in the Lake Simcoe Basin, under Source Water Protection or Lake Simcoe Protection Plan initiatives, have explicitly characterized and quantified groundwater flow in the model domain (Figure 2.2). This thesis will evaluate the capability of two regional scale sub-models within the Lake Simcoe Basin to simulate groundwater discharge to the lake, by comparing simulated groundwater discharge estimates to $^{222}$Rn-derived groundwater discharge estimates.
Figure 2.2: Map of Ontario Source Water Protection Assessment areas (modified from Kornelsen & Coulibaly, 2014). Areas shown (a) do not consider groundwater discharge, (b) qualitatively consider groundwater discharge, and quantify groundwater discharge (c) within the watershed, and (d) at reach scale.

2.2.3 Intercomparison of Regional Scale Field Methods and Groundwater Modelling

Comparing independent regional scale estimates of groundwater discharge can improve confidence in the discharge results, as well as provide insight into the strengths and limitations of each method within a particular study environment (Burnett et al., 2006). In general, $^{222}\text{Rn}$-derived groundwater discharge estimates have been compared with independent local scale field estimates from seepage meters and hydraulic gradient measurements, as well as electrical resistivity tomography (ERT) imaging, and qualitative tracers including conductivity, methane, and temperature (e.g. Corbett et al., 2000; Dimova & Burnett, 2011; Dulaiova & Burnett, 2006; Ji et al., 2017; Ono et al., 2013). For example, Dimova et al. (2013) found good agreement between continuous along-shore $^{222}\text{Rn}$, methane, and conductivity measurements in several small, shallow lakes. Comparisons have also been made between $^{222}\text{Rn}$-derived and water budget estimates of groundwater
discharge, although water budget analyses provide limited insight into the spatial variability of discharge patterns (Corbett et al., 1997).

Similarly, groundwater discharge estimates from numerical groundwater models have been compared quantitatively to seepage meter and hydraulic gradient measurements, and qualitatively to groundwater chemistry, for field validation (Bouff, 2001; Kazmierczak et al., 2016; Turner & Townley, 2006). For example, Kidmose et al. (2011) compared the model simulated distribution of groundwater discharge to a small lake in Denmark with stable isotope (O$^{18}$), and chlorofluorocarbon (CFC) measurements for groundwater age. Results of the study showed good agreement between average simulated groundwater travel time and CFC-derived groundwater age estimates of 13yr and 16yr, respectively.

Although geochemical tracers, such as CFC and stable isotopes, provide regional scale field data for model validation, these methods do not provide independent quantification of groundwater discharge. Studies combining regional scale groundwater discharge field data with numerical groundwater flow simulation results are limited. To our knowledge, there is only one intercomparison study for $^{222}$Rn-derived and model simulated groundwater discharge estimates. Lambert & Burnett (2003) and Smith & Zawadzki (2003) used continuous offshore $^{222}$Rn measurements and a density-dependent FEFLOW groundwater model to quantify submarine groundwater discharge along the Gulf of Mexico coast in Florida. Comparison between the two methods over time show that the results were not consistent - this was attributed to insufficient salinity data for model calibration and the effect of tidal pumping on measured field $^{222}$Rn inventories. To our knowledge, no studies have been done to compare groundwater discharge estimates determined from $^{222}$Rn-surveys and model simulations in a lake environment, at a large watershed scale. This thesis will compare the magnitude and spatial variability of groundwater discharge determined from $^{222}$Rn field measurements and regional scale numerical modelling for two >40km shoreline areas along the shoreline of a large glacial lake.
2.3 Linking Field Estimates of Groundwater Discharge to Regional Geology

In general, there are several factors that have been shown to control the spatial variability of groundwater discharge. Groundwater discharge from a homogeneous aquifer is controlled by the hydraulic gradient between the groundwater table and surface water level; whereby groundwater discharge will be highest in the nearshore area and decrease exponentially with distance offshore (Freeze, 1967; McBride & Pfannkuch, 1975). In reality, however, groundwater discharge is often spatially variable due to heterogeneities that occur in aquifer geology and hydraulic conductivity, nearshore topography, and distribution of recharge (Burnett et al., 2006; Cherkauer & Hensel, 1986; Lee, 1977). For example, Kishel & Gerla (2002) found that preferential groundwater flow paths to Shingobee Lake, Minnesota developed as a result of heterogeneities in both surficial aquifer and lake bed sediments. Geologic heterogeneities can also result in offshore groundwater discharge from confined layers (Cherkauer & Nader, 1989; Kidmose et al., 2011). Studies have shown that shoreline configuration and anthropogenic modifications to the shoreline can also affect the spatial distribution of direct groundwater discharge. For example, Cherkauer & McKereghan (1991) found that groundwater discharge can be elevated along shoreline embayments as a result of distorted equipotential flow paths caused by irregular shoreline configuration. Santos et al. (2008) found that 70% of measured groundwater discharge along the shoreline of a coastal lagoon was coming from several small dredged irrigation canals; the construction of which was hypothesized to have penetrated through the regional aquitard layer. Heterogeneities lead to variability in groundwater discharge at both local and regional scales. For example, in the Laurentian Great Lakes Basin each of the lakes is a discharge feature because they act as topographic low points in their individual sub-basins. However, on the basin scale Lake Michigan is thought to have the highest direct groundwater discharge because of the abundance of high permeability (i.e. sand and gravel) nearshore aquifer sediment (Grannemann et al., 2000).

It is important to understand the factors that control spatial variability of groundwater discharge, because areas with preferential or higher groundwater inputs, may also have higher groundwater-derived pollutant inputs depending on the quality of the discharging
groundwater (Dulaiova et al., 2010; Kidmose et al., 2015; Ono et al., 2013). Although it is important to link groundwater discharge patterns with regional hydrogeologic features, studies that have applied regional scale field methods to quantify direct groundwater discharge (including the use of $^{222}\text{Rn}$ as a tracer) generally provide little insight into these controls (Corbett et al., 1997; Dimova & Burnett, 2011; Kornelsen & Coulibaly, 2014; Lewandowski et al., 2015; Tecklenburg & Blume, 2017). In a study done by Burnett et al. (2006), investigations were conducted in five different geologic environments (i.e. karst, crystalline bedrock, volcanic, coastal plain, and glacial) to quantify groundwater discharge using different techniques/ combinations of techniques at each site, including $^{222}\text{Rn}$ as a tracer. The objective of this study was to make recommendations about which measurement techniques are best suited for which specific environment. While the focus of this study was not to investigate the relationship between spatial variability of groundwater discharge and the regional geologic environment, they explore how the regional geology may affect the groundwater flow systems and thus groundwater discharge. For example a karstic or fractured bedrock environment might require a field measurement technique that could adequately capture high variability in the magnitude of groundwater discharge within a small area.

Several studies have been conducted in marine coastal environments, linking local and regional scale field measurement tools to regional geologic controls on groundwater discharge patterns. Mulligan & Charette (2006) used $^{222}\text{Rn}$ point measurements, seepage meters, and hydraulic gradient measurements to characterize the site-specific groundwater discharge in a coastal estuary, whose environment was shaped primarily by glacial processes. They showed that spatial variability in discharge measurements was consistent with changes in nearshore topography; where a high discharge corresponded to a steeper gradient. The findings of the study in this coastal environment, however, are focused on the largest influence on measured $^{222}\text{Rn}$ concentrations: temporal variations caused by tidal fluctuation. More recently, in a coastal estuary environment, Russoniello et al. (2013) investigated the influence of small and large-scale geologic heterogeneities on the spatial variability of groundwater discharge and groundwater salinity using geophysical survey, seepage meter, and groundwater salinity measurements. They showed that both local-scale changes in aquifer material, and a large shore-perpendicular paleo valley features affected
the quantity and salinity of groundwater discharge, which may have important implications for surface water chemistry and pollutant loading to the estuary.

In lake environments specifically, several smaller scale studies have been conducted to investigate hydrogeologic controls on groundwater discharge. Schneider et al. (2005) investigated groundwater discharge into a large glacial lake, along a 240 m shoreline segment, using seepage meter measurements. They found that the deeper regional groundwater flow system (i.e. bedrock geology, regional aquifer recharge) had a larger influence on the spatial distribution of groundwater discharge than the local surficial sediment type, despite the high variability in surficial sediments along the shoreline in the study area. More recently, Tecklenburg & Blume (2017) investigated controls on the spatial variability of groundwater discharge to a small glacial lake (0.49 km$^2$) in Germany. They used vertical temperature profiles to quantify nearshore groundwater inputs, in addition to fibre optic temperature sensor and $^{222}\text{Rn}$ sample measurements as qualitative tracers. Results of this study showed that large scale variability in the groundwater discharge patterns was strongly correlated to topography and the groundwater flow field (i.e. hydraulic gradient), and small scale variability was strongly correlated to aquifer sediment type; where large grain size was associated with high groundwater flow and small grain size was associated with low groundwater flow. Although this study provides a comprehensive analysis of the geologic controls on observed spatial variability of groundwater discharge to this small lake, this method would be difficult to apply in more heterogeneous large lake environments. The only study to show a relationship between nearshore groundwater discharge spatial patterns and hydrogeologic features in a large glacial lake environment is Ji et al. (2017). This study conducted a regional scale $^{222}\text{Rn}$ survey and electrical resistivity measurements along 17 km shoreline in Nottawasaga Bay, Lake Huron and demonstrated that groundwater discharge ‘hotspots’ (areas of elevated groundwater discharge) occur in areas where tunnel channel aquifers intercept the shoreline. Tunnel channel aquifers are hydrogeologic features that are characteristic to glacial environments.

Overall, a review of studies linking groundwater discharge field measurements with hydrogeologic features shows the variety of factors that can influence spatial patterns of
groundwater discharge in different environments, and on different scales. Hydrogeologic controls that may influence spatial variability of groundwater inputs to large glacial lakes are largely unknown, particularly at the regional scale. Understanding the variability in groundwater discharge patterns to large lakes is challenging, but identifying areas with high groundwater discharge is needed to inform lake water quality management actions. In addition, regional scale field campaigns to characterize groundwater discharge to a large lake can be resource intensive, and there is a need to develop broadly applicable transferable knowledge to focus future investigations in large lake environments. This thesis will evaluate key hydrogeologic controls on the spatial variability of groundwater discharge measured using $^{222}$Rn as a tracer for quantification.

2.4 Lake Simcoe

The research presented in this thesis focuses on Lake Simcoe, Ontario which is a large glacial lake with an area of 722 km$^2$ and approximately 240 km of shoreline. Lake Simcoe is located within the Laurentian Great Lakes Basin between the northwest shoreline of Lake Ontario and the southeast shoreline of Lake Huron’s Georgian Bay, and represents the largest lake in Ontario aside from the Great Lakes themselves (shown in Figure 2.2; North et al., 2013). The Lake Simcoe watershed features seventeen subwatersheds, drained by thirty-five tributaries, including five major rivers that drain approximately 60% of the watershed area (i.e. Beaver River, Talbot River, Black River, Holland River, and Pefferlaw River; Eimers et al., 2005). The Lake Simcoe watershed also includes a wide range of geologic environments with varying surficial sediment, aquifer systems, topography, and depositional/erosional features. The surficial geology along the Lake Simcoe shoreline is dominated by areas of continuous sand, gravel, and diamicton (glacial till) in the south and east, and thin sediment and exposed Paleozoic bedrock in the north (Ontario Ministry of Northern Development and Mines, 2016). Large moraine features within the watershed (i.e. the Oro Moraine, and Oak Ridges Moraine) represent topographic high points and regional recharge features (AquaResource Inc., 2013; Earthfx Inc., 2014a, 2008; Genivar Inc., 2013). Large erosional channels along the lake’s southern shoreline are also important hydrogeologic features (Earthfx Inc., 2013, Earthfx Inc. & Gerber Geosciences, 2008). Both Kempenfelt Bay, which is situated within a narrow and deep sand-gravel corridor on
the west side of the lake, and Cook’s Bay, which is found within a shallow and wide silt-organics area along the lakes southern shoreline, are large channel valley features (Lake Simcoe Region Conservation Authority, 2010b).

In the past several decades, the Lake Simcoe watershed has been subject to rapid urbanization and agricultural development, with the population doubling between 1981 and 2005 (Palmer et al., 2011). Additionally, the southern shoreline of Lake Simcoe represents the northern extent of the Greater Toronto Area (Canada’s largest urban center) and has become a popular home for commuters and cottagers; with a transient population of approximately 50,000 people during the summer months. The changing land use within the watershed has caused deterioration of lake water quality and ecosystem health. This is a major issue for Lake Simcoe’s fisheries and tourism industries that generate an estimated $200 million annually (North et al., 2013; Palmer et al., 2011). The greatest issue is nutrient loading to the lake, with increases in phosphorous loading being attributed to excessive algal growth and decreasing dissolved oxygen that has had detrimental effects on the lake’s cold-water fish population (Winter et al., 2007; Young et al., 2011). The source of phosphorous and other pollutants of concern have been linked to current and historic land use within the watershed. Nutrient (predominantly phosphorous) loading along the lake’s southern shoreline area has been associated with agricultural activities (Eimers et al., 2005). O’Connor et al. (2013) estimated that septic systems within a 100 m band around the lake discharge approximately 3.87 tonnes of phosphorous annually. In the City of Barrie, the largest urban center in the watershed, and an area that has been historically associated with industrial activity, Roy & Malenica (2013) found elevated concentrations of phosphorous, nitrate, chloride, chlorinated solvents, and petroleum compounds in shallow nearshore groundwater.

In 2009 the Lake Simcoe Protection Plan (LSPP) was passed by the Government of Ontario, with an urgency to protect and restore the water quality and ecosystem health of Lake Simcoe (MOECC, 2009). Current and historical water quality management initiatives in the Lake Simcoe watershed have focused on point source and tributary pollutant sources, with few studies considering groundwater inputs to the lake (Eimers et al., 2005; O’Connor et al., 2006; O’Connor et al., 2013; Roy & Malenica, 2013; Winter et al., 2007; Young et
al., 2011). Results of seismo-stratigraphic surveys in Kempenfelt Bay (the western area of Lake Simcoe) conducted by Lewis et al. (2007) showed evidence of the presence of large submarine hollows in the lake floor, which may be locations of offshore groundwater discharge to the bay. Subsequently, North et al. (2013) hypothesized that observed differences in the $O_2$ concentrations between Kempenfelt Bay and the lake’s main basin may be the result of groundwater discharge from these submarine pathways to Kempenfelt Bay. Aside from this work focused on offshore groundwater discharge, there has been no attempt to quantify direct nearshore groundwater discharge into Lake Simcoe, despite recognition of its potential to impact lake water quality. This thesis will aim to address knowledge gaps by evaluating the quantity, spatial variability, and potential impact of groundwater discharge through the use of both field and model investigations.
2.5 References


Chapter 3

3 Hydrogeologic Controls on Groundwater Discharge to a Large Glacial Lake

3.1 Introduction

Increased urban development, and intensification of agricultural and industrial activities over recent decades have led to the deterioration of water quality in lakes worldwide (International Institute for Sustainable Development, 2017; Palmer et al., 2011). While lake water quality management efforts typically focus on reducing point pollution sources and pollutant inputs from tributaries, groundwater discharge can also be an important pathway for delivering pollutants into lakes (Burnett et al., 2006; International Joint Commission, 2013; Stets et al., 2010). Although the magnitude of groundwater inputs is often a small component of the lake water budget, concentrations of dissolved pollutants can be elevated in groundwater compared to receiving surface waters leading to high pollutant loads from groundwater discharge (Burnett et al., 2008; Dimova & Burnett, 2011; Moore, 2010; Robinson, 2015). For instance, shallow urban groundwater adjacent to Lake Simcoe, a large inland lake in southern Ontario and the focus of this study, has been shown to have high concentrations of pollutants including nitrate, ammonium, phosphorous, and chlorinated solvents (Roy & Malenica, 2013).

The impact of groundwater discharge on lake water quality is often poorly understood, although studies have demonstrated that groundwater pollutant inputs can adversely impact the quality of receiving surface waters and ecosystem health (Haack et al., 2005; Kazmierczak et al., 2016; Lewandowski et al., 2015; Meinikmann et al., 2015; Roy & Malenica, 2013). The lack of understanding regarding the contribution of groundwater discharge is in part due to the challenge of quantifying groundwater discharge as it is generally characterized by low specific fluxes as well as high spatial and temporal variability (Burnett et al., 2006; Dimova et al., 2015; Kornelsen & Coulibaly, 2014; Russoniello et al., 2013).

Local and regional scale field methods for quantifying groundwater discharge into lakes are available and selecting an appropriate method depends on the characteristics of the
study site and study objectives. Methods that have been applied include seepage meters, hydraulic gradient/piezometer measurements, numerical groundwater models, electrical resistivity tomography (ERT) imaging, thermal imaging, and geochemical/isotopic tracer methods (Burnett et al., 2006; Dimova et al., 2015; Kidmose et al., 2015; Meinikmann et al., 2013; Santos et al., 2008; Smith & Swarzenski, 2012). A geochemical tracer method that has been widely applied to estimate groundwater discharge in lake, marine, and riverine environments is the naturally occurring radium isotope, radon-222 ($^{222}$Rn) (e.g. Cable et al., 1996; Burnett & Dulaiova, 2003; Dimova & Burnett, 2011; Dimova et al., 2013; Dulaiova et al., 2010; Santos et al., 2009). Geochemical tracers, such as $^{222}$Rn, are generally more suitable for regional scale groundwater discharge investigations, as opposed to seepage meters and hydraulic gradient measurements, which are local scale techniques (Burnett et al., 2006; Cherkauer & McKereghan, 1991; Ji et al., 2017; Mulligan & Charette, 2006).

Groundwater discharge to lakes is often highly spatially variable. Understanding this variability is challenging but identifying areas with high groundwater discharge is needed to develop effective lake water quality management actions. Groundwater can enter a lake indirectly, through tributaries that are groundwater fed and flow into the lake, or directly through discharge from (often multiple) aquifer layers. This study focuses specifically on direct nearshore groundwater discharge. There are several factors known to control the spatial variability of direct groundwater discharge. These include regional aquifer sediment type (i.e., permeability, hydraulic conductivity, fractures), nearshore topography, and the hydraulic gradient between the groundwater piezometric levels and the lake (Burnett et al., 2006; Cherkauer & Hensel, 1986; Feinstein & Reeves, 2010; Kishel & Gerla, 2002; Kornelsen & Coulibaly, 2014; Mulligan & Charette, 2006; Russoniello et al., 2013). For example, the Laurentian Great Lakes are all known discharge features because they are located within topographic low areas, but Lake Michigan has the highest direct groundwater discharge in part because of the prevalence of high permeability nearshore aquifer sediment (Grannemann et al., 2000). Studies have shown that shoreline configuration, and anthropogenic modifications to the shoreline (e.g., manmade canals and erosion control structures) can also affect the spatial distribution of direct groundwater discharge. In lake environments, the distance offshore and lake bathymetry also play a role
In general, direct groundwater discharge into a lake from a shallow unconfined aquifer will be highest close to the shoreline and decrease further offshore but this depends on the aquifer homogeneity. In addition, depending on the lake bathymetry and nearshore hydrogeology, deeper confined layers may intersect the lake bed and discharge further offshore (Fukuo & Kaihotsu, 1988; McBride & Pfannkuch, 1975). Areas with preferential or higher groundwater inputs (herein called groundwater discharge ‘hotspots’) may also have higher groundwater-derived pollutant inputs, depending on the quality of the discharging groundwater.

Although $^{222}\text{Rn}$ has been used previously to quantify direct groundwater discharge at different scales into lake environments, these studies often provide limited insight into the factors controlling the observed spatially variability, particularly the hydrogeological controls (Burnett et al., 2002; Cable et al., 1996; Corbett et al., 1997; Dimova & Burnett, 2011; Dulaiova & Burnett, 2006; Santos et al., 2008). There is a need to examine the relationship between the geologic environment and the observed spatial distribution of nearshore groundwater discharge to develop transferrable knowledge that can be applied to more easily identify areas of high direct groundwater discharge to lakes. Schneider et al. (2005) investigated groundwater discharge into a glacial lake using seepage meter measurements, and found that the deeper regional groundwater flow system had a larger influence on the spatial distribution of groundwater discharge than the local surficial sediment type. More recently, Ji et al. (2017) conducted a regional scale $^{222}\text{Rn}$ survey and electrical resistivity measurements along 17 km shoreline in Nottawasaga Bay, Lake Huron and demonstrated a relationship between groundwater discharge ‘hotspots’ and tunnel channel aquifer features that are characteristic to glacial environments. These findings vary from other studies conducted in more homogeneous geological settings that indicate the importance of surficial sediment type and topography (e.g., Tecklenburg & Blume, 2017).

Lake Simcoe, a large glacial lake in Southern Ontario (Figure 3.1), represents an ideal setting for this study due to the large variation in the geology around the lake with surficial sediment ranging from sand and gravel to exposed bedrock, as well as large erosional channels, and moraine features. Over the past several decades rapid population growth
combined with intensification of agriculture in the Lake Simcoe watershed has led to the deterioration of the ecological health of Lake Simcoe (North et al., 2013; Young et al., 2011). For instance, increased nutrient loading to the lake, particularly phosphorous (P), has stimulated algal growth, reduced dissolved oxygen levels and in turn impaired cold-water fish habitats (Eimers et al., 2005; North et al., 2013; Winter et al., 2007). Deterioration in recreational water quality and decline in fish population is a major concern as the Lake Simcoe fisheries and tourism industries generate an estimated $200 million annually (Eimers et al., 2005; North et al., 2013; Winter et al., 2007). Lake water quality management actions have largely focused on point source and tributary inputs in an effort to reduce nutrient (and other pollutant) inputs, while little attention has been given to quantifying groundwater inputs (Roy & Malenica, 2013; Winter et al., 2007). Seismic surveys have revealed the presence of large submarine hollows at the bottom of Kempenfelt Bay (Lewis et al., 2007). These are speculated to be zones of groundwater discharge causing differences in oxygen concentrations and affecting habitat for cold-water fish (North et al., 2013). Aside from this work focused on offshore groundwater discharge, there has been no attempt to quantify direct nearshore groundwater discharge into Lake Simcoe.

The study objectives are to (i) characterize the large scale spatial distribution of direct groundwater discharge to Lake Simcoe including identification of groundwater discharge ‘hotspots’, (ii) estimate the total direct nearshore groundwater input and compare it to tributary inputs, and (iii) evaluate potential links between observed large scale groundwater discharge patterns and regional hydrogeologic features in a large glacial lake environment. To our knowledge, no prior studies have used $^{222}$Rn to quantify the total nearshore direct groundwater discharge to a lake as large as Lake Simcoe (722 km$^2$), and to assess the geological controls on the large scale spatial variability in groundwater discharge to a glacial lake. In this Chapter, results from $^{222}$Rn surveys that were conducted along the Lake Simcoe shoreline are first presented with groundwater discharge ‘hotspots’ identified. The total direct nearshore groundwater discharge to the lake is then quantified and compared with tributary inputs. Finally, field results are used to evaluate the influence of surficial geology, regional recharge features, tunnel channel deposits, and groundwater-fed streams on the observed spatial groundwater discharge patterns. Insight into key controls on
groundwater discharge in large glacial lake environments is needed to improve understanding and quantification of this pollutant delivery pathway and thus to develop more effective lake water quality management strategies.

3.2 Field Site

Lake Simcoe is the largest lake in Ontario, aside from the Laurentian Great Lakes, with an area of 722 km² and approximately 240 km of shoreline (Eimers et al., 2005). Lake Simcoe’s southern shoreline represents the northern-most extent of the Greater Toronto area, which is the largest urban center in Canada (North et al., 2013). Generally, the lake is divided into three areas: Kempenfelt Bay, Cook’s Bay, and the main basin (Figure 3.1) (North et al., 2013; Palmer et al., 2011; Winter et al., 2007). The Lake Simcoe watershed area features thirty-five tributaries, including five major rivers that drain approximately 60% of the total area (North et al., 2013). Lake Simcoe provides a complex and diverse environment to study geologic controls on direct groundwater discharge. The Lake Simcoe watershed includes a wide range of geologic environments with varying surficial sediment, aquifer systems, topography, and depositional/erosional features. The surficial geology along the Lake Simcoe shoreline is dominated by areas of continuous sand, gravel, and diamicton (glacial till) in the south and east, and thin sediment and exposed Paleozoic bedrock in the north (Figure 3.2). Large moraine features within the watershed, most notably the Oro Moraine near the northwest shoreline (405 masl) and the Oak Ridges Moraine near the southern shoreline (340 masl), represent topographic high points and regional recharge features (AquaResource Inc., 2013; Earthfx Inc., 2014a, 2008; Genivar Inc., 2013). Large erosional channels in the regionally extensive Newmarket Till aquitard layer are also important hydrogeology features particularly along the lake’s southern shoreline. The channels are characterized by a fining upward sequence of sediment, that often has hydraulic conductivities an order of magnitude higher than adjacent sediments (Earthfx Inc., 2013, Earthfx Inc. & Gerber Geosciences, 2008). Both Kempenfelt Bay, which is situated within a narrow and deep sand-gravel corridor, and Cook’s Bay, which is found within a shallow and wide silt-organics area, are large channel valley features (Lake Simcoe Region Conservation Authority, 2010b). $^{222}\text{Rn}$ surveys were conducted along 80% of shoreline of Lake Simcoe, with multiple surveys conducted in areas
identified as groundwater discharge hotspots (Figure 3.1). Detailed description of the characteristics of the groundwater discharge hotspot areas are provided in Appendix 1.

Figure 3.1: (a) Map of the location of Lake Simcoe in southern Ontario, Canada, and (b) map showing shoreline where $^{222}$Rn boat surveys were conducted (denoted by white lines along shoreline). Black boxes indicate focus areas for the study where groundwater discharge hotspots were repeatedly identified (red dots).
Figure 3.2: Surficial geology around Lake Simcoe with groundwater sampling (+), sediment sampling (○), and creek water sampling (●) locations shown. Figure modified from Ontario Ministry of Northern Development and Mines (2016).

3.3 Methods

3.3.1 Field $^{222}\text{Rn}$ Measurements

3.3.1.1 $^{222}\text{Rn}$ Boat Surveys

Boat surveys during which $^{222}\text{Rn}$ was continuously measured were conducted between June 2015 and July 2018 to quantify nearshore groundwater discharge to Lake Simcoe. Surveys were performed along 190 km of the lake’s shoreline (Figure 3.1). $^{222}\text{Rn}$ is produced from the radioactive decay of radium ($^{226}\text{Ra}$) and is naturally occurring in almost all aquifer materials including glacial sediment (Je & Eyles, 1998; Schmidt et al., 2010; Schmidt & Schubert, 2007). $^{222}\text{Rn}$ is a suitable tracer for evaluating groundwater discharge because it (i) is a chemically inert gas, (ii) is often present in groundwater in concentrations that are
3-4 orders of magnitude higher than receiving surface waters, and (iii) has a half-life of 3.82 days which is suitable for nearshore measurements (Burnett et al., 2013; Burnett et al., 2001; Cable et al., 1996).

Surveys were conducted by continuously sampling $^{222}$Rn in lake water from a boat travelling between 3 – 5 km/h along the shoreline at an offshore distance of 50 - 200 m. A submersible pump was used to pump lake water continuously from 0.5 – 1 m below the lake water surface to an air-water exchanger (RAD AQUA; Durridge Co.). As the water was passed through the RAD AQUA, $^{222}$Rn reached equilibrium between the air and water in a closed air loop system. $^{222}$Rn concentrations were then measured by RAD7 portable electronic radon detectors (Durridge Co., Burnett et al., 2002; Burnett et al., 2001; Dulaiova et al., 2005). Following Dulaiova et al. (2005), multiple RAD7 units were used in parallel to decrease the initial equilibration time, improve sensitivity to changes in concentrations, and reduce the overall measurement errors. Several equipment combinations were used including 5 RAD7s and 1 RAD AQUA, 4 RAD7s and 1 RAD AQUA, and 2 RAD7s with 2 RAD AQUAs. Details on the instrument response time and measurement error for each system set up are provided in Appendix 2. The reported error for each $^{222}$Rn measurement is the standard deviation ($\sigma$) following Poisson statistics (Taylor, 1982; Durridge Co.). Measurements were taken over a 15 minute integration cycle as the boat travelled along the shoreline. Electrical conductivity measurements were also taken alongside $^{222}$Rn measurements, but no clear relationship was observed as in-lake conductivity spatial trends were found to be overwhelmed by urban sources (shown in Appendix 3). The track of each offshore survey was recorded using a handheld GPS unit (Trimble GEO5T handheld, Trimble). As in-lake $^{222}$Rn concentrations were found to vary temporally, potential groundwater discharge ‘hotspot’ areas (shown in Figure 3.1) were surveyed multiple times to ensure the repeatability of results.

### 3.3.1.2 Groundwater $^{222}$Rn Sampling

To estimate the magnitude of nearshore groundwater discharge from in-lake $^{222}$Rn concentrations, $^{222}$Rn was also measured in nearshore shallow groundwater around Lake Simcoe. Groundwater $^{222}$Rn concentrations were measured using two methods: shallow groundwater samples, and sediment equilibration experiments (Burnett et al., 2007).
Groundwater samples were collected using temporary, shallow, nearshore wells installed at 20 beaches around Lake Simcoe from 2015-2017 (Figure 3.2). Dissolved $^{222}$Rn decays and re-equilibrates with radium in the aquifer sediments as groundwater travels towards the lake (Schmidt et al., 2010). As the half-life of $^{222}$Rn is relatively short (3.82 d), groundwater samples were taken within 5 - 10 m of the shoreline so that the sample would accurately reflect the $^{222}$Rn concentration in groundwater that is discharging to the lake (Burnett et al., 2013). Generally, three wells were installed approximately 5-10 m apart in an along-shore transect to capture the spatial variability in groundwater $^{222}$Rn concentrations at each site. Samples were analyzed using the RAD H2O (Durridge Co.) system with a RAD7 detector (Durridge Co.). The second method of measuring groundwater $^{222}$Rn concentrations is based on a sediment equilibration method (Chanyotha et al., 2014). For this method, nearshore lakebed sediment was collected in 250 mL airtight bottles and left to equilibrate with ambient lake water in the closed bottle for at least 20 d (or 5 x $^{222}$Rn half-life). Following equilibration, the samples were analyzed using the RAD H2O (Durridge Co.) system and RAD7 detector (Durridge Co.). Groundwater $^{222}$Rn concentrations were estimated using the equilibrium concentration of $^{222}$Rn in the water after the 20 d period and the sediment porosity and bulk density. Samples for the sediment equilibration method were collected from 13 beach and dock sites around the lake (Figure 3.2).

3.3.1.3 Tributary $^{222}$Rn Sampling

$^{222}$Rn concentrations in tributaries discharging into Lake Simcoe were measured to qualitatively evaluate the contribution of tributary inputs on the measured in-lake $^{222}$Rn concentrations (Figure 3.2). $^{222}$Rn concentrations were measured using a continuous sampling system whereby water was pumped continuously from the middle of a tributary to a RAD AQUA air-water exchanger, connected to one RAD7 unit (Burnett et al., 2010). The system was run for approximately 2 hours at each site using a 15-minute measurement cycle.
3.3.2 $^{222}$Rn Mass Balance

Nearshore groundwater discharge rates along the shoreline were estimated using measured in-lake $^{222}$Rn and groundwater $^{222}$Rn concentrations as inputs for a mass balance model (e.g. Burnett et al., 2001; Burnett & Dulaiova, 2003; Cable et al., 1996; Corbett et al., 1997). The model considers the various sources and sinks of $^{222}$Rn for a well-mixed surface water volume. Sources considered in the model include (1) $^{222}$Rn input from groundwater discharge ($J_{GW}$, dpm m$^{-2}$ d$^{-1}$), (2) diffusion of $^{222}$Rn from lake bed sediment ($J_{diff}$, dpm m$^{-2}$ d$^{-1}$), and (3) production of $^{222}$Rn from $^{226}$Ra decay ($J_{prod}$, dpm m$^{-2}$ d$^{-1}$). In lake environments, $J_{prod}$ is negligible as $^{226}$Ra concentrations are low in freshwater and therefore this source could be neglected (Dulaiova & Burnett, 2008; Moore, 1996). In addition, $J_{diff}$ is also negligible where there is advective flux of groundwater across the sediment-water interface (Dimova et al., 2013; Dulaiova et al., 2010; Ji et al., 2017). Losses considered in the model include (1) evasion of $^{222}$Rn to the atmosphere ($J_{atm}$, dpm m$^{-2}$ d$^{-1}$), (2) decay of $^{222}$Rn ($J_{decay}$, dpm m$^{-2}$ d$^{-1}$), and (3) offshore mixing with low $^{222}$Rn waters ($J_{mix}$, dpm m$^{-2}$ d$^{-1}$). Although $J_{mix}$ is often neglected for non-tidal lake environments with studies showing that $J_{atm}$ accounts for the largest loss of $^{222}$Rn from the coastal water volume (Corbett et al., 1997; Santos et al., 2008; Schmidt & Schubert, 2007; Tuccimei et al., 2005), $J_{mix}$ was considered in our mass balance model given the large size of Lake Simcoe and therefore potential importance of coastal processes driving offshore mixing.

Following the method outlined by Dulaiova et al. (2010), in-lake $^{222}$Rn concentrations and groundwater $^{222}$Rn concentrations were used to calculate the groundwater discharge rates ($Q$, m$^3$d$^{-1}$) along the Lake Simcoe shoreline using the following:

$$Q = \frac{C_{sw}V}{\frac{C_{gw}}{\tau}}$$ (3-1)

where $C_{sw}$ (dpm m$^3$) is the in-lake $^{222}$Rn concentration corrected for sources and losses, $C_{gw}$ (dpm m$^3$) is the groundwater $^{222}$Rn concentration, $V$ (m$^3$) is the surface water volume over which the mass balance was applied, and $\tau$ (d) is the rate at which the surface water volume is flushed. The surface water volume ($V$) was calculated using the average water column (lake) depth ($z$), distance offshore, and length of shoreline travelled during each
222Rn measurement cycle. The flushing rate was assumed to be equal to the mean life of 222Rn, which is 5.53 d (Burnett et al., 2013). As 222Rn concentrations in groundwater often exhibit high spatial variability (Dimova et al., 2013), detailed analysis was conducted to determine a representative value for $C_{GW}$ (see Section 3.4.1.2 for further details).

The in-lake 222Rn concentration ($C_{sw}$, dpm m$^{-3}$) was corrected for 222Rn losses using the following:

1. Evasion of 222Rn to the atmosphere ($J_{atm}$, dpm m$^{-2}$d$^{-1}$) was calculated using (Burnett & Dulaiova, 2003; Dulaiova et al., 2010; Macintyre et al., 1995):

$$J_{atm} = k(C_{sw} - \alpha C_{air})$$ (3-2)

where $C_{air}$ (dpm m$^{-3}$) is the measured ambient air 222Rn concentration and $\alpha$ (dimensionless) is the Ostwald’s solubility coefficient. Based on ambient air 222Rn measurements taken at the start of each survey, $C_{air}$ was set to be 500 dpm m$^{-3}$. The gas transfer coefficient, $k$ (m d$^{-1}$), was calculated by (Macintyre et al., 1995):

$$k(600) = 0.45 \times u_{10}^{1.6} \times \left(\frac{S_c}{600}\right)^{-b}$$ (3-3)

where $u_{10}$ (m s$^{-1}$) is the wind speed which was taken from the nearest Environment Climate Change Canada weather station (located 5 - 28 km away from a survey site), $b$ is 0.5 for wind speeds greater than 3.6 m s$^{-1}$ and 0.667 for wind speeds less than 3.6 m s$^{-1}$ (Baskaran, 2016; Dimova et al., 2013; Macintyre et al., 1995), and $S_c$ is the Schmidt number for 222Rn, which is 1000 (Baumert et al., 2005). The Oswald solubility coefficient, $\alpha$, is dependent on the temperature at the air-water interface (°C) and was calculated by (Dimova et al., 2013; Macintyre et al., 1995):

$$\alpha = 0.105 + 0.405 \exp(-0.05027T)$$ (3-4)

2. Decay of 222Rn in the surface water volume, ($J_{decay}$, dpm m$^{-2}$d$^{-1}$) was calculated by (Schmidt & Schubert, 2007):

$$J_{decay} = z\lambda_{Rn}C_{sw}$$ (3-5)
where $\lambda_{\text{Rn}}$ (d$^{-1}$) is the decay rate of $^{222}\text{Rn}$ which is 0.181 d$^{-1}$.

3. The horizontal offshore mixing loss, $J_{\text{mix}}$ (dpm m$^{-2}$ d$^{-1}$) was estimated using an iterative method described in Santos et al. (2008). Shore-perpendicular $^{222}\text{Rn}$ and conductivity transect data was collected in three different locations of Lake Simcoe (Kempenfelt Bay, Cook’s Bay, and the main basin) to calculate the offshore concentration gradients at each location (Appendix 7). Both the $^{222}\text{Rn}$ (non-conservative tracer) and the conductivity (conservative tracer) gradients were used in the steady state advection diffusion equation to estimate the horizontal eddy diffusion coefficient $K_h$ (m$^2$ d$^{-1}$):

$$K_h \frac{\partial^2 C}{\partial x^2} - \omega \frac{\partial C}{\partial x} - \lambda C = 0 \quad (3-6)$$

where $\omega$ (m d$^{-1}$) is the horizontal surface advection, $C$ is the concentration of the tracer, and $\lambda$ is the $^{222}\text{Rn}$ decay constant. When using the equation for the offshore conductivity transect, the decay term was removed. To calculate $K_h$ the measured offshore $^{222}\text{Rn}$ concentrations must be corrected for $J_{\text{decay}}$ and $J_{\text{atm}}$ over the time taken for the $^{222}\text{Rn}$ to travel to the measurement location. As such an iterative approach was used whereby the $^{222}\text{Rn}$ concentrations were corrected for $J_{\text{decay}}$ and $J_{\text{atm}}$ and then used in equation 3-6 to estimate a new $K_h$ and $\omega$. Iterations were continued until convergence was reached. The final $K_h$ value was then used to calculate $J_{\text{mix}}$ using (Burnett et al., 2008; Moore, 2000; Santos et al., 2008):

$$J_{\text{mix}} = -K_h \left( \frac{C_{S+1} - C_{S-1}}{2\Delta x} \right) x \frac{A_{CS}}{A_{LB}} \quad (3-7)$$

where $\frac{C_{S+1} - C_{S-1}}{2\Delta x}$ (dpm m$^{-3}$ m$^{-1}$) is the measured offshore $^{222}\text{Rn}$ gradient, and $\frac{A_{CS}}{A_{LB}}$ is the ratio of the along-shore cross sectional area ($A_{CS}$, m$^2$) to the lake bed area ($A_{LB}$, m$^2$) for the surface water volume represented by the cycle measurement.

Groundwater discharge values ($Q$) were calculated for each 15-minute measurement cycle and the values were divided by the shoreline length travelled during the respective cycle to obtain the discharge rate per unit of shoreline (m$^3$ d$^{-1}$ m$^{-1}$). Reported uncertainties for each groundwater discharge value represent the propagation of uncertainties associated with
each source and sink term. As tributary inputs were not included as a $^{222}$Rn source term in the mass balance, calculations were not performed for shoreline areas within 500 m of a tributary.

3.4 Results and Discussion

3.4.1 $^{222}$Rn Field Measurement Results

3.4.1.1 In-lake $^{222}$Rn Concentrations

A summary of the in-lake $^{222}$Rn concentrations along the Lake Simcoe shoreline is shown in Figure 3.3. $^{222}$Rn concentrations exhibit high spatial variability along the shoreline with values ranging from $0.00 \pm 0.15$ to $5.10 \pm 2.12$ dpm L$^{-1}$. Combining data from surveys conducted on separate days was challenging as in-lake $^{222}$Rn concentrations vary temporally in response to environmental (e.g., precipitation, wind) and lake (e.g., waves) conditions. To address this, multiple surveys were conducted along most of the shoreline and an 8-day stationary time series test was performed at a location on the southern shoreline to evaluate the response of in-lake $^{222}$Rn concentrations to atmospheric and lake conditions. The stationary test showed that $^{222}$Rn concentrations sharply decreased following sustained onshore wind speeds greater than $20$ km h$^{-1}$ for 6-12 hours, and also decreased for over 12 hours following a 10 mm precipitation event (Appendix 4 for additional details). Based on these results, surveys that were conducted within at least 12 hours of these high wind and precipitation conditions were discounted. The concentrations shown in Figure 3.3 are the average concentration values for all other surveys. To aid discussion of the results, the lake has been divided into the eight main areas: Oro North, Kempenfelt Bay North, Kempenfelt Bay South, Cook’s Bay West, Cook’s Bay East, Georgina, East Shore, and North Shore (Figure 3.3).

Five shoreline areas were consistently identified during repeat surveys to have elevated in-lake $^{222}$Rn concentrations (referred to as $^{222}$Rn hotspot areas). These areas, and the average and maximum $^{222}$Rn concentrations measured in the areas, are: Shingle Bay (SB; average = $2.26 \pm 0.53$ dpm L$^{-1}$; max = $3.03 \pm 0.21$ dpm L$^{-1}$), Johnson’s Beach (JB; average = $1.48 \pm 0.52$ dpm L$^{-1}$; max = $3.57 \pm 0.38$ dpm L$^{-1}$) Keswick Beach (KB; average= $3.47 \pm 0.89$ dpm L$^{-1}$; max= $5.10 \pm 1.12$ dpm L$^{-1}$), Duclos Point (DP; average= $2.62 \pm 0.52$ dpm L$^{-1}$;
max=3.80 ± 0.94 dpm L⁻¹), and Thorah Centennial Park (TCP; average=3.24 ± 0.41 dpm L⁻¹; max=3.92 ± 0.30 dpm L⁻¹) (Figure 3.3). The average $^{222}$Rn concentration at the hotspot areas were 15 - 40% higher than in-lake $^{222}$Rn concentrations measured along the adjacent shoreline. Elevated in-lake $^{222}$Rn concentrations were observed in the Kempenfelt Bay South (max= 3.30 ± 0.39 dpm L⁻¹) and North Shore areas (max= 4.93 ± 0.71 dpm L⁻¹), however repeat surveys were not conducted along these shoreline areas. The following discussion of $^{222}$Rn survey results focuses on areas repeatedly identified as $^{222}$Rn hotspots. Data from individual surveys together with the weather conditions during each survey are provided in Appendix 6.

$^{222}$Rn hotspots were observed in sheltered bay areas (e.g., Shingle Bay [SB], Duclos Point [DP]) as well as more exposed shoreline areas (e.g. Johnson’s Beach [JB], Keswick Beach [KB]) where there was higher potential for $^{222}$Rn losses due to strong wind and currents (i.e. higher offshore mixing). Hotspots were also observed in both shallow and deep nearshore waters (e.g., 2 – 4 m depth in Oro North, and 6-20 m depth in Kempenfelt Bay; Ontario Ministry of Natural Resources, 2006). The in-lake $^{222}$Rn concentrations were also influenced by indirect groundwater discharge with high $^{222}$Rn concentrations observed adjacent to some tributaries. For instance, the highest average $^{222}$Rn concentrations were measured at Keswick Beach (KB, 3.47 ± 0.89 dpm L⁻¹) and Thorah Centennial Park (TCP, 3.24 ± 0.41 dpm L⁻¹), which are located adjacent to the Maskinonge River and the Talbot River/ Beaver River, respectively. The influence of indirect groundwater discharge on in-lake $^{222}$Rn concentrations is discussed in Section 3.4.1.3.
Figure 3.3: Average in-lake $^{222}\text{Rn}$ concentrations measured during boat surveys from June 2015- July 2018. Radon hotspot areas identified on repeat surveys are labelled (white letters): Shingle Bay (SB), Johnson’s Beach (JB), Keswick Beach (KB), Duclos Point (DP) and the Thorah Centennial Park (TCP).

3.4.1.2 Groundwater $^{222}\text{Rn}$ Concentrations

Determining a representative $^{222}\text{Rn}$ groundwater endmember concentration ($C_{gw}$) is a key challenge in using $^{222}\text{Rn}$ as a tracer to quantify groundwater discharge rates as groundwater concentrations typically exhibit high spatial variability (Dimova et al., 2013; Schmidt et al., 2010). Uncertainty in $C_{gw}$ can be reduced by using multiple methods to measure the endmember value (e.g., monitoring well and piezometer sampling, seepage meter sampling, and sediment equilibration experiments) and collecting multiple samples to quantify the spatial heterogeneity (Burnett et al., 2007). For our study, groundwater samples were collected at 20 public beaches, and sediment samples (for sediment equilibration experiments) were collected at 13 beaches from 2015-2017 (locations shown
in Figure 3.2; sample results provided in Appendix 5). $^{222}\text{Rn}$ concentrations in groundwater samples ranged from 50.0 ± 5.0 dpm L$^{-1}$ to 462.8 ± 43.7 dpm L$^{-1}$, with an average concentration of 178.3 ± 17.2 dpm L$^{-1}$. $^{222}\text{Rn}$ concentrations derived from sediment equilibration experiments were slightly higher ranging from 132.6 ± 6.3 dpm L$^{-1}$ to 478.8 ± 10.0 dpm L$^{-1}$ with an average value of 321.6 ± 9.0 dpm L$^{-1}$. For the 9 beaches where both groundwater and sediment samples were collected (results shown in Figure 3.4), the average groundwater sample concentration was 243.7 ± 18.0 dpm L$^{-1}$, while sediment sample $^{222}\text{Rn}$ results had an average concentration of 342.6 ± 8.5 dpm L$^{-1}$ (see Appendix 5 for direct comparison of results). This difference may be a result of shallow recharge to the shallow aquifer layers along the shoreline. The higher concentrations from sediment samples is consistent with other studies (Burnett & Dulaiova, 2006).

To evaluate whether the observed spatial variability in the in-lake $^{222}\text{Rn}$ concentrations is due to large-scale spatial variability in the groundwater $^{222}\text{Rn}$ concentrations, the relationship between these concentrations was examined (Figure 3.4). The correlation between both the groundwater sample and sediment equilibration sample $^{222}\text{Rn}$ concentrations, and the in-lake $^{222}\text{Rn}$ concentrations measured near each groundwater/sediment sampling location is poor. This indicates that the variability in the in-lake $^{222}\text{Rn}$ concentrations is not controlled by the variability in the groundwater concentrations and may be due to varying groundwater discharge along the shoreline. The measured groundwater concentrations (groundwater and sediment samples) were also grouped based on the surficial geology type (gravel, sand, diamicton, and silt; surficial geology shown in Figure 3.2) and the surficial permeability (high and low permeability; Ontario Ministry of Mines and Development (2016)) at each of the groundwater and sediment sampling sites. No relationships were found between the $^{222}\text{Rn}$ groundwater concentrations and the surficial geology or permeability (Figure 3.4). Importantly, however, low in-lake $^{222}\text{Rn}$ concentrations generally occur in areas with low permeability sediments regardless of the groundwater $^{222}\text{Rn}$ concentration value (Figure 3.4c, d). As higher permeability surficial sediments may represent more productive shallow aquifers, this result supports the relationship between measured in-lake $^{222}\text{Rn}$ and the groundwater discharge rate (Feinstein & Reeves, 2010; Kornelsen & Coulibaly, 2014). While the relationship between the surficial permeability and in-lake concentration does not hold as
well for the sediment sample $^{222}$Rn concentrations (Figure 3.4d), the three highest in-lake $^{222}$Rn concentrations are all adjacent to sampling sites located in high permeability areas. Based on the poor correlation between the in-lake and groundwater $^{222}$Rn concentrations, together with the lack of relationship between the groundwater concentrations and the surficial geology and permeability, an average groundwater endmember $^{222}$Rn concentration of 234.8 ± 14.0 dpm L$^{-1}$ was used for the mass balance calculations (Equation 3-1). This value represents the average $^{222}$Rn concentration for all sampled beach sites, including both groundwater sample and sediment equilibrium sample results.

![Figure 3.4: Measured (a, c) groundwater sample and (b, d) sediment equilibrium sample $^{222}$Rn concentrations for all sampling locations plotted against average in-lake $^{222}$Rn concentrations measured during boat surveys. Points have been grouped by (a, b) surficial sediment type, and (c, d) surficial sediment permeability (Ontario Ministry of Northern Development and Mines, 2016).](image)

### 3.4.1.3 Tributary $^{222}$Rn Concentrations

Measured $^{222}$Rn concentrations in 11 tributaries discharging into Lake Simcoe ranged from 1.76 ± 0.26 dpm L$^{-1}$ to 15.68 ± 0.31 dpm L$^{-1}$ with an average of 8.32 ± 0.54 dpm L$^{-1}$ (shown
in Figure 3.5). The highest $^{222}$Rn concentrations were observed in Hawkestone Creek (HC; 15.68 ± 0.31 dpm L$^{-1}$), Talbot River (TR; 15.07 ± 1.74 dpm L$^{-1}$), Lovers Creek (LC; 9.11 ± 0.10 dpm L$^{-1}$), and Shelswell Creek (SC; 10.14 ± 0.15 dpm L$^{-1}$). Elevated $^{222}$Rn concentration in a tributary indicates that groundwater inputs into that tributary may be high (Mullinger et al., 2007). Measured in-lake $^{222}$Rn concentrations can be elevated in areas where a tributary with high $^{222}$Rn concentration is discharging indicating the impact of in-direct groundwater discharge on the lake water quality. From comparing measured tributary $^{222}$Rn concentrations and in-lake $^{222}$Rn concentrations, it is evident that some measured $^{222}$Rn hotspots are influenced by indirect groundwater discharge rather than only direct groundwater discharge. For example, $^{222}$Rn concentrations and thus lake water quality are likely influenced by indirect groundwater discharge along the Kempenfelt Bay South shoreline where four creeks discharge along the 16 km stretch (Hotchkiss Creek (HTC), Whisky Creek (WC), Lovers Creek (LC), and Hewitt’s Creek (not measured)) and tributary $^{222}$Rn concentrations were ~ 3-5 times higher than in-lake $^{222}$Rn concentrations (Figure 3.6).

Measured $^{222}$Rn hotspots at Keswick Beach (KB) and Thorah Centennial Park (TCP) are both adjacent to major rivers in the Lake Simcoe watershed: the Maskinonge River (MR) and the Talbot River (TR)/ Beaver River (BR) respectively. $^{222}$Rn concentrations in the Maskinonge River were a similar magnitude to the average lake $^{222}$Rn concentration in the area (~ 1-2 times higher) suggesting that while the Keswick Beach area may be under mild influence from indirect groundwater discharge, the primary source of $^{222}$Rn in this area is likely direct groundwater discharge. Conversely, in the Thorah Centennial Park area the results show that the high in-lake $^{222}$Rn concentrations are likely due to a combination of direct and indirect groundwater discharge, where the Talbot River has a strong influence (~ 5 times higher than lake) and the Beaver River has a mild influence (~1-2 times higher than lake) on measured lake $^{222}$Rn. Evaluating $^{222}$Rn loading from streams can also be useful to interpret the influence of in-direct groundwater discharge, however, stream discharge data was not available for the above-mentioned tributaries. Due to the influence of in-direct groundwater discharge on the in-lake $^{222}$Rn concentrations, mass balance calculations were not done for shoreline areas within 500 m of the mouth of a tributary.
Figure 3.5: $^{222}$Rn concentrations measured in tributaries discharging into Lake Simcoe. Some of these tributaries are located in $^{222}$Rn hotspot areas (boxes).

Figure 3.6: Average in-lake $^{222}$Rn concentrations plotted against tributary $^{222}$Rn concentrations. Colored areas represent ratios of tributary concentration and lake concentrations, to illustrate the degree to which tributaries are influencing nearby in-lake $^{222}$Rn concentrations (mild influence= blue, medium influence= orange, strong influence= red).
3.4.2 $^{222}$Rn Mass Balance Results

Groundwater discharge along the shoreline was calculated using $^{222}$Rn data for all non-discounted surveys (i.e. surveys that were not conducted within 12 hours of high wind and precipitation events) with the average groundwater discharge values shown Figure 3.7 (all groundwater discharge results shown in Appendix 6). Calculated groundwater discharge per meter of shoreline range from 0.01 ± 0.06 m$^3$ d$^{-1}$ m$^{-1}$ to 12.78 ± 0.85 m$^3$ d$^{-1}$ m$^{-1}$. Areas of high groundwater discharge (referred to as groundwater discharge hotspots) occur at Shingle Bay (SB; average= 1.60 ± 0.28 m$^3$ d$^{-1}$ m$^{-1}$), Johnson’s Beach (JB; average= 3.81 ± 0.91 m$^3$ d$^{-1}$ m$^{-1}$), Keswick Beach (KB; average= 4.33 ± 0.57 m$^3$ d$^{-1}$ m$^{-1}$), Duclos Point (DP; average= 4.56 ± 0.47 m$^3$ d$^{-1}$ m$^{-1}$), and Thorah Centennial Park (TCP; average=8.08 ± 0.61 m$^3$ d$^{-1}$ m$^{-1}$).

Groundwater discharge results generally show a similar pattern to the in-lake $^{222}$Rn data, where the same hotspot locations (i.e. Shingle Bay, Johnson’s Beach, Keswick Beach, Duclos Point, and Thorah Centennial Park) were identified as having groundwater discharge values 25-80% higher than adjacent shoreline areas. However, the magnitudes of the groundwater discharge are adjusted based on the survey conditions (i.e. lake depth, distance covered during survey cycle, atmospheric losses due to wind speed on survey day etc.). For example, $^{222}$Rn concentrations in the Oro North (lake depth ~2-4 m) and Kempenfelt Bay North (lake depth ~6-20 m) areas showed a similar range, but calculated groundwater discharge rates are much higher at the Johnson’s Beach hotspot (3.81 ± 0.91 m$^3$ d$^{-1}$ m$^{-1}$) than the Shingle Bay hotspot (1.60 ± 0.28 m$^3$ d$^{-1}$ m$^{-1}$).
Figure 3.7: Average groundwater discharge (m$^3$ d$^{-1}$ m$^{-1}$) along the Lake Simcoe Shoreline. Areas (boxes) are delineated with groundwater discharge hotspots at Shingle Bay (SB), Johnson’s Beach (JB), Keswick Beach (KB), Duclos Point (DP) and the Thorah Centennial Park (TCP). Calculated groundwater discharge in the Kempenfelt Bay South, Cook’s Bay West, and North Shore area are shown, but as multiple surveys were not conducted in these areas the results are preliminary estimates.

The total nearshore groundwater discharge calculated based on $^{222}$Rn surveys conducted along 190 km of the lake’s shoreline is 197,200 ± 37,800 m$^3$ d$^{-1}$ (Table 3.1). Comparing this to the total volume of tributary inputs to Lake Simcoe for the 2010-2011 year (2,546,200 m$^3$ d$^{-1}$; O’Connor et al., 2013), the total groundwater discharge is approximately 7.6 ± 1.5 % the total volume of tributary inputs. Results of a sensitivity analysis, where the upper and lower quartile groundwater endmember $^{222}$Rn concentrations were used to calculate the total groundwater discharge, show that the total groundwater discharge is
between 5.2 ± 1.0 % and 10.9 ± 2.3 % respectively (additional information shown in Appendix 5). It is also important to note that quantifying groundwater discharge using 222Rn boat surveys only accounts for the nearshore direct groundwater discharge and does not include offshore groundwater discharge which may also be important given the complexity of the geologic environment in the Lake Simcoe area.

Table 3.1: Maximum, minimum and total direct nearshore groundwater discharge along the shoreline for each area of the lake (shoreline areas shown in Figure 3.7).

<table>
<thead>
<tr>
<th>Shoreline Area (Figure 3.7)</th>
<th>Shoreline Length (m)</th>
<th>Maximum (m³d⁻¹m⁻¹)</th>
<th>Minimum (m³d⁻¹m⁻¹)</th>
<th>Total Groundwater Discharge (m³ d⁻¹)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Oro North</td>
<td>24,800</td>
<td>1.60 ± 0.28</td>
<td>0.03 ± 0.03</td>
<td>12,200 ± 4,000</td>
</tr>
<tr>
<td>Kempenfelt Bay North</td>
<td>24,800</td>
<td>3.81 ± 0.91</td>
<td>0.07 ± 0.07</td>
<td>18,700 ± 5,600</td>
</tr>
<tr>
<td>Kempenfelt Bay South</td>
<td>16,500</td>
<td>3.33 ± 0.93</td>
<td>0.75 ± 0.07</td>
<td>20,100 ± 2,700</td>
</tr>
<tr>
<td>Cook’s Bay West</td>
<td>16,300</td>
<td>1.96 ± 0.18</td>
<td>0.16 ± 0.12</td>
<td>11,700 ± 2,800</td>
</tr>
<tr>
<td>Cook’s Bay East</td>
<td>19,400</td>
<td>4.33 ± 0.57</td>
<td>0.57 ± 0.06</td>
<td>26,400 ± 4,100</td>
</tr>
<tr>
<td>Georgina</td>
<td>28,200</td>
<td>4.56 ± 0.47</td>
<td>0.70 ± 0.17</td>
<td>44,400 ± 8,600</td>
</tr>
<tr>
<td>East Shore</td>
<td>17,800</td>
<td>8.08 ± 0.61</td>
<td>0.29 ± 0.09</td>
<td>31,300 ± 3,100</td>
</tr>
<tr>
<td>North Shore</td>
<td>42,300</td>
<td>3.27 ± 0.24</td>
<td>0.08 ± 0.01</td>
<td>32,400 ± 6,900</td>
</tr>
<tr>
<td><strong>Total</strong></td>
<td><strong>190,100</strong></td>
<td></td>
<td></td>
<td><strong>197,200 ± 37,800</strong></td>
</tr>
</tbody>
</table>
While nearshore groundwater discharge accounts for approximately 5-11% of the total tributary input into the lake, pollutant concentrations in groundwater may be higher than in tributaries, and therefore groundwater discharge may account for a higher percentage of the pollutant loading. To illustrate this, the reported P loading from tributaries is compared to an estimate of P loading from groundwater discharge (Table 3.2). Preliminary estimates of P loading from groundwater discharge were calculated using average groundwater concentrations of 0.18 mg L\(^{-1}\) P and 0.40 mg L\(^{-1}\) SRP (soluble reactive phosphorous, readily bioavailable P fraction), from the provincial monitoring well network data (average of six wells around the lake; MOECC Provincial Groundwater Monitoring Network, 2009) and a study of shallow nearshore groundwater chemistry in Kempenfelt Bay (Roy & Malenica, 2013), respectively. O’Connor et al. (2013) estimated that tributary inputs represent the largest source of P to the lake at approximately 67% of the total P inputs. Preliminary calculations suggest that loading from groundwater discharge may represent up to 25% and 57% of the total annual P loading from tributaries based on regional groundwater monitoring well and nearshore groundwater sampling data, respectively (Table 3.2). It is important to note that the high SRP concentrations measured in shallow nearshore groundwater may be due to the decomposition of organic matter that is recirculating across the lake-groundwater interface (Anwar et al., 2014). Therefore these concentrations and associated loading may not be representative of P derived from the aquifer. More broadly, average groundwater SRP loading was estimated using SRP concentrations from 1041 overburden and 1237 bedrock groundwater samples taken across Ontario (Table 3.2; OMNDM Ambient Geochemistry Data, 2014). The average overburden and bedrock groundwater SRP concentrations of 0.01 mg L\(^{-1}\) and 0.003 mg L\(^{-1}\) represent 1.4% and 3.7% of the annual P loading from tributaries respectively.
Table 3.2: Comparison of annual total P and SRP loading from tributaries and estimated from groundwater discharge (t= metric tonne)

<table>
<thead>
<tr>
<th>Sample Type</th>
<th>Groundwater Concentration</th>
<th>Groundwater Loading</th>
<th>Tributary P Loading(^1)</th>
<th>Percent of Tributary Loading (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Nearshore Groundwater Sampling in Kempenfelt Bay area(^2)</td>
<td>0.18 mg L(^{-1}) P</td>
<td>12.4 t P yr(^{-1})</td>
<td></td>
<td>24.9</td>
</tr>
<tr>
<td>Groundwater Monitoring Wells in Lake Simcoe Watershed(^3)</td>
<td>0.40 mg L(^{-1}) SRP</td>
<td>28.4 t SRP yr(^{-1})</td>
<td></td>
<td>57.1</td>
</tr>
<tr>
<td>Overburden Groundwater Samples in Southern Ontario(^4)</td>
<td>0.01 mg L(^{-1}) SRP</td>
<td>0.72 t SRP yr(^{-1})</td>
<td></td>
<td>1.40</td>
</tr>
<tr>
<td>Bedrock Groundwater Samples in Southern Ontario(^4)</td>
<td>0.03 mg L(^{-1}) SRP</td>
<td>2.16 t SRP yr(^{-1})</td>
<td></td>
<td>3.73</td>
</tr>
</tbody>
</table>

\(^1\) O’Connor et al. (2013) and the LSRCA; \(^2\) Groundwater P concentration estimated from Ministry of Environment and Climate Change (MOECC) Provincial Groundwater Monitoring Network (PGMN) samples taken in 2009; \(^3\) Groundwater SRP concentrations estimated from Roy & Malenica (2013); \(^4\) Average PO\(_4\)\(^{3-}\) concentrations in bedrock and overburden estimated from Ontario Ministry of Northern Development and Mines (OMNDM) ambient groundwater geochemistry data set.

The observed spatial variability of groundwater discharge may also cause large differences in P loading to different areas of the lake. For example, two 5 km areas along the lake’s southern shoreline - the first along the base of Duclos Point and the second 20 km west of
Duclos Point - have estimated P loadings of 0.7 t SRP yr\(^{-1}\) and 0.2 t SRP yr\(^{-1}\) respectively, assuming a 0.40 mg L\(^{-1}\) SRP concentration in shallow nearshore groundwater.

**3.4.3 Controls on Direct Groundwater Discharge**

The spatial variability in groundwater discharge was compared with the hydrogeologic conditions around the Lake Simcoe shoreline to identify the main factors controlling the groundwater discharge hotspots. It is resource intensive to conduct a detailed assessment of groundwater inputs into a lake, as was done in this study. To focus available resources in other glaciated lake environments, identifying relationships between the hydrogeologic conditions and groundwater discharge patterns is needed to help identify areas with high groundwater inputs. Factors generally known to affect direct groundwater discharge include regional aquifer sediment type and hydraulic conductivity, recharge to the aquifers connected to the lake, hydraulic gradient between the groundwater and surface water, and nearshore topography (Cherkauer & Hensel, 1986; Feinstein & Reeves, 2010; Kazmierczak et al., 2016; Mulligan & Charette, 2006; Schneider et al., 2005; Tecklenburg & Blume, 2017). In addition, direct groundwater discharge can be lower in areas where there are streams entering the lake as the streams can intercept shallow groundwater travelling towards the lake (see Section 3.4.1.3; Sawyer et al., 2016). Consistent with these factors, the main hydrogeologic features found to be associated with high direct groundwater discharge were: permeable nearshore surficial sediments, proximity to regional recharge features and presence of tunnel channel deposits. Shoreline configuration also influenced direct groundwater discharge hotspots, where higher groundwater discharge was observed in bays/embayments due to convergence of groundwater flow paths (Cherkauer & McKereghan, 1991). Identified groundwater discharge hotspot areas were found to be under the influence of one, or a combination of, these hydrogeologic features as described below.

While deep regional aquifer units can provide a productive source for municipal groundwater supplies, shallow surficial aquifer layers often provide a higher portion of shallow nearshore groundwater discharge to lakes (Grannemann et al., 2000; Kornelsen & Coulibaly, 2014). The identified groundwater discharge hotspots at Johnson’s Beach (JB), Keswick Beach (KB), and Thorah Centennial Park (TCP; Figure 3.8) correspond to areas
with surficial sand and gravel deposits along the shoreline (surficial geology map shown in Figure 3.2). The high direct groundwater discharge in these areas may be associated with shallow nearshore unconfined aquifer units. For example, Johnson’s Beach is located within a continuous sand and gravel corridor associated with the shallowest aquifer units in the Kempenfelt Bay valley (AquaResource Inc., 2013). Direct groundwater discharge is highest where these shallow aquifer units intersect the lakebed in the nearshore area. This is illustrated in the calculated groundwater discharge along the shoreline in Kempenfelt Bay (Figure 3.8a), where a decrease in direct groundwater discharge occurs from east to west, where the surficial sediment changes from gravel to diamicton (lower permeability). A similar pattern was also seen near Keswick Beach (KB; Figure 3.8b) and Thorah Centennial Park (TCP; Figure 3.8c), despite these areas having distinctively different deeper geology. For instance, the subwatershed area where Thorah Centennial Park is located is characterized by flat topography and very thin soil covering karst limestone, while Kempenfelt Bay is characterized by a large valley feature infilled by a series of four glaciolacustrine sand and gravel aquifer units (AquaResource Inc., 2013; Earthfx Inc., 2014a).

There are two large glacial moraines in the Lake Simcoe area that are topographic high points and regional recharge features characterized by unconsolidated glacial sand and gravel deposits: the Oro Moraine and the Oak Ridges Moraine (shown in Figure 3.2). The observed groundwater discharge hotspot at Shingle Bay (SB) is likely influenced by its proximity and hydraulic connection to the Oro Moraine. The steep topography between the moraine and shoreline, and high recharge to the shallow aquifer layers adjacent to the lake likely contribute to the high observed groundwater discharge (Figure 3.9a, c). Similarly, the Oak Ridges Moraine and its associated sediments may influence the groundwater discharge hotspots at Keswick Beach (KB) and Duclos Point (DP). While the Oak Ridges Moraine is not as close to the Lake Simcoe shoreline as the Oro Moraine, the Oak Ridges Moraine sediments may be hydraulically connected to both shallow surficial aquifers and deeper productive channel aquifers along the lake’s southern shoreline (Figure 3.9: Average groundwater discharge (m$^3$ d$^{-1}$ m$^{-1}$) along the shoreline at (a) the Shingle Bay (SB) in the Oro North area, (b) the Keswick Beach (KB) and Duclos Point (DP) in the Cook’s Bay East and Georgina areas. Surficial sand (yellow) and gravel (orange) sediment
associated with the (c) Oro Moraine and (d) Oak Ridges Moraine in these areas is shown (modified from surficial sediment maps by the Ontario Ministry of Northern Development and Mines 2016). Tunnel channel deposits along the southern shoreline may help facilitate this connection between the moraine and the lake with the location of tunnel channel deposits aligning with the groundwater discharge hotspots observed at Keswick Beach (KB) and Duclos Point (DP) (Figure 3.10). These tunnel channel deposits are large erosional channel features, within the regional Newmarket Till aquitard layer, formed by subglacial meltwater flow and infilled by a fining upward sequence of gravel to silt sediment (Earthfx Inc. & Gerber Geosciences, 2008; Sharpe et al., 2004). These channels affect the groundwater flow systems with channel infill sediments often having hydraulic conductivities an order of magnitude higher than the surrounding till material which otherwise acts as a barrier between shallow and deep aquifer layers in the region (Earthfx Inc. & Gerber Geosciences, 2008). This presence of the tunnel channel deposits could effectively recharge the shallow aquifer layers that are in direct contact with the lake, which may explain the high groundwater discharge near KB and DP.
Figure 3.8: Average groundwater discharge (m³ d⁻¹ m⁻¹) along the shoreline near the Johnson’s Beach (JB), Keswick Beach (KB), and Thorah Centennial Park (TCP) hotspots. Areas surrounding hotspots are delineated (boxes) and surficial geology is shown for (a) JB, (b) KB, and (c) TCP modified from Ontario Ministry of Northern Development and Mines (2016).
Figure 3.9: Average groundwater discharge (m$^3$ d$^{-1}$ m$^{-1}$) along the shoreline at (a) the Shingle Bay (SB) in the Oro North area, (b) the Keswick Beach (KB) and Duclos Point (DP) in the Cook’s Bay East and Georgina areas. Surficial sand (yellow) and gravel (orange) sediment associated with the (c) Oro Moraine and (d) Oak Ridges Moraine in these areas is shown (modified from surficial sediment maps by the Ontario Ministry of Northern Development and Mines 2016).
Figure 3.10: Interpreted tunnel channel locations along the southern shoreline of Lake Simcoe, modified from Earthfx Inc. and Gerber Geosciences (2008). Areas of interest are delineated (boxes) and groundwater discharge hotspots at Keswick Beach (KB) and Duclos Point (DP) are marked (red dots).
3.5 Conclusions

Nearshore direct groundwater discharge to Lake Simcoe varies considerably along the shoreline with $^{222}$Rn boat survey data revealing five main groundwater discharge hotspot areas. Hotspots were identified in sheltered bay as well as exposed shoreline areas, and in areas with both shallow and deep nearshore waters. Poor correlation between measured in-lake $^{222}$Rn concentrations and shallow nearshore groundwater $^{222}$Rn concentrations suggests that the spatial variability in the in-lake $^{222}$Rn concentrations is due to higher groundwater inputs and not higher $^{222}$Rn groundwater concentrations. The total direct groundwater discharge to Lake Simcoe was estimated to be 197,200 ± 37,800 m$^3$ d$^{-1}$, which represents approximately 7.6 ± 1.5 % of the total estimated volume of tributary inputs. The contribution of groundwater discharge to pollutant loading however may be considerably higher with preliminary calculations illustrating that groundwater discharge may be an important pathway for P loading to the lake. Groundwater discharge hotspots were related to the presence of one or more of the following features: permeable surficial sediments, tunnel channel deposits, and proximity to large regional recharge features (moraines). These features are associated with high recharge to the adjacent aquifer system as well as providing connectivity between the aquifer system and lake. Groundwater fed streams can also play a role in the in-direct delivery of groundwater to the lake. Although mass balance calculations were not done within 500 m of a tributary inlet, a qualitative assessment of tributary $^{222}$Rn concentrations shows that several of the observed $^{222}$Rn hotspots are likely influenced by a combination of direct and indirect groundwater inputs. Evaluating linkages between groundwater discharge hotspots and regional geologic controls is important to better target future field campaigns investigating groundwater discharge to glaciated lakes. The findings of this study can be broadly applied to other glacial lake environments to understand the potential importance of groundwater discharge and the potential implications for lake water quality.
3.6 References


Earthfx Inc. (2014). *Tier 2 Water Budget, Climate Change, and Ecologically Significant Groundwater Recharge Area Assessment for the Ramara Creeks, Whites Creek and
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Chapter 4

4 Use of the Tracer $^{222}\text{Rn}$ and Regional Scale Groundwater Models to Investigate Groundwater Inputs to a Large Glacial Lake

4.1 Introduction

Groundwater discharge can be an important pathway for delivering pollutants to surface waters, including lakes (Burnett et al., 2006; Kazmierczak et al., 2016; Kornelsen & Coulibaly, 2014; Tecklenburg & Blume, 2017). Pollutant concentrations can be higher in groundwater than receiving surface waters, and as a result groundwater discharge can adversely affect surface water quality (Haack et al., 2005; Howard & Livingstone, 2000; Lewandowski et al., 2015; Roy & Malenica, 2013). However, the impact of groundwater discharge on receiving surface water quality depends on the chemical composition of the discharging groundwater. This is influenced by (i) the land use in the recharge area, where pollutants may contaminate the groundwater at its source, (ii) the groundwater flow path and residence time, and (iii) transformations that take place as groundwater travels along its flow path (Hill, 1990). The groundwater residence time also controls the timing of pollutant loading to surface waters. For example, Howard & Livingstone (2000) showed that 80% of conservative pollutants (e.g. chloride) released into shallow nearshore aquifers in the Toronto area over the past several decades will discharge to Lake Ontario over the next 100 years, thus representing a legacy water quality issue.

Groundwater discharge to lakes is often a poorly quantified and an overlooked component of water and chemical budgets. This is in part due to the complexity of quantifying groundwater discharge, due to its high spatial and temporal variability, and limited techniques available to characterize this variability at a regional scale (Burnett & Dulaiiova, 2006; McBride & Pfannkuch, 1975; Mulligan & Charette, 2006; Schneider et al., 2005). While numerous studies have shown how land use activities can degrade groundwater quality, the relationship between land use activities, subsurface flow paths, and potential implications for lake water quality is challenging to evaluate (Kornelsen & Coulibaly, 2014; Robinson, 2015). Groundwater can enter a lake either directly from an aquifer (direct
groundwater discharge), or indirectly through groundwater-fed streams that flow into the lake (indirect groundwater discharge). This study focuses on direct groundwater discharge. Spatial variability of direct groundwater discharge is influenced by the hydraulic gradient between the groundwater level and lake water level, distribution and volume of recharge to aquifers hydraulically connected to the lake, heterogeneities in nearshore hydrogeology and corresponding hydraulic conductivity, and the presence of embayments along the lake shoreline (Cherkauer & Nader, 1989; Feinstein & Reeves, 2010; Meinikmann et al., 2013; Schneider et al., 2005). Methods for quantifying groundwater discharge at the regional scale include geochemical/isotopic tracer methods, numerical groundwater models and water budget calculations (Dimova et al., 2013; Kidmose et al., 2015; Lambert & Burnett, 2003; Smith & Zawadzki, 2003). Radon-222 (222Rn), a naturally occurring radium isotope, has been widely used as a tracer to quantify groundwater discharge to surface waters, including lakes (Burnett et al., 2006; Burnett & Dulaiova, 2006; Corbett et al., 1997; Dimova & Burnett, 2011; Dulaiova et al., 2010; Ono et al., 2013). A number of studies have compared 222Rn results to other groundwater discharge measurement techniques including other geochemical tracers for groundwater (e.g., methane and conductivity), local scale physical techniques (e.g., seepage meters, hydraulic gradient measurements), and larger scale water budget estimates. The comparison studies demonstrate 222Rn is a suitable tracer for regional scale characterization of groundwater discharge to lakes (Burnett & Dulaiova, 2006; Corbett et al., 2000; Ji et al., 2017).

Although studies have shown the suitability of 222Rn as a tracer for quantifying regional scale groundwater discharge, little attention is given to the groundwater history (i.e., groundwater flow paths and residence time, recharge areas) and potential implications for pollutant fluxes to the lake, and thus lake water quality management. To understand the underlying regional scale groundwater flow systems, for which flow and transport processes generally occur on a temporal scale that cannot be directly captured by field measurements, regional scale numerical groundwater models can be used (Kornelsen & Coulibaly, 2014; Marchildon et al., 2016). Integrated groundwater-surface water modelling has been used extensively to characterize regional scale groundwater flow systems and inform management decisions related to, for example, drinking water protection zones, climate change impacts, land use management changes, and well pumping (Tanvir Hassan
et al., 2014; Huntington & Niswonger, 2012; Niswonger et al., 2014; Woolfenden & Nishikawa, 2014). In Ontario, Canada, a large number of regional scale groundwater models have been developed for drinking water source protection, water resource management, and water quality protection initiatives (e.g. Holysh & Gerber, 2014; Marchildon et al., 2016; Sharpe et al., 2004). Groundwater flow models are also often used to simulate contaminant transport with some being applied to evaluate the potential discharge of groundwater contaminants to receiving surface waters (Boutt et al., 2001; Howard & Livingstone, 2000; Kidmose et al., 2015). To evaluate specific groundwater flow paths that connect areas of recharge and discharge, forward or backward particle tracking analysis can also be conducted (Batelaan et al., 2003; Marchildon et al., 2016; Matula et al., 2014; Modica et al., 1998; Rock & Kupfersberger, 2002).

Comparing $^{222}$Rn-derived estimates and model simulated groundwater discharge to a lake can provide insight into the strengths and limitations of both methods, as well as improve confidence in the groundwater discharge estimates (Burnett et al., 2006). It is also recommended that future research on groundwater-surface water interactions in the Laurentian Great Lakes Basin, and in any basin of interest, should use both regional scale groundwater modelling and field techniques (Kornelsen & Coulibaly, 2014). To our knowledge, the only studies to compare $^{222}$Rn-derived and model simulated groundwater discharge estimates were Lambert & Burnett (2003) and Smith & Zawadzki (2003) who conducted a comparison experiment to quantify groundwater discharge along the Gulf of Mexico coast, Florida. To our knowledge, no studies have compared these two methods of quantifying regional scale groundwater discharge to a lake, and more importantly, applied regional scale groundwater models to evaluate the history of the discharging groundwater.

Lake Simcoe, a large inland lake (722 km$^2$) in southern Ontario, provides an ideal large lake system in which to conduct a comparison of $^{222}$Rn-derived and model simulated direct groundwater discharge estimates. Over the past several decades the Lake Simcoe watershed has experienced rapid population growth, and an increase in nutrient and chloride loading to the lake has degraded the lake water quality and affected fish populations. The Lake Simcoe Protection Plan was approved by the government of Ontario in 2009, with the goal of protecting and maintaining ecological health, and restoring self-
sustaining cold-water fish communities in the Lake Simcoe watershed (Marchildon et al., 2016; MOECC, 2009). Lake water quality management efforts have mainly focused on quantifying and controlling point source pollutants and tributary inputs (Palmer et al., 2011), and the relative importance of direct groundwater discharge as a pathway for delivering pollutants to Lake Simcoe remains unclear. Under the Lake Simcoe Protection Plan (MOECC, 2009) integrated groundwater and surface water models were developed for subwatershed areas within the Lake Simcoe Basin to assess ecologically significant recharge areas (ESGRAs) and complete water budget and water quality stress assessments. While the objective of these models was not to quantify the amount and spatial distribution of groundwater discharge to the lake, it may be possible to apply the models for this purpose, provided they incorporate sufficiently characterized recharge and hydrogeologic information. Lake Simcoe is also an ideal setting for this study as it provides an opportunity to investigate the effects of large glacial features including moraines and erosional channels on the regional groundwater flow systems using field and groundwater modelling methods.

The objectives of this study are (i) to compare $^{222}$Rn-derived and model simulated estimates of groundwater discharge to Lake Simcoe for two areas along the shoreline, and (ii) to evaluate the potential impact of groundwater discharge on lake water quality by characterizing groundwater discharge pathways and associated recharge areas. In this Chapter, field and model estimates of direct groundwater discharge are first compared along the north-western and southern shorelines of Lake Simcoe. Groundwater flow paths, recharge areas and the potential implications for lake water quality are then evaluated by conducting back particle tracking along the north-western shoreline. This study demonstrates the value of comparing regional scale field data and model simulation results to provide comprehensive understanding of groundwater discharge and inform lake water quality management initiatives.

### 4.2 Field Site

This field site for this study is Lake Simcoe, a large glacial lake in Southern Ontario (Figure 4.1). Lake Simcoe has a watershed area of 2900 km$^2$, and approximately 240 km of shoreline (Eimers et al., 2005; Palmer et al., 2011). Land use in the watershed is largely agricultural (approximately 47%), however, the population has grown significantly -
doubling between 1981 and 2005 - leading to a rapid urbanization of otherwise undeveloped areas around the lake (Eimers et al., 2005; Palmer et al., 2011). In addition, the southern shoreline of Lake Simcoe represents the northern most extent of the Greater Toronto Area, Canada’s largest urban center (North et al., 2013; Palmer et al., 2011). This changing land use in the watershed has caused a deterioration of lake water quality, and declines in fish populations (Eimers et al., 2005). Elevated phosphorus loading to the lake is of particular concern. Increases in phosphorous loading have stimulated excessive algal growth and depleted dissolved oxygen concentrations, which has affected cold-water fish populations (North et al., 2013; O’Connor et al., 2013). Road salting in urban areas has also caused increased chloride loading in the watershed (O’Connor et al., 2006).

The Lake Simcoe watershed represents a complex geologic and hydrogeologic environment that includes a variety of surficial sediment types, aquifer systems, topography, and depositional/erosional features. Within the watershed, there are two large glacial moraine features: the Oro Moraine and the Oak Ridges Moraine (shown in Figure 4.1a). These moraines are regional recharge features and topographic high points in the north-western and southern shoreline areas (AquaResource Inc., 2013; Earthfx Inc., 2014a, 2008; Genivar Inc., 2013). The two focus areas for this study are the Oro-Hawkestone area, and the York Region area which include the Oro Moraine and Oak Ridges Moraine, respectively (Figure 4.1).

### 4.2.1 Oro- Hawkestone Area

The Oro-Hawkestone study area is located along the north-western shoreline of Lake Simcoe. The model area and field surveyed area (shown in Figure 4.1a,b) represents approximately 43 km of the lake’s shoreline and encompasses three subwatersheds in the Lake Simcoe Basin: Oro Creeks North, Oro Creeks South, and Hawkestone Creek. This area is largely agricultural, with the city of Orillia which is the largest urban center located in the northern region. Land surface topography in the area is characterized by a topographic high point at the Oro Moraine (405 masl) and a downward slope to the Lake Simcoe shoreline (219 masl) (Earthfx Inc., 2013b). The Oro Moraine is an east to west trending glacial moraine, whose thick sand and gravel deposits act as recharge features for regional aquifer layers (Marchildon et al., 2016). Surficial sediments in the lowland areas,
near the lake shoreline, are predominately diamicton and lacustrine sand deposits, with some clay plains in the north (shown in Figure 4.1a; Earthfx Inc., 2013a; Marchildon et al., 2016). Drift thickness along the shoreline, which contains a series of aquifer and aquitard layers, is 50-100 m in the southern area thinning to less than 15 m in the north (Burt & Dodge, 2011; Marchildon et al., 2016).

4.2.2 York Region Area

The York Region study area is located along the southern shoreline of Lake Simcoe. The extents for the York Region model are bordered by Lake Simcoe in the north, and Lake Ontario in the south, with the peak of the Oak Ridges Moraine (390masl) as a surface water divide in the center (Earthfx Inc., 2014b). The delineated model and field-surveyed area, shown in Figure 4.1a, c only represents the northern portion of the total model extent. The study area includes approximately 48 km of the Lake Simcoe shoreline, and includes four subwatersheds: East Holland, Maskinonge River, Black River, and Georgina Creeks. Land use in these subwatersheds is predominately agricultural, with developed areas including urban areas and roads covering approximately 28% of the subwatershed (Earthfx Inc., 2013c). The Oak Ridges Moraine is the main recharge feature with thick sandy deposits and hummocky terrain (Earthfx Inc., 2014b). Nearshore surficial sediment is characterized by sand and diamicton, with some localized clay and organic deposits (Figure 4.1a). The groundwater flow system in the York Region area is influenced by large erosional channels which often have hydraulic conductivities an order of magnitude higher than adjacent geologic layers (Earthfx Inc. & Gerber Geosciences, 2008). Although the channels are often quite large, and in some areas may be deeper than the lake in the nearshore area, the channel sediment may facilitate a hydraulic connection between shallow and deep groundwater flow systems (Earthfx Inc., 2014b).
Figure 4.1: (a) Surficial geology around Lake Simcoe, Ontario, Canada (modified from Ontario Ministry of Northern Development and Mines (2016)). Solid black lines represent shoreline areas where $^{222}$Rn surveys were performed, and boxes indicate the model extents for the (b) Oro-Hawkestone groundwater model and (c) York Region groundwater model.

4.3 Methods

4.3.1 $^{222}$Rn Field Measurements and Mass Balance Model

$^{222}$Rn boat surveys were conducted along approximately 35 km of shoreline in the Oro-Hawkestone area, and 48 km of shoreline in the York Region area from June 2015- July 2018 to quantify direct nearshore groundwater discharge to Lake Simcoe (Figure 4.1). $^{222}$Rn is a suitable tracer for groundwater discharge as it occurs naturally in a variety of aquifer materials, including glacial sediment, and concentrations are often 3-4 orders of magnitude higher in groundwater than receiving surface waters (Burnett et al., 2006; Burnett et al., 2001; Je & Eyles, 1998; Schmidt et al., 2010). Measured in-lake $^{222}$Rn concentrations can be used to estimate nearshore groundwater discharge rates along the
shoreline by applying a mass balance model (Burnett & Dulaiova, 2003; Cable et al., 1996; Corbett et al., 1997; Dulaiova et al., 2010; Schmidt & Schubert, 2007).

Details of the $^{222}\text{Rn}$ survey approach and mass balance calculations are provided in Chapter 3 with a brief summary provided here. Continuous $^{222}\text{Rn}$ measurements were taken at an offshore distance of 50-200 m, from a boat travelling approximately 3-5 km h$^{-1}$. Lake water was continuously pumped, via a submersible pump (Rule 3700GPH Bilge Pump), into a closed loop system. This system consisted of an air-water exchanger (RAD Aqua; Durridge Co.), which allowed $^{222}\text{Rn}$ concentrations between air and incoming lake water to reach equilibrium, and multiple RAD7 (Durridge Co.) detectors to measure the $^{222}\text{Rn}$ concentrations in the air (Burnett et al., 2001; Dulaiova et al., 2005; Lane-Smith et al., 2002). Given the relatively low in-lake $^{222}\text{Rn}$ concentrations, multiple RAD7 units were connected in parallel to minimize measurement error while still maintaining sensitivity to in-lake concentration changes (Dimova et al., 2013; Dulaiova et al., 2005; see Appendix 2 for details on equipment response time and measurement uncertainty). $^{222}\text{Rn}$ measurements were taken over a 15-minute integration cycle, and survey locations were recorded using a handheld GPS unit (Trimble Geo5T handheld, Trimble). Measured $^{222}\text{Rn}$ concentrations were spatially referenced by applying the concentration measured during the 15 minute cycle to the alongshore distance travelled during the cycle. The error reported alongside each $^{222}\text{Rn}$ measurement is the standard deviation ($\sigma$) following Poisson statistics (Taylor, 1982; Durridge Co.). Survey areas were surveyed multiple times to ensure repeatable and representative results.

Groundwater discharge rates were estimated using the $^{222}\text{Rn}$ data by applying a steady state mass balance model. The model considers the sources and losses of $^{222}\text{Rn}$ in a well-mixed surface water volume; in this case the representative lake volume for each 15 minute survey cycle (Burnett et al., 2001; Corbett et al., 1997; Dulaiova et al., 2010; Schmidt et al., 2010). Following Dulaiova et al. (2010) the groundwater discharge rate ($Q$, m$^3$d$^{-1}$) along the shoreline was calculated by:

$$Q = \frac{C_{sw}V}{C_{gw}t}$$  \hspace{1cm} (4-1)
where $C_{sw}$(dpm m$^{-3}$) is the in-lake $^{222}$Rn concentration corrected for sources and losses, $C_{gw}$ (dpm m$^{-3}$) is the groundwater endmember $^{222}$Rn concentration, $V$ (m$^3$) is the surface water volume, and $\tau$ (d) is the flushing rate for that volume. The flushing rate was assumed to be equal to 5.53 d, the mean life of the $^{222}$Rn isotope (Burnett et al., 2013). Shallow groundwater samples were taken at 20 beach sites and sediment samples (for sediment equilibration experiments; Chanyotha et al., 2014) were taken at 13 beach sites around the lake to determine a representative value for $C_{gw}$ (Burnett et al., 2007; Dimova et al., 2013). Additional details on groundwater and sediment sampling and analysis methods are provided in Sections 3.3.1.2 and 3.4.1.2 with sample results provided in Appendix 5. A representative groundwater endmember $^{222}$Rn concentration of 234.8 ± 14.0 dpm L$^{-1}$ was used in the mass balance calculations.

In-lake $^{222}$Rn concentrations ($C_{sw}$, dpm m$^{-3}$) were corrected by considering the following loss terms:

1. Evasion of $^{222}$Rn to the atmosphere, $J_{atm}$, dpm m$^{-2}$d$^{-1}$ (Burnett & Dulaiova, 2003; Dulaiova et al., 2010; Macintyre et al., 1995):

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

(4-2)

where $C_{air}$ (dpm m$^{-3}$) is the ambient air $^{222}$Rn concentration (set to 500 dpm m$^{-3}$ based on field measurements), and $\alpha$ (dimensionless) is the Ostwald’s solubility coefficient. The gas transfer coefficient, $k$ (m d$^{-1}$), was calculated by (Macintyre et al., 1995):

$$k(600) = 0.45 \times u_{10}^{1.6} \times \left(\frac{S_c}{600}\right)^{-b}$$

(4-3)

where $u_{10}$ (m s$^{-1}$) is the wind speed which was taken from the nearest Environment Climate Change Canada (ECCC) weather station (located 5 - 28 km away from a given survey site), $b$ is 0.5 for wind speeds greater than 3.6 m s$^{-1}$ and 0.667 for wind speeds less than 3.6 m s$^{-1}$ (Baskaran, 2016; Dimova et al., 2013; Macintyre et al., 1995), and $S_c$ is the Schmidt number for $^{222}$Rn (1000; Baumert et al., 2005). The Oswald solubility coefficient, $\alpha$, is dependent on the temperature at the air-
water interface ($T$, °C) and was calculated by (Dimova et al., 2013; Macintyre et al., 1995):

$$\alpha = 0.105 + 0.405 \exp(-0.05027T)$$  (4-4)

2. Decay of $^{222}$Rn in the surface water volume, $J_{\text{decay}}$, dpm m$^2$d$^{-1}$ (Schmidt & Schubert, 2007):

$$J_{\text{decay}} = z\lambda_{Rn}C_{sw}$$  (4-5)

where $\lambda_{Rn}$ (d$^{-1}$) is the $^{222}$Rn decay rate which is 0.181 d$^{-1}$.

3. The horizontal offshore mixing loss, $J_{\text{mix}}$, dpm m$^2$d$^{-1}$, was estimated using an iterative method described in Santos et al. (2008). Shore-perpendicular $^{222}$Rn and conductivity shore-perpendicular transect data was collected in three different locations of Lake Simcoe to evaluate the offshore concentration gradient at each location (Appendix 7). Concentration gradients were used in the steady state advection diffusion equation to estimate the horizontal eddy diffusion coefficient $K_h$ (m$^2$d$^{-1}$):

$$K_h \frac{\partial^2C}{\partial x^2} - \omega \frac{\partial C}{\partial x} - \lambda C = 0$$  (4-6)

where $\omega$ (m d$^{-1}$) is the horizontal surface advection, $C$ is the concentration of the tracer ($^{222}$Rn or conductivity), and $\lambda$ is the $^{222}$Rn decay constant. To calculate $K_h$, measured in-lake $^{222}$Rn concentrations had to be corrected for losses ($J_{\text{decay}}, J_{\text{atm}}$) over the time taken for $^{222}$Rn to travel to from the source to the measurement location. An iterative approach was used to perform this correction, whereby the $^{222}$Rn concentrations were corrected and in turn used in equation 4-6 to estimate a new $K_h$ and $\omega$. Iterations continued until both values converged. The final $K_h$ value was then used to calculate $J_{\text{mix}}$ (Burnett et al., 2008; Moore, 2000; Santos et al., 2008):

$$J_{\text{mix}} = -K_h \left(\frac{C_{S+1} - C_{S-1}}{2\Delta x}\right) x \frac{A_{CS}}{A_{LB}}$$  (4-7)
where \( \frac{C_{S+1} - C_{S-1}}{2 \Delta x} \) (dpm m\(^{-3}\) m\(^{-1}\)) is the measured offshore \(^{222}\)Rn gradient, and \( \frac{A_{CS}}{A_{LB}} \) is the ratio of the along-shore cross sectional area (\( A_{CS} \), m\(^2\)) to the lake bed area (\( A_{LB} \), m\(^2\)) for the surface water volume represented by the cycle measurement.

Mass balance calculations were done for all \(^{222}\)Rn survey cycles to calculate the groundwater discharge rate along the shoreline (\( Q \), m\(^3\) d\(^{-1}\)). Values were divided by the shoreline length traveled during the survey cycle to obtain discharge rates per unit of shoreline (m\(^3\) d\(^{-1}\) m\(^{-1}\)).

Weather conditions, particularly storms causing high wind, precipitation, and waves, can affect measured in-lake \(^{222}\)Rn concentrations such that they cannot be adequately corrected using the \( J_{atm} \) and \( J_{mix} \) terms (Burnett et al., 2007). To address this, an 8 day time series test was performed at a location on the southern shoreline of Lake Simcoe to evaluate the response of in-lake \(^{222}\)Rn concentrations to inclement weather conditions. Test results showed that in-lake \(^{222}\)Rn concentrations were rapidly depleted following prolonged high onshore wind speeds (greater than 20 km h\(^{-1}\) for 6-12 hours) and a 10 mm precipitation event, and remained depleted for over 12 hours (see Appendix 4 for additional details). Based on these results, \(^{222}\)Rn surveys conducted within 12 hours of high wind and precipitation events were discarded. The average in-lake \(^{222}\)Rn concentrations, and the corresponding average groundwater discharge values along the shoreline for non-discarded survey days were used for comparison with the groundwater model results.

4.3.2 Integrated groundwater-surface water subwatershed models

Integrated groundwater-surface water models developed for the Oro-Hawkestone area and York Region were applied to simulate direct groundwater discharge to the lake. The Oro-Hawkestone model was originally developed and calibrated to (i) perform a Tier 2 Water Budget Analysis and Stress Assessment and (ii) identify Ecologically Significant Groundwater Recharge Areas (ESGRAs) as mandated in the Lake Simcoe Protection Plan (LSPP) (Earthfx Inc., 2013b). The York Region model was originally developed and calibrated for a risk assessment of the municipal water supply in the York Region presented as a Tier 3 Water Budget and Local Area Risk Assessment report under the Clean Water Act (MOECC, 2006; Earthfx Inc., 2013c).
Both integrated groundwater-surface water models were developed in GSFLOW (Earthfx Inc., 2013b, 2013c; Huntington & Niswonger, 2012; Markstrom et al., 2008; Woolfenden & Nishikawa, 2014). GSFLOW combines PRMS to simulate recharge, with MODFLOW to simulate groundwater flow (Harbough et al., 1996; Leavesley et al., 1983; Markstrom et al., 2008). Precipitation, climate conditions, soil type, land surface topography and land use are incorporated into the PRMS model to simulate the groundwater recharge, runoff and stream flow (Leavesley et al., 1983). The MODFLOW groundwater flow model is based on hydrogeological conceptualization of the regional aquifer-aquitard systems with corresponding properties assigned (i.e. storage coefficient, hydraulic conductivity). The groundwater flow model enables simulation of groundwater levels in the model domain, exchange of groundwater between shallow and deep aquifer layers, and rates of groundwater discharge across model boundaries including surface water features, over the simulated time period (Earthfx Inc., 2013b; Harbough et al., 1996).

Data for model input and calibration for both models includes long term climate data from ECCC weather stations, SOLRIS (southern Ontario land use resource information system) land use data, and Water Survey of Canada stream gauge data (Earthfx Inc., 2013b, 2013c). The York Region model domain (~2700 km²) incorporates a much larger and more complex area than the Oro-Hawkestone model (~400 km²) (Earthfx Inc., 2013c). In the Oro-Hawkestone area, model hydrostratigraphy was based on a comprehensive three-dimensional geologic model compiled by the Ontario Geologic Survey (OGS, 2011), and detailed geologic and groundwater flow modelling work done for the Oro Moraine aquifer system (Beckers & Frind, 2000; Burt & Dodge, 2011; Earthfx Inc., 2013b). The geology in the York Region model area is not as well characterized, and model hydrostratigraphy was defined using a compilation of multiple sources including previous modelling work for subset areas within the larger model area and MOECC borehole logs (Earthfx Inc., 2013c). Groundwater levels, which served as the primary calibration target, were determined from provincial, municipal and private well data across the model areas (Earthfx Inc., 2013b, 2013c). Additional details of model development and calibration for the Oro-Hawkestone and York Region models can be found in Earthfx Inc., 2013b and Earthfx Inc. 2013c, respectively. Boundary conditions for the models include constant head conditions along the lake boundary and no flow conditions along the remaining lateral
boundaries (Earthfx Inc., 2013b, 2013c). In both the Oro-Hawkestone and York Region models, the Lake Simcoe shoreline was set to be a constant head boundary (220 masl), and therefore, one of the model outputs for the steady state simulation was leakage through this boundary. The simulated leakage (m$^3$ d$^{-1}$) for cells along the shoreline were divided by the 100 m cell size to obtain groundwater discharge rates per unit of shoreline (m$^3$ d$^{-1}$ m$^{-1}$). These values were then compared with direct groundwater discharge rates estimated from the $^{222}$Rn mass balance calculations (m$^3$ d$^{-1}$ m$^{-1}$).

Backward particle tracking was performed with particles placed along the Lake Simcoe shoreline to determine the origins, flow paths, and travel times of groundwater discharging to the lake (Buxton et al., 1991; Earthfx Inc., 2013a; Rock & Kupfersberger, 2002). The Oro-Hawkestone model was chosen for particle tracking analysis because of the well-defined hydrogeology incorporated into the model, and the range of conditions along the shoreline that may affect the groundwater flow paths (i.e. variable surficial geology, drift thickness, lake bathymetry, and proximity to the Oro Moraine). Particle tracking was completed using MODPATH, which uses the velocity flow field simulated by the MODFLOW groundwater flow model.

4.4 Results and Discussion

4.4.1 Comparison of $^{222}$Rn-Derived and Model Simulated Groundwater Discharge Estimates

Measured in-lake $^{222}$Rn concentrations, $^{222}$Rn-derived groundwater discharge, and simulated groundwater discharge to Lake Simcoe in the Oro-Hawkestone and York Region areas are shown in Figure 4.2 and Figure 4.4, respectively. In the Oro-Hawkestone area, $^{222}$Rn boat surveys and corresponding mass balance calculations were performed for 35 km (~80%) of the 43 km of shoreline length simulated in the model. In the York Region area, all data spans the complete 48 km shoreline length. The $^{222}$Rn concentrations and calculated groundwater discharge values shown in Figure 4.2b, c and Figure 4.4b, c represent the average of all surveys not performed within 12 hours of high sustained wind speeds and precipitation. Data from all $^{222}$Rn surveys, along with weather conditions during each survey, are provided in Appendix 6.
4.4.1.1 Oro-Hawkestone Area

The in-lake $^{222}\text{Rn}$ concentrations vary considerably along the shoreline, ranging from 0.13 ± 0.13 dpm L$^{-1}$ to 2.26 ± 0.53 dpm L$^{-1}$ (Figure 4.2b). Groundwater discharge values calculated using the $^{222}\text{Rn}$ data also show considerable spatial variability, ranging from 0.07 ± 0.07 m$^3$ d$^{-1}$ m$^{-1}$ to 2.45 ± 0.32 m$^3$ d$^{-1}$ m$^{-1}$ (Figure 4.2c). While the highest $^{222}\text{Rn}$ concentrations were generally measured along the northern shoreline (~0 - 24 km), the groundwater discharge was calculated to be highest along the southern shoreline (~33 - 43 km). This is because the mass balance calculations consider the $^{222}\text{Rn}$ inventory within a given water column depth and the nearshore lake depth is much greater in the south (6 – 20 m) compared to the north (2 – 4 m; Ontario Ministry of Natural Resources, 2006). The model simulated groundwater discharge values compare well with the $^{222}\text{Rn}$-derived groundwater discharge along the shoreline, ranging from 0 m$^3$ d$^{-1}$ m$^{-1}$ to 3.10 m$^3$ d$^{-1}$ m$^{-1}$, with a RMSE of 0.61 m$^3$ d$^{-1}$ m$^{-1}$ (Figure 4.2b, c). The total direct groundwater discharge calculated using the $^{222}\text{Rn}$ survey data along 35 km of the Oro-Hawkestone shoreline is 20,800 ± 5700 m$^3$ d$^{-1}$. This compares well with the simulated total direct groundwater discharge along this shoreline length (20,300 m$^3$ d$^{-1}$). The model results show that groundwater discharge to streams and wetlands in the Oro-Hawkestone areas is considerably larger than direct groundwater discharge (83,200 m$^3$ d$^{-1}$), indicating that indirect groundwater inputs are also important in in the Oro-Hawkestone area (Earthfx Inc., 2013b).

Three areas with consistently elevated $^{222}\text{Rn}$ concentrations and calculated groundwater discharge relative to the remainder of the shoreline were identified at Shingle Bay (SB; 4 - 9 km), Carthew Bay (CB; 19 - 24 km), and Shanty Bay (STB; 33 - 42 km; locations shown in Figure 4.2). The spatial distribution of $^{222}\text{Rn}$-derived and simulated groundwater discharge is in good agreement along the shoreline length, with values of similar magnitude in the three areas with high discharge (Figure 4.2c). The highest $^{222}\text{Rn}$-derived and simulated groundwater discharge values are in Shanty Bay (RMSE= 0.78 m$^3$ d$^{-1}$ m$^{-1}$; max = 2.45 ± 0.32 m$^3$ d$^{-1}$ m$^{-1}$ and 3.10 m$^3$ d$^{-1}$ m$^{-1}$, respectively). The simulated groundwater discharge in the Shanty Bay area (33 – 42 km), however, is more variable along the shoreline compared to the $^{222}\text{Rn}$-derived discharge estimates which show a more defined
peak area of high discharge (36 km, Figure 4.2c). The simulated groundwater discharge was similar to the $^{222}$Rn-derived discharge at Carthew Bay (RMSE= 0.55 m$^3$ d$^{-1}$ m$^{-1}$; max = 1.18 m$^3$ d$^{-1}$ m$^{-1}$ and 1.24 ± 0.13 m$^3$ d$^{-1}$ m$^{-1}$, respectively) but lower at Shingle Bay (RMSE= 0.93 m$^3$ d$^{-1}$ m$^{-1}$; max = 0.56 m$^3$ d$^{-1}$ m$^{-1}$ and 1.60 ± 0.30 m$^3$ d$^{-1}$ m$^{-1}$, respectively).

Analysis of the contribution of each geologic layer to the total simulated groundwater discharge show that the geologic layers that contribute to the groundwater discharge vary along the shoreline. The simulated groundwater discharge from each model layer is shown in Figure 4.2d, with cross sections in the northern, central, and southern shoreline areas shown in Figure 4.3. The model geologic layers 5, 6, and 7 represent regional lower drift (upper), lower drift (lower), and basal aquifer units respectively. The lower drift units are glaciolacustrine sand/ silty-sand aquifers, while the basal aquifer unit is comprised of weathered carbonate bedrock and gravel. Along the northern shoreline including the Shingle Bay (SB) area, the simulated groundwater discharge is predominately from aquifer layers 5 (26%), 6 (17%), and 7 (26%), with the remaining 31% contributed from layers 1-4. In the Carthew Bay (CB) area the majority of groundwater discharge is contributed from layer 5 (38%) and layer 6 (47%), with a small contribution from layers 1-4 (11%) and layer 7 (4%). Along the southern shoreline, including the Shanty Bay (STB) area, the groundwater discharge comes from a combination of layers 1-4 (42%), and layer 7 (45%), with lower contribution (13%) from layers 5 and 6. The depth of these geologic layers relative to the lake surface vary along the shoreline. For example, at the lake model boundary intermediate layers 5 and 6 are closer to the lake surface (< 10m) in the Shingle Bay area (Figure 4.3 section A-A’), but are much deeper (> 20m) in the Shanty Bay area (Figure 4.3 section C-C’). The potential implications of this on the groundwater discharge estimates are discussed in Section 4.4.1.3.
Figure 4.2: Comparison results for (a) the Oro-Hawkestone model area, scale is shown. Data from (b) $^{222}$Rn boat surveys and (c) $^{222}$Rn mass balance calculations are compared with the simulated groundwater discharge along the shoreline. The simulated discharge from each geologic model layer is shown in (d). Three groundwater discharge hotspots, shown with red dashed boxes, were identified at Shingle Bay (SB), Carthew Bay (CB) and Shanty Bay (STB).
Figure 4.3: Cross section view of the Oro-Hawkestone model geologic layers in the northern (A-A’), central (B-B’), and southern (C-C’) shoreline areas within the model domain, shown on map (right). High and low permeability geologic layers (i.e. aquifer and aquitard) are shown in yellow and blue, respectively. Lake level is shown at 219 masl (Earthfx Inc. (2013 b, c)).

4.4.1.2 York Region Area

Measured in-lake $^{222}$Rn concentrations along the shoreline in the York Region area ranged from 0.27 $\pm$ 0.25 dpm L$^{-1}$ to 3.69 $\pm$ 0.88 dpm L$^{-1}$ (Figure 4.4b). $^{222}$Rn-derived groundwater discharge values show a similar spatial variability along the shoreline, ranging from 0.57 $\pm$ 0.06 m$^3$ d$^{-1}$ m$^{-1}$ to 4.56 $\pm$ 0.47 m$^3$ d$^{-1}$ m$^{-1}$ (Figure 4.4c). The in-lake $^{222}$Rn concentrations and calculated groundwater discharge along the shoreline show a similar pattern because, in contrast to the Oro-Hawkestone area, the nearshore lake depth is relatively constant along the shoreline length (varies from 2 - 8m; Ontario Ministry of Natural Resources, 2006). The simulated groundwater discharge shows greater variability along the shoreline compared to the $^{222}$Rn-derived estimates ranging from from 0.00 m$^3$ d$^{-1}$ m$^{-1}$ to 10.22 m$^3$ d$^{-1}$ m$^{-1}$ (Figure 4.4b, c). The overall comparison yields a RMSE of 3.40 m$^3$ d$^{-1}$ m$^{-1}$. The total $^{222}$Rn-derived direct groundwater discharge along the 48 km shoreline in the York Region area is 72,400 $\pm$ 12,900 m$^3$ d$^{-1}$, which is slightly higher than the total simulated direct
groundwater discharge \((62,100 \text{ m}^3 \text{ d}^{-1})\). While the model predicts higher groundwater discharge in the western shoreline area \((0 – 17 \text{ km})\) compared to the \(^{222}\text{Rn}\)-derived estimates, the simulated groundwater discharge is lower along the central and eastern shoreline \((18 – 48 \text{ km}; \text{Figure 4.4c})\). Similar to the Oro-Hawkestone model, groundwater discharge to streams and wetlands \((335,800 \text{ m}^3 \text{ d}^{-1})\) in York Region is estimated to be considerably larger than direct groundwater discharge, suggesting indirect groundwater inputs may also be important in this region.

Along the York Region shoreline, there are three areas with consistently elevated \(^{222}\text{Rn}\) concentrations and \(^{222}\text{Rn}\)-derived groundwater discharge, relative to the adjacent shoreline area (shown in Figure 4.4). These are at Keswick Beach (KB; 4 – 13 km), Willow Beach (WB; 23 – 27 km) and Duclos Point (DP; 32 – 48 km). The simulated groundwater discharge agrees does not agree very well with \(^{222}\text{Rn}\)-derived discharge values in the Keswick Beach (RMSE= 4.57 m\(^3\) d\(^{-1}\) m\(^{-1}\)), Willow Beach (RMSE= 2.31 m\(^3\) d\(^{-1}\) m\(^{-1}\)) or Duclos Point (RMSE= 2.60 m\(^3\) d\(^{-1}\) m\(^{-1}\)). While the simulated groundwater discharge values are more variable along the western shoreline compared to the \(^{222}\text{Rn}\)-derived estimates, the discharge values are similar at KB (~8 km) where the highest \(^{222}\text{Rn}\)-derived groundwater discharge was calculated \((4.44 \text{ m}^3 \text{ d}^{-1} \text{ m}^{-1} \text{ and } 4.33 \pm 0.57 \text{ m}^3 \text{ d}^{-1} \text{ m}^{-1} \) for simulated and \(^{222}\text{Rn}\)-derived values, respectively). Peak groundwater discharge values are also similar in the Willow Beach area, with simulated and \(^{222}\text{Rn}\)-derived estimates of 2.93 m\(^3\) d\(^{-1}\) m\(^{-1}\) and 2.48 \pm 0.29 m\(^3\) d\(^{-1}\) m\(^{-1}\) respectively. At Duclos Point, although the \(^{222}\text{Rn}\)-derived groundwater discharge shows several distinct peak discharge values \((\text{max} = 4.55 \pm 0.47 \text{ m}^3 \text{ d}^{-1} \text{ m}^{-1})\), the simulated groundwater discharge is relatively low compared to the adjacent shoreline area \((\text{max} = 2.18 \text{ m}^3 \text{ d}^{-1} \text{ m}^{-1})\).

The contribution from the different model layers to the total groundwater discharge varies along the shoreline, although discharge along the shoreline is predominately from layers 6 and 7 (Figure 4.4d). Model geologic layers 5 and 6 represent the Inter-Newmarket aquifer and Lower Newmarket till aquitard units respectively, while layer 7 represents the regionally extensive Thorncliffe Aquifer complex. The Inter-Newmarket aquifer sediments range from silt to gravel, while the Thorncliffe aquifer is comprised of glaciolacustrine silt and fluvial gravel. Layer 6 and 7 also incorporate the discontinuous channel silt and
channel aquifer units respectively. The York Model has a total of 9 geologic model layers (Figure 4.5), however layers 8 and 9 are ~20 m deeper than lake level (219 masl) and therefore do not intersect the lake bed in the nearshore area (lake depth 2 - 8 m; Ontario Ministry of Natural Resources, 2006). Therefore, it was assumed that these layers do not contribute to nearshore groundwater discharge. In the western shoreline area, near Keswick Beach, the simulated groundwater discharge is mostly from layer 7 (51%), with some contribution from layer 6 (26%). Conversely, in the Willow Beach and Duclos Point areas, the largest contribution is from layer 6 (50% for both areas), with some contribution from layer 7 (26% and 19% respectively). Similar to the Oro-Hawkeststone area, the depth of geologic model layers relative to the lake surface at the lake model boundary varies along the shoreline. For example, in the Duclos Point area layer 7 is shallower (< 10 m) at the lake boundary (Figure 4.5 section A-A’) than in the Keswick Beach area (~20 m; Figure 4.5 section C-C’).
Figure 4.4: Comparison results for (a) the York Region model area, scale is shown. Data from (b) $^{222}\text{Rn}$ boat surveys, and (c) $^{222}\text{Rn}$ mass balance calculations are compared with model estimates of groundwater discharge along the shoreline. Model estimates can be broken down into contribution from each model layer, shown in (d). Three areas with elevated $^{222}\text{Rn}$ concentrations and groundwater discharge are shown at Keswick Beach (KB), Willow Beach (WB), and Duclos Point (DP).
Figure 4.5: Cross section view of the York Region model geologic layers in the western (A-A’), central (B-B’), and eastern (C-C’) shoreline areas within the model domain, shown on map (right). High and low permeability geologic layers (i.e. aquifer and aquitard) are shown in yellow and blue respectively, channel sediment is shown in white, and the Oak Ridges Complex is shown in orange. Lake level is shown at 219 masl (Earthfx Inc., 2013 b,c).

4.4.1.3 Reasons for Discrepancies between Groundwater Discharge Estimates

Discrepancies between the simulated and $^{222}\text{Rn}$-derived groundwater discharge in the Oro-Hawkestone and York Region areas highlight the limitations of the $^{222}\text{Rn}$ survey method as well as the regional scale model simulations. The accuracy of the model results are only as good as the hydrogeologic information incorporated into the model. The underlying hydrogeologic framework in the Oro-Hawkestone model is well defined with the geology and hydrogeology in this area well characterized (Beckers & Frind, 2001; Burt & Dodge, 2011; Earthfx Inc., 2013b). In contrast, the hydrogeologic framework in the York Region model is not as well defined with information compiled from many sources given the large size and complexity of the model domain (Earthfx Inc., 2013c). As the objective of the model was not to quantify groundwater inputs into Lake Simcoe, efforts to refine the hydrogeologic framework were not focused along the shoreline area. Based on this, it is
not unexpected that the $^{222}$Rn-derived and simulated groundwater discharge compare better along the Oro-Hawkestone shoreline than the York Region shoreline.

The difference between the $^{222}$Rn-derived and simulated discharge values may also partially be because the $^{222}$Rn survey method is more sensitive to groundwater inputs from aquifer layers that are intersecting the lake bed in the shallow nearshore area. Along the Oro-Hawkestone shoreline, the aquifer layers that contribute the highest groundwater discharge to the lake are shallow relative to the lake surface along the northern shoreline, and become deeper relative to the lake surface along the central and southern shoreline (shown in cross sections Figure 4.3). This may explain why, for example, the $^{222}$Rn-derived discharge values are higher relative to the simulated discharge values in the Shingle Bay area compared to Carthew Bay and Shanty Bay - the layers contributing the highest groundwater discharge are intersecting the lakebed in the shallow, nearshore area in Shingle Bay. A similar finding can be made for the York Region area (shown in Figure 4.4d, cross sections in Figure 4.5). For instance, the simulated groundwater discharge around Keswick Beach is primarily from layer 7, which is 10 – 20 m below the lake surface, while the nearshore lake depth is less than 10 m (Figure 4.5; Ontario Ministry of Natural Resources, 2006). Conversely, in the Duclos Point area, the simulated groundwater discharge is predominately from layer 6, which is within 10 m of the lake surface, the same range as the nearshore lake depth. This may explain why the $^{222}$Rn-derived groundwater discharge near Duclos Point and Keswick Beach are similar, but the simulated groundwater discharge near Keswick Beach is much higher. Additionally, there may be some uncertainty associated with $^{222}$Rn-derived groundwater discharge values as a result of assumptions made within the mass balance calculations (i.e. use of average $C_{gw}$, similar offshore mixing patterns for large areas of the lake).

The observed differences between the simulated and $^{222}$Rn-derived groundwater discharge may also be due to the resolution of the model and field survey methods, and their respective ability to characterize smaller scale groundwater discharge features. For example, the model simulated shoreline around Duclos Point may not be able to capture smaller scale discharge features around the point itself that are captured in the $^{222}$Rn survey. In addition, anthropogenic alterations to the shoreline (i.e. dredged canals, presence of pier
structures) may cut through aquitard layers and alter groundwater flow patterns in the nearshore area—these alterations are not captured in the model simulations (Burnett et al., 2006; Santos et al., 2008). This may contribute to discrepancy between the $^{222}$Rn-derived and simulated measurements at the base of Duclos Point on the west side where there is a large marina/pier structure that may connect shallow and deeper aquifers layers.

### 4.4.2 History of Discharging Groundwater

The Oro-Hawkestone model was used to determine the history of the discharging groundwater and thus evaluate the potential influence of the groundwater inputs on the lake water quality. This analysis was not performed using the York Region model due to the lower resolution of the underlying hydrogeologic model and its weaker comparison with the $^{222}$Rn field results. The simulated groundwater discharge along the Oro-Hawkestone area is predominately from geologic model layer 6 (25%) and layer 7 (36%) with a lesser contribution from layer 4 (11%) and layer 5 (14%). As such, back particle tracking was conducted with particles initially placed along Oro-Hawkestone shoreline in the geologic model layers 4, 5, 6 and 7 (Figure 4.6 and Figure 4.7). The results show that the particle flow paths and travel times vary considerably along the shoreline. In general, flow paths in the northern area are long, originating in the Oro Moraine, while flow paths in the southern area are shorter originating within 1–2 km of the shoreline. In the central shoreline area, the flow paths are both long and short, with short flow paths associated with discharge from the shallower aquifer layers (layers 4 and 5) and longer flow paths associated with discharge from the deeper aquifer layers (layers 6 and 7). The particle travel times, however, vary based on the hydraulic conductivity and porosity of the different geologic layers. For example, in the central shoreline areas, particles originating in the shallow layer 4 have short flow paths (< 2 km from the shoreline) but very long travel times of ~400 years (Figure 4.6a), whereas particles originating in the deep layer travel over 5 km in less than 100 years (Figure 4.6c).
Figure 4.6: Back particle tracking from the Lake Simcoe shoreline with particles originating along the lake model boundary in geologic model layers (a) 4, (b) 5, (c) 6, and (d) 7 and tracking back to their recharge point.
Figure 4.7: Cross sections within the Oro-Hawkestone model in the northern (A-A’), central (B-B’), and southern (C-C’) shoreline areas, shown on map (right), with flow paths of particles backward tracked from the shoreline to the water table recharge point, and travel times labelled. Note that some particles enter and exit in the transverse direction to the cross-sections shown.

Land use at the groundwater recharge point and groundwater travel time influence the chemistry of the groundwater discharging to the lake. The particle tracking results indicate that groundwater discharging in the northern shoreline area travels along deep groundwater flow paths, and may take thousands of years to travel from the recharge point at the Oro Moraine to the lake (Figure 4.7). This not only provides long time scales for interactions between the aquifer sediments and groundwater but the groundwater age suggests that discharging groundwater in this area is unlikely to be adversely impacting the lake water quality with respect to key anthropogenic pollutants of concern (i.e. P and chloride). Conversely, in the central and southern shoreline areas, the groundwater discharge is
associated with short and shallow groundwater flow paths with travel times as low as 50 years. In the shallower layers 4 and 5, for example, the percentage of flow paths whose total travel time is less than 50 years is 1.4% and 0.5% respectively. Further, the percentage of flow paths in these layers with total travel times less than 100 years is 12% and 6% respectively. The land use in the nearshore areas where the groundwater recharges are predominately agricultural with some urban areas (Figure 4.8). As such, the discharging groundwater is more likely to be enriched with nutrients and other urban pollutants (including chloride) that may degrade lake water quality in the areas where it is discharging to the lake. In particular, shallow groundwater flow paths with relatively short travel times, may be more vulnerable to anthropogenic pollutants. The deeper layers 6 and 7 are characterized by longer travel times, in general, and the percentage of vulnerable flow paths is smaller.

Travel times of 50 years and greater for these flows paths also indicates that pollutant inputs via groundwater discharge may represent a legacy issue, with delay between land application of pollutants (e.g., nutrients) and the ultimate discharge of the pollutants to the lake. This may have potential long term implications for lake water quality. Moreover, it means that the current management efforts targeted at reducing pollutant inputs to the lake may be partially buffered by the long travel times for pollutants to reach the lake via groundwater pathways. It is also important to consider the lake conditions in the areas where groundwater is discharging. For example, a shallow bay area such as the northern Shingle Bay, is largely sheltered from lake mixing effects and therefore the impacts of groundwater pollutants on the lake water quality may be greater than in the central shoreline area where the lake is much deeper and more exposed to in-lake mixing effects that may dilute groundwater pollutants discharging.
Figure 4.8: (a) Land use in the Oro-Hawkestone model area, modified from Ontario Ministry of Natural Resource and Forestry SOLRIS (2011) and (b) flow path locations of particles released all model layers.

4.5 Conclusions

Comparison between $^{222}\text{Rn}$-derived and model simulated groundwater discharge to Lake Simcoe along two shoreline areas show good agreement with respect to the total groundwater discharge amounts as well as the spatial pattern of discharge. The good agreement between the methods builds confidence in groundwater discharge results, and discrepancies that were observed highlight the limitations of these regional scale approaches. The results illustrate that the goodness of the model results depends on the accuracy and resolution of the hydrogeologic information incorporated into the model. Adequate characterization of the hydrogeological system is not always available and in areas without adequate characterization use of regional scale groundwater flow models for estimating direct groundwater discharge may not be feasible. The results also illustrate that the $^{222}\text{Rn}$ approach is more sensitive to groundwater discharging to the nearshore from shallow geologic layers, whereas the regional scale models have better capability to estimate discharge also from deeper aquifer layers that may intercept the lake bed further offshore.
Particle tracking analysis performed using the Oro-Hawkestone regional scale model demonstrated the value of understanding the groundwater history in evaluating the potential impact of direct groundwater discharge on the lake water quality. While groundwater discharging in the northern shoreline area is characterized by long flow paths and long travel times (>1000 years), groundwater discharging in the southern shoreline area is characterized by short flow paths originating in nearshore agricultural areas with shorter travel times (50-200 years). This is important for evaluating the potential impact of groundwater discharge on lake water quality in that groundwater discharge in the north is not expected to be a major source of anthropogenic pollutants (i.e. nutrients and chlorides) to the lake, while groundwater discharging to the south is more likely to deliver pollutants associated with agricultural land use in the nearshore areas. This information is needed together with volume of groundwater discharge to inform management efforts focused on evaluating and, if needed, managing the groundwater pathway as a source of pollutants to the lake.
4.6 References


Earthfx Inc. (2013c). *Tier 3 Water Budget - Water Quantity Risk Level Assignment Study*


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Chapter 5

5 Summary and Recommendations

5.1 Summary

Direct groundwater discharge may be an important transport pathway for delivering pollutants to large glacial lakes. Typically, this pathway is poorly understood; in part due to high spatial variability of groundwater discharge. Groundwater discharge can be associated with high pollutant inputs to a lake, however, this depends on the history of the discharging groundwater (i.e. activities in recharge area, groundwater flow paths, and travel time). Understanding the spatial distribution and history of groundwater discharge is necessary to develop effective lake water quality management actions. This thesis evaluated groundwater discharge to Lake Simcoe, a large glacial lake, using field \(^{222}\text{Rn}\) data and regional scale groundwater models. Four specific research objectives were addressed.

The first objective was to quantify direct nearshore groundwater discharge and identify groundwater discharge ‘hotspots’ in a large glacial lake using \(^{222}\text{Rn}\) as a tracer. \(^{222}\text{Rn}\) has been used extensively as a tracer for groundwater discharge in many environments, including lakes. \(^{222}\text{Rn}\) boat surveys were conducted along 80% of the Lake Simcoe shoreline, with data used to estimate nearshore direct groundwater discharge using a \(^{222}\text{Rn}\) mass balance approach. Groundwater discharge showed considerable spatial variability around the lake, and distinct and repeatable groundwater discharge ‘hotspots’ were identified at Shingle Bay, Johnson’s Beach, Keswick Beach, Duclos Point, and Thorah Centennial Park. Analysis of the influence of indirect groundwater discharge on identified ‘hotspots’ areas suggests that some of the areas (e.g. Thorah Centennial Park, Keswick Beach) may be under the influence of a combination of direct and in-direct groundwater inputs. While direct groundwater discharge represents a relatively small contribution by volume, preliminary calculations of phosphorus loading associated with the total groundwater discharge estimate suggest that groundwater discharge may be an important source of pollutants to the lake.
The second objective was to evaluate the hydrogeologic controls on the observed regional scale spatial variability of groundwater discharge to a large glacial lake. Studies that have used the $^{222}\text{Rn}$ mass balance method to quantify direct groundwater discharge often provide limited insight into the factors controlling spatial groundwater patterns, particularly the hydrogeological controls. By comparing the field estimated groundwater discharge to the hydrogeology around Lake Simcoe, factors that were found to potentially influence identified groundwater discharge hotspots included: permeable nearshore surficial sediments, proximity to recharge features (i.e. the Oro Moraine and the Oak Ridges Moraine), and the presence of tunnel channel erosional features along the lake’s southern shoreline. Identified groundwater discharge hotspots were found to be under the influence of one or a combination of these factors. These factors are consistent with parameters known to control the magnitude and spatial variability of direct groundwater discharge, including; regional aquifer sediment permeability and hydraulic conductivity, nearshore topography, and amount of aquifer recharge. Conducting field assessments of groundwater discharge along several hundred kilometers of shoreline (as was done in this study) can be resource intensive, and the linkages between groundwater discharge hotspots and regional hydrogeologic controls can be broadly applied to other glacial lake environments to target future field investigations.

The third objective was to compare $^{222}\text{Rn}$-derived estimates of groundwater discharge to groundwater discharge simulated using regional scale numerical models. Comparison between groundwater discharge estimates was done along two areas of the Lake Simcoe shoreline: the Oro-Hawkestone area (northwestern shoreline), and the York Region area (southern shoreline). Groundwater discharge estimates from both regional scale methods showed good agreement with consistent results in several areas of elevated groundwater discharge. Discrepancies between the two methods suggest that model simulations were dependent on the accuracy of the underlying hydrogeologic information and model resolution, while $^{222}\text{Rn}$-derived estimates were dependent on the relative depth of geologic layers in the nearshore survey area (i.e. discharge from deeper layers could not be measured by the $^{222}\text{Rn}$ surveys). Comparing results from $^{222}\text{Rn}$ field surveys and groundwater models highlights the strengths and limitations of each method and increases confidence in the groundwater discharge estimates.
The fourth objective was to evaluate the history of discharging groundwater, and the potential implications of different groundwater flow paths and travel times on the lake water quality. Particle tracking analysis was performed using the regional scale groundwater model in the Oro-Hawkestone area with particle tracked backward from their discharge point at the lake shoreline to their recharge point. Results showed that groundwater discharging to the lake along the northern shoreline is characterized by long flow paths and travel times, while groundwater discharge in the south originates in the nearshore areas with shorter travel times. This may have important implications for the quality of discharging groundwater and its subsequent effect on lake water quality depending on: land use in the recharge area, chemical transformations that may take place in the given travel time, and lake conditions in the discharge area. Groundwater flowing along relatively short, shallow, flow paths may be more vulnerable to urban and agricultural contaminants than groundwater flowing along deep pathways, traveling thousands of years from its recharge point to the lake.

5.2 Recommendations

Chapter 3 evaluated direct nearshore groundwater discharge to a large glacial lake, and hydrogeologic controls on the observed regional scale spatial variability of this discharge. Recommendations for improving estimates of groundwater discharge ($Q_{gd}$) using the $^{222}$Rn mass balance method are as follows:

- A key uncertainty in evaluating $Q_{gd}$ using the $^{222}$Rn mass balance method is in quantifying the $^{222}$Rn losses due to horizontal offshore mixing ($J_{mix}$) in a large lake environment. The importance of this term in non-tidal environments is unclear, and it is therefore often largely ignored (Santos et al., 2008). In this study, $J_{mix}$ was considered in the mass balance due to the potential importance of offshore mixing processes, given the large size of Lake Simcoe. To estimate this term, $^{222}$Rn and conductivity data from three shore-perpendicular transects (Kempenfelt Bay, Cook’s Bay, and the main basin) were applied to solve for a $K_h$ value in each area of the lake. While this provided a reasonable initial estimate of $J_{mix}$ it is recommended that future analysis in large lakes calculate $K_h$ values at a higher
spatial resolution. Given that additional offshore $^{222}\text{Rn}$ boat surveys are resource intensive, continuous stationary $^{222}\text{Rn}$ measurements can be used as proposed by Burnett et al. (2001) and Dulaiova & Burnett (2008). This method involves investigating the changes in $^{222}\text{Rn}$ inventory (corrected for atmospheric evasion) during continuous stationary sampling, on short time scale (<1 hr), where the maximum negative flux rate is assumed to be a conservative estimate of $J_{\text{mix}}$.

- Further investigation of the effects of weather conditions (i.e. high wind speeds, waves, and precipitation) is recommended to better define the relationship between these events and decreases in measured in-lake $^{222}\text{Rn}$ concentrations. The time series investigation performed in this study served as preliminary analysis to discount surveys performed under high wind and precipitation, but it is only representative of a single set of storm conditions. Additional experiments performed in more areas of the lake, under a variety of wind speed, precipitation intensity and duration, and wave height conditions would help to better characterize the relationship.

- The analysis of seasonal and annual groundwater discharge as a percentage of total tributary inputs to Lake Simcoe presented in this study assumes that groundwater discharge is constant over time. To investigate this assumption, future field measurements should focus on characterizing the temporal variability in groundwater discharge.

- To refine estimates of pollutant loading associated with groundwater discharge, it is recommended that shallow groundwater samples be collected around the lake, particularly in areas identified as having elevated groundwater inputs (i.e. in groundwater discharge hotspots). The preliminary calculations for nutrient loading presented in Chapter 3 serve as an initial estimate, but this estimate needs to be better refined. Measurements of urban and rural pollutants (i.e. chloride, nitrate, phosphorous) in shallow groundwater discharging to different areas of the lake could provide insight into the spatial variability in groundwater quality, and a more accurate estimate of the pollutant loading to the lake from groundwater discharge.
Chapter 4 of this thesis compared $^{222}$Rn-derived and model simulated estimates of groundwater discharge along two shoreline areas of Lake Simcoe and evaluated the history of discharging groundwater through particle tracking analysis. Key recommendations for strengthening understanding of the spatial variability of groundwater discharge and potential implications for lake water quality are as follows:

- Future work should focus on characterizing groundwater chemistry in areas where elevated groundwater inputs were identified. The findings of the particle tracking analysis in the Oro-Hawkestone area indicate that groundwater discharge along the shoreline from north to south may have very different chemical composition given variable travel times and recharge areas. To confirm these findings, and to evaluate potential implications for lake water quality, field investigations should also examine the lake water chemistry in these areas. For example, Dulaiova (2010) used a commercially available automated nutrient analyzer to measure nitrate and nitrite, alongside a continuous $^{222}$Rn measurement system.

- To strengthen the conclusions drawn from the particle tracking analysis, future work should investigate historical land uses in recharge areas associated with both long and short groundwater transport pathways. This investigation, along with flow path and groundwater travel time information, could provide additional insight into the potential for legacy groundwater pollutant inputs that may act to buffer the results of current land use and pollution management strategies.

- Although particle tracking provides valuable information about groundwater flow paths including vulnerable recharge areas, it does not directly simulate the pollutant flux through each aquifer layer. To better understand the movement of pollutants through the aquifer system, from their recharge zone to discharge points along the lake shoreline, it is recommended that contaminant transport modelling be conducted.

- The comparison between $^{222}$Rn-derived and model simulated estimates for groundwater discharge to Lake Simcoe provides confidence in the groundwater discharge results presented in this thesis and also demonstrates the ability of previously developed regional groundwater flow models to be applied for
estimating groundwater discharge to lakes. Given that many regional scale groundwater models exist for source water and water resource protection initiatives, it is recommended that additional comparisons be done for other model areas in the Lake Simcoe area, and in the Great Lakes basin more generally, where the model resolution and underlying hydrogeologic model are sufficient.
5.3 References


## Appendices

### Appendix 1: Additional Site Information

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<th>Area</th>
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<tr>
<td><strong>Kempenfelt Bay North</strong> <em>(Johnson’s Beach, JB)</em></td>
<td><strong>Geography</strong>&lt;br&gt;- Kempenfelt Bay is in the western area of Lake Simcoe&lt;br&gt;- Kempenfelt Bay north, and Johnson’s Beach, are within the Barrie Creeks and Oro Creeks South subwatersheds&lt;br&gt;<strong>Topography</strong>&lt;br&gt;- Situated within a topographic low (220 masl) that is likely a tunnel channel valley&lt;sup&gt;1&lt;/sup&gt;&lt;br&gt;- Glacial deposits form topographic high to the north (375 masl) and south (300 masl) of the valley&lt;sup&gt;2,3&lt;/sup&gt;&lt;br&gt;<strong>Physiography</strong>&lt;br&gt;- North of Kempenfelt Bay, in the Oro Moraine region, is the Simcoe Uplands&lt;sup&gt;3&lt;/sup&gt;&lt;br&gt;- Kempenfelt Bay is within the Simcoe Lowlands&lt;sup&gt;3&lt;/sup&gt;&lt;br&gt;- South of Kempenfelt Bay is part of the Peterborough Drumlin Field&lt;sup&gt;7&lt;/sup&gt;&lt;br&gt;<strong>Surficial Sediment</strong>&lt;br&gt;- Surficial sediment consists of glacial diamicton, ice contact and outwash sands and gravels, and glaciolacustrine silts and clays&lt;sup&gt;3&lt;/sup&gt;&lt;br&gt;- Uplands are till and fine sediment, with other areas, such as the Oro Moraine, characterized by sands and gravels&lt;sup&gt;3&lt;/sup&gt;&lt;br&gt;- East-west trending valley surficial sediment is glaciolacustrine sand&lt;sup&gt;3&lt;/sup&gt;&lt;br&gt;<strong>Hydrogeology</strong>&lt;br&gt;- General groundwater flow to Kempenfelt Bay comes from topographic highs to the north and south&lt;sup&gt;3&lt;/sup&gt;&lt;br&gt;- In the north, the Oro Moraine acts as a large recharge feature, recharging local and regional aquifer layers&lt;br&gt;- There are four prominent aquifer units in the area, and the upper two are unconfined and associated with nearshore groundwater discharge&lt;sup&gt;3&lt;/sup&gt;&lt;br&gt;- The Oro Moraine, and large tunnel channel features may influence groundwater flow in the area&lt;sup&gt;4&lt;/sup&gt;</td>
</tr>
<tr>
<td><strong>Oro North</strong> <em>(Shingle Bay, SB)</em></td>
<td><strong>Geography</strong>&lt;br&gt;- Northwestern shoreline area of Lake Simcoe&lt;br&gt;- The Oro North area, and Shingle Bay are located within the Oro Creeks North subwatershed&lt;br&gt;<strong>Topography</strong>&lt;br&gt;- Topographic high point in the area occurs along the Oro Moraine (405 masl)&lt;sup&gt;2&lt;/sup&gt;&lt;br&gt;- Adjacent topographic low points occur along the Lake Simcoe shoreline (219 masl)&lt;sup&gt;2&lt;/sup&gt;</td>
</tr>
</tbody>
</table>
Physiography
- Dominant physiographic regions are the Simcoe Uplands and the Simcoe Lowlands

Surficial Sediment
- Lowlands are dominated by lacustrine sands, with sparse silt and clay in the north
- Uplands are dominated by diamictons: silty-sand, some clay, Oro Moraine sediment
- Drift thickness in the area ranges from 50-100m, with thinner drift along the Oro North shoreline (<15m)

Hydrogeology
- The area is characterized by a deep, regional aquifer, and several more discontinuous local aquifer layers
- Oro Moraine aquifer is restricted to moraine boundaries
- Two local unconfined aquifer layers may contribute to nearshore groundwater discharge

Cook’s Bay East
(Keswick Beach, KB)
and Georgina Area
(Duclos Point, DP)

Geography
- This area is located along the southern shoreline of the lake’s main basin and the eastern shoreline of Cook’s Bay
- Keswick beach is near the mouth of the Maskinonge River, on the east side of Cook’s Bay, within the Maskinonge River subwatershed
- Duclos Point is on the southern shoreline, near Georgina Island, within the Black River subwatershed
- Three major rivers flow into the lake in this area: East Holland River, Maskinonge River, and the Black River

Topography
- Topography in the area is characterized by a high point at the Oak Ridges moraine (340 masl) in the southern part of the Black River watershed, which slopes down to topographic low points along the lake’s shoreline (219 masl)

Physiography
- Highland areas are part of the Oak Ridges moraine physiographic region
- Lowland areas belong to the Schomberg Clay Plain, and the Simcoe Lowlands
- Thickness of quaternary sediments varies within the study area- thickest within erosional channels and beneath the Oak Ridges Moraine and thinnest in the northern part of the Black River subwatershed

Surficial Sediment
- The Oak ridges moraine area is dominated by sand and gravel deposits
- Surficial sediment in the lowland areas, closer to the lake, consists of lacustrine sand, silt, and clay deposits

References:
1. Earthfx Inc. & Gerber Geosciences (2008)
2. Genivar Inc. (2013)
3. LSRCA (2010)
4. LSRCA (2010a)
### Hydrogeology
- There are three main aquifer units in the area. The shallowest is the Oak Ridges aquifer complex, and the two deeper units are the Thorncliffe aquifer complex and the Scarborough aquifer complex.
- The shallow aquifer is separated from the deeper units by the Newmarket Till regional aquitard.
- Erosional channels in the Newmarket Till play a role in the groundwater flow system and are present below Cook’s bay in the Holland Market and within the Maskinonge and Black River subwatersheds.

### Geography
- The East Shore area is located within the Talbot River, Whites Creek, and Beaver River subwatersheds along the eastern shoreline of Lake Simcoe.
- The Talbot and Beavers rivers represent major rivers in the lake’s watershed.

### Topography
- Topography in the area is variable.
- The north is relatively flat, with high points north of the Talbot river (304 masl) and low points along the lake’s shoreline (217 masl).
- In the south, the topography is slightly steeper with the high point at the Oak Ridges Moraine (340 masl).

### Physiography
- Dominant physiographic regions are Simcoe Lowlands, and the Carden Plain in the north, and Oak Ridges Moraine in the south.

### Surficial Sediment
- Simcoe lowland regions are characterized by clay and organics, with lacustrine sand plains in lower lying areas.
- The Carden plain is characterized by thin soil covering limestone.
- Oro Moraine deposits consist of surficial sand and gravel deposits.
- Sediment thickness increases from north to south.

### Hydrogeology
- Hydrogeology in the northern area is characterized by a regionally discontinuous sand and gravel aquifer, underlain by several till, silt and clay layers. Bedrock weathering also plays a role in regional groundwater flow in this area, due to the shallow bedrock depth.
- In the southern area, there are several geologic layers that are laterally continuous across watersheds, but are most prominent in the Black river subwatershed area, including the Oak Ridges Aquifer complex (an aquifer system influenced by recharge of Oak Ridges Moraine sediment) and discontinuous Newmarket till (represents an aquitard, discontinuous because erosional channel features have incised this layer).

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**East Shore (Thorah Centennial Park, TCP)**

References:
1. Earthfx Inc. (2014)
2. LSRCA (2012b)
3. LSRCA (2012a)
References


Appendix 2: Instrument Response Time and Uncertainty

$^{222}$Rn boat survey measurements were conducted using several equipment combinations: including 5 RAD7s and 1 RAD AQUA, 4 RAD7s and 1 RAD AQUA, and 2x 2 RAD7s with 1 RAD AQUA (shown in Figure A2.1). For each equipment combination, the response time to changes in $^{222}$Rn concentrations was tested using control experiments in the laboratory. For these experiments measurements were taken over several cycles in one water source, and then switched to a different water source with a different $^{222}$Rn concentrations to observe the response.

![Diagram of equipment combinations](image)

**Figure A2.1:** Schematic diagrams of the equipment combinations used to conduct $^{222}$Rn boat surveys, including (a) 5 RAD7s and 1 RAD AQUA, (b) 4 RAD7s and 1 RAD AQUA, and (c) 2x 2 RAD7s with 1 RAD AQUA

The results of the control experiments are shown in Figure A2.2. Experiments were conducted by alternating measurements between de-gassed water (blue sections in Figure A2.2), which acted as the lower $^{222}$Rn concentration source, and tap water (orange sections in Figure A2.2), which served as the high $^{222}$Rn concentration source. The response time
of each system was analyzed by taking the average $^{222}\text{Rn}$ concentration over the number of cycles represented by each water source, and determining which cycle lag time had the best fit with the recorded $^{222}\text{Rn}$ data (recall that 1 RAD7 cycle is 15 minutes). For example, the data from the 5-RAD7 system experiment in figure Figure A2.2a is compared to a 1-cycle lag in response time (dotted line). $^{222}\text{Rn}$ measurements were taken in de-gassed water for cycles 1-5 and 16-20, so the ‘1 cycle lag time’ plot shows average measurements taken from cycles 2-6 and 17-21. Similarly, $^{222}\text{Rn}$ measurements were taken in tap water for cycles 6-15 and 21-27, so the average measurements taken from cycles 7-16 and 22-27 are shown. The ‘Average $^{222}\text{Rn}$ concentration’ cycle lags shown in Figure A2.2 represent the best fit for each system. Results show that the 5-RAD7 and 2-RAD7 systems are subjected to a 1-cycle lag in response time, while the 4-RAD7 system has a 2-cycle lag time. This data was used to adjust spatial $^{222}\text{Rn}$ maps accordingly, depending on the equipment combination used for each survey.

Figure A2.2: Results of control experiments conducted using (a) 5 RAD7s and 1 RAD AQUA, (b) 4 RAD7s and 1 RAD AQUA, and (c) 2x 2 RAD7s with 1 RAD AQUA.
The measurement uncertainty for each system was also investigated. The reported error for each $^{222}\text{Rn}$ measurement is the standard deviation ($\sigma$) following Poisson statistics; whereby an increase in the number of counts and the number of RAD7 units will reduce the measurement uncertainty (Taylor, 1982; Durridge Co.). A summary of the average $^{222}\text{Rn}$ concentrations recorded during each experiment, for each water source measurement, is shown in Table A2.1. The results show that the average error, as a percentage of the $^{222}\text{Rn}$ measurement, is lowest for the 5-RAD7 system and highest for the 2-RAD7 system. The 2-RAD7 system, however, appears to have the best sensitivity to changes in source water $^{222}\text{Rn}$ concentration. For example, the average $^{222}\text{Rn}$ concentrations between the first and second set of de-gassed and tap water cycles are very similar, compared to the other two equipment systems.

<table>
<thead>
<tr>
<th>Equipment System</th>
<th>Average Concentration (dpm L$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>De-gassed Water (1)</td>
</tr>
<tr>
<td>5 RAD7</td>
<td>0.45 ± 0.10</td>
</tr>
<tr>
<td>4 RAD7</td>
<td>0.55 ± 0.14</td>
</tr>
<tr>
<td>2 RAD7</td>
<td>0.42 ± 0.39</td>
</tr>
</tbody>
</table>

Table A2.0.1: Results of control experiments conducted using (a) 5 RAD7s and 1 RAD AQUA, (b) 4 RAD7s and 1 RAD AQUA, and (c) 2x 2 RAD7s with 1 RAD AQUA. The average $^{222}\text{Rn}$ concentrations are based on the best-fit cycle lag time shown in Figure A2.2.
Appendix 3: In-lake Electrical Conductivity Measurements

In-lake electrical conductivity measurements were taken alongside $^{222}\text{Rn}$ measurements during boat surveys. Figures A3.1-A3.4 are examples of surveys conducted in different areas of the lake. No clear relationship was observed between $^{222}\text{Rn}$ and conductivity measurements. The lack of relationship may because electrical conductivity values in Lake Simcoe are overwhelmed by urban sources.

![Graph showing electrical conductivity and $^{222}\text{Rn}$ measurements over time.]

**Figure A3.1:** $^{222}\text{Rn}$ and electrical conductivity measurements for a boat survey performed on August 24, 2017 in the Oro North area.

![Graph showing electrical conductivity and $^{222}\text{Rn}$ measurements over time.]

**Figure A3.2:** $^{222}\text{Rn}$ and electrical conductivity measurements for a boat survey performed on August 28, 2017 in the East shore area.
Figure A3.3: $^{222}\text{Rn}$ and electrical conductivity measurements for a boat survey performed on May 25, 2018 in the Cook’s Bay East area.

Figure A3.4: $^{222}\text{Rn}$ and electrical conductivity measurements for a boat survey performed on July 4, 2018 in the Kempenfelt Bay North area.
Appendix 4: Time Series Testing

Time series testing was conducted from September 23-30, 2017 at a location along the southern shoreline of Lake Simcoe, to assess the effects of high wind speeds, precipitation, and waves on measured in-lake $^{222}$Rn concentration. A summary of the results is shown in Figure A4.1. There are two areas of particular interest indicated in the figure; (1) where the in-lake $^{222}$Rn inventory is affected by sustained high wind speeds (>20 km h$^{-1}$) and waves, and (2) where the in-lake $^{222}$Rn inventory is affected by precipitation and waves.

In the first area, from 0:00 September 27th to 6:00 September 28th (approximately 30 hours) there is a significant drop in measured in-lake $^{222}$Rn concentrations that corresponds to increases in wind speed up to 40 km h$^{-1}$ and waves (represented by the water pressure measurements in Figure A4.1c). The $^{222}$Rn concentration begin to drop after ~6-12hr. The $^{222}$Rn inventory, corrected for losses due to atmospheric evasion predicted by the gas transfer equations (Section 3.3.2; Figure A4.1b) is over-estimated under sustained wind speeds of this magnitude, which suggests that these equations may not be suitable under these conditions. In the second area, from 6:00 September 29th to 12:00 September 30th (approximately 30 hours), the average wind speeds are lower compared to the first area. However, due to a cumulative 10mm of precipitation the $^{222}$Rn concentrations do not recover to their initial levels (Figure A4.1a). Further, despite the drop in wind speed from 0:00 September 30th to 12:00 September 30th, no increase in $^{222}$Rn concentration was observed for the remainder of the experiment.

The results of this testing indicate that the equations to correct for $^{222}$Rn losses due to atmospheric evasion may not be suitable to ‘correct’ $^{222}$Rn concentrations during the conditions observed (i.e. high sustained wind speeds, waves, and precipitation). These results were used as the basis to discount $^{222}$Rn surveys performed within 6-12hr of sustained onshore wind speeds >20 km h$^{-1}$ and precipitation events.
Figure A4.1: A summary of the data collected during time series testing. Plots show (a) the average $^{222}$Rn concentrations, the instantaneous wind speed from two different weather stations, and precipitation, (b) the $^{222}$Rn concentrations corrected for atmospheric evasion, and the corresponding average wind speeds used in the correction, and (c) pressure transducer measurements at the sampling location.
## Appendix 5: Groundwater Endmember

Table A5.1 Average groundwater endmember $^{222}\text{Rn}$ concentrations for groundwater samples and sediment equilibration samples from all sampling sites.

<table>
<thead>
<tr>
<th>Sample Site</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Groundwater Sample $^{222}\text{Rn}$ Concentration (dpm L$^{-1}$)</th>
<th>Sediment Sample $^{222}\text{Rn}$ Concentration (dpm L$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>OB</td>
<td>44.44828</td>
<td>-79.50852</td>
<td>76.5 ± 40.9</td>
<td>-</td>
</tr>
<tr>
<td>JB</td>
<td>44.39295</td>
<td>-79.65737</td>
<td>235.2 ± 43.7</td>
<td>373.8 ± 9.1</td>
</tr>
<tr>
<td>CB</td>
<td>44.38058</td>
<td>-79.68924</td>
<td>66.4 ± 14.8</td>
<td>-</td>
</tr>
<tr>
<td>MPB</td>
<td>44.37622</td>
<td>-79.66841</td>
<td>99.0 ± 17.1</td>
<td>-</td>
</tr>
<tr>
<td>WKB</td>
<td>44.36987</td>
<td>-79.63274</td>
<td>92.0 ± 19.9</td>
<td>-</td>
</tr>
<tr>
<td>WLB</td>
<td>44.31160</td>
<td>-79.42447</td>
<td>109.3 ± 7.9</td>
<td>-</td>
</tr>
<tr>
<td>JPB</td>
<td>44.32049</td>
<td>-79.38495</td>
<td>159.0 ± 10.2</td>
<td>260.6 ± 8.1</td>
</tr>
<tr>
<td>BMB</td>
<td>44.46381</td>
<td>-79.49012</td>
<td>309.3 ± 26.4</td>
<td>195.5 ± 8.5</td>
</tr>
<tr>
<td>HPB</td>
<td>44.38750</td>
<td>-79.68573</td>
<td>50.0 ± 9.0</td>
<td>-</td>
</tr>
<tr>
<td>TB</td>
<td>44.37430</td>
<td>-79.64308</td>
<td>109.0 ± 27.0</td>
<td>478.8 ± 10.0</td>
</tr>
<tr>
<td>LHB</td>
<td>44.3574</td>
<td>-79.53226</td>
<td>462.8 ± 38.5</td>
<td>-</td>
</tr>
<tr>
<td>10B</td>
<td>44.34345</td>
<td>-79.53593</td>
<td>55.0 ± 7.0</td>
<td>-</td>
</tr>
<tr>
<td>IPB</td>
<td>44.32171</td>
<td>-79.53147</td>
<td>50.0 ± 5.0</td>
<td>-</td>
</tr>
<tr>
<td>PB</td>
<td>44.31038</td>
<td>-79.43453</td>
<td>72.0 ± 12.0</td>
<td>-</td>
</tr>
<tr>
<td>CBP</td>
<td>44.23202</td>
<td>-79.47025</td>
<td>243.6 ± 10.4</td>
<td>477.6 ± 7.1</td>
</tr>
<tr>
<td>BPB</td>
<td>44.32878</td>
<td>-79.36707</td>
<td>239.8 ± 10.3</td>
<td>-</td>
</tr>
<tr>
<td>BB</td>
<td>44.43202</td>
<td>-79.16697</td>
<td>196.2 ± 9.5</td>
<td>132.6 ± 8.6</td>
</tr>
<tr>
<td>HPP</td>
<td>44.33804</td>
<td>-79.22674</td>
<td>295.4 ± 11.2</td>
<td>268.7 ± 8.7</td>
</tr>
<tr>
<td>SPB</td>
<td>44.33231</td>
<td>-79.31826</td>
<td>352.6 ± 13.4</td>
<td>436.0 ± 10.3</td>
</tr>
<tr>
<td>MB</td>
<td>44.58477</td>
<td>-79.36084</td>
<td>293.3 ± 9.9</td>
<td>460.0 ± 6.3</td>
</tr>
<tr>
<td>LC</td>
<td>44.54892</td>
<td>-79.21715</td>
<td>-</td>
<td>269.6 ± 8.3</td>
</tr>
<tr>
<td>MCP</td>
<td>44.56704</td>
<td>-79.33255</td>
<td>-</td>
<td>204.0 ± 9.0</td>
</tr>
<tr>
<td>TCP</td>
<td>44.46856</td>
<td>-79.15945</td>
<td>-</td>
<td>255.4 ± 11.7</td>
</tr>
<tr>
<td>DPD</td>
<td>44.32686</td>
<td>-79.27896</td>
<td>-</td>
<td>368.7 ± 10.5</td>
</tr>
</tbody>
</table>
Table A5.2: Distribution of groundwater endmember $^{222}$Rn concentrations for groundwater samples and sediment equilibration samples

<table>
<thead>
<tr>
<th>Radon Concentration (dpm L$^{-1}$)</th>
<th>Groundwater Samples</th>
<th>Sediment Samples</th>
<th>Average</th>
</tr>
</thead>
<tbody>
<tr>
<td>Minimum</td>
<td>50.0 ± 5.0</td>
<td>132.6 ± 8.6</td>
<td>91.3 ± 6.8</td>
</tr>
<tr>
<td>$25^{th}$ percentile</td>
<td>75.4 ± 11.8</td>
<td>255.4 ± 21.8</td>
<td>165.4 ± 16.8</td>
</tr>
<tr>
<td>$75^{th}$ percentile</td>
<td>256.0 ± 21.8</td>
<td>436.0 ± 28.4</td>
<td>346.0 ± 25.0</td>
</tr>
<tr>
<td>Maximum</td>
<td>462.8 ± 38.5</td>
<td>478.8 ± 10.0</td>
<td>470.8 ± 24.3</td>
</tr>
<tr>
<td>Median</td>
<td>134.2 ± 15.8</td>
<td>269.6 ± 22.3</td>
<td>201.9 ± 19.1</td>
</tr>
<tr>
<td>Average</td>
<td>178.3 ± 17.2</td>
<td>321.6 ± 9.0</td>
<td>234.8 ± 14.0</td>
</tr>
</tbody>
</table>

Figure A5.1: Box plot of groundwater endmember $^{222}$Rn concentrations for groundwater samples and sediment equilibration samples
Differences in the range of $^{222}$Rn concentrations measured using groundwater and sediment equilibration samples are primarily due to the differences in sampling sites for each method. Comparing results from the nine beach sites, where both types of samples were collected, shows that similar groundwater endmember $^{222}$Rn concentrations were measured using shallow groundwater sampling and sediment equilibration methods (Figure A5.2).

**Figure A5.2: Comparison of groundwater sample and sediment equilibration results for the nine beach sites where both types of samples were taken.**

Groundwater discharge calculations were done using the overall average groundwater concentration of $234.8 \pm 14.0$ dpm L$^{-1}$. To investigate the sensitivity of calculated groundwater discharge using the $^{222}$Rn mass balance method, groundwater discharge was re-calculated using the average 25$^{th}$ percentile (lower quartile) value of $165.4 \pm 16.8$ dpm L$^{-1}$ and the 75$^{th}$ percentile (upper quartile) value of $346.0 \pm 25.0$ dpm L$^{-1}$.
Table A5.3: Average direct nearshore groundwater discharge for each area of shoreline, using the 25th percentile groundwater endmember $^{222}$Rn concentration of $165.4 \pm 16.8$ dpm L$^{-1}$ (shoreline areas shown in Figure 3.7).

<table>
<thead>
<tr>
<th>Shoreline Area</th>
<th>Maximum (m$^3$d$^{-1}$m$^{-1}$)</th>
<th>Minimum (m$^3$d$^{-1}$m$^{-1}$)</th>
<th>Total Groundwater Discharge (m$^3$d$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Oro North</td>
<td>$2.27 \pm 0.45$</td>
<td>$0.05 \pm 0.05$</td>
<td>$17,300 \pm 6000$</td>
</tr>
<tr>
<td>Kempenfelt Bay North</td>
<td>$5.41 \pm 1.39$</td>
<td>$0.10 \pm 0.10$</td>
<td>$26,500 \pm 8500$</td>
</tr>
<tr>
<td>Kempenfelt Bay South</td>
<td>$4.72 \pm 1.38$</td>
<td>$1.07 \pm 0.13$</td>
<td>$28,500 \pm 4600$</td>
</tr>
<tr>
<td>Cook’s Bay West</td>
<td>$2.79 \pm 0.34$</td>
<td>$0.23 \pm 0.18$</td>
<td>$16,600 \pm 4200$</td>
</tr>
<tr>
<td>Cook’s Bay East</td>
<td>$6.15 \pm 0.97$</td>
<td>$0.81 \pm 0.11$</td>
<td>$37,500 \pm 6700$</td>
</tr>
<tr>
<td>Georgina</td>
<td>$6.47 \pm 0.87$</td>
<td>$0.99 \pm 0.26$</td>
<td>$63,000 \pm 13600$</td>
</tr>
<tr>
<td>East Shore</td>
<td>$11.46 \pm 1.28$</td>
<td>$0.41 \pm 0.14$</td>
<td>$44,400 \pm 5800$</td>
</tr>
<tr>
<td>North Shore</td>
<td>$4.64 \pm 0.52$</td>
<td>$0.11 \pm 0.02$</td>
<td>$46,200 \pm 10800$</td>
</tr>
<tr>
<td><strong>Total</strong></td>
<td></td>
<td></td>
<td><strong>280,000 \pm 60200</strong></td>
</tr>
</tbody>
</table>
Table A5.4: Average direct nearshore groundwater discharge for each area of shoreline using the 75th percentile groundwater endmember $^{222}\text{Rn}$ concentration of $346.0 \pm 25.0 \text{ dpm L}^{-1}$ (shoreline areas shown in Figure 3.7).

<table>
<thead>
<tr>
<th>Shoreline Area</th>
<th>Maximum (m$^3$d$^{-1}$m$^{-1}$)</th>
<th>Minimum (m$^3$d$^{-1}$m$^{-1}$)</th>
<th>Total Groundwater Discharge (m$^3$d$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Oro North</td>
<td>1.09 ± 0.20</td>
<td>0.02 ± 0.02</td>
<td>8,300 ± 2800</td>
</tr>
<tr>
<td>Kempenfelt Bay North</td>
<td>2.59 ± 0.63</td>
<td>0.05 ± 0.05</td>
<td>12,700 ± 3900</td>
</tr>
<tr>
<td>Kempenfelt Bay South</td>
<td>2.26 ± 0.64</td>
<td>0.51 ± 0.05</td>
<td>13,600 ± 1900</td>
</tr>
<tr>
<td>Cook’s Bay West</td>
<td>1.33 ± 0.13</td>
<td>0.11 ± 0.08</td>
<td>7,900 ± 1900</td>
</tr>
<tr>
<td>Cook’s Bay East</td>
<td>2.94 ± 0.41</td>
<td>0.39 ± 0.05</td>
<td>17,900 ± 2900</td>
</tr>
<tr>
<td>Georgina</td>
<td>3.09 ± 0.34</td>
<td>0.47 ± 0.12</td>
<td>30,100 ± 6000</td>
</tr>
<tr>
<td>East Shore</td>
<td>5.48 ± 0.47</td>
<td>0.19 ± 0.06</td>
<td>21,200 ± 2300</td>
</tr>
<tr>
<td>North Shore</td>
<td>2.22 ± 0.19</td>
<td>0.05 ± 0.01</td>
<td>22,100 ± 4800</td>
</tr>
<tr>
<td><strong>Total</strong></td>
<td></td>
<td></td>
<td><strong>133,800 ± 26500</strong></td>
</tr>
</tbody>
</table>
Appendix 6: $^{222}$Rn Survey and Mass Balance Results

Table A6.1: Summary of $^{222}$Rn boat survey dates and shoreline distance surveyed in each shoreline area, from June 2015- July 2018 (shoreline area also shown in Figure 3.7). The 24-hr average winds speeds shown are were calculated using wind speed records from the nearest Environment Climate Change Canada weather station (located 5 - 28 km away from a survey site).

<table>
<thead>
<tr>
<th>Area</th>
<th>Survey Date (MM/DD/YY)</th>
<th>Shoreline Distance Surveyed (km)</th>
<th>24-hr Average Wind Speed (km/h)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Oro North</td>
<td>08/26/17</td>
<td>11.3, 11.4</td>
<td>5.9</td>
</tr>
<tr>
<td></td>
<td>08/27/17</td>
<td>12.3, 14.4</td>
<td>6.0</td>
</tr>
<tr>
<td></td>
<td>08/31/17</td>
<td>5.6</td>
<td>15.9</td>
</tr>
<tr>
<td>Kempenfelt Bay North</td>
<td>06/09/15</td>
<td>3.8</td>
<td>11</td>
</tr>
<tr>
<td></td>
<td>06/11/15</td>
<td>14.1</td>
<td>18.5</td>
</tr>
<tr>
<td></td>
<td>07/06/15</td>
<td>8.6</td>
<td>9.9</td>
</tr>
<tr>
<td></td>
<td>07/08/15</td>
<td>14.0</td>
<td>11.4</td>
</tr>
<tr>
<td></td>
<td>07/09/15</td>
<td>8.7</td>
<td>5.2</td>
</tr>
<tr>
<td></td>
<td>07/04/18</td>
<td>9.6, 11.3</td>
<td>5.5</td>
</tr>
<tr>
<td>Kempenfelt Bay South</td>
<td>06/09/15</td>
<td>13.7</td>
<td>11</td>
</tr>
<tr>
<td></td>
<td>06/11/15</td>
<td>2.4</td>
<td>18.5</td>
</tr>
<tr>
<td>Cook’s Bay West</td>
<td>07/08/15</td>
<td>12.2</td>
<td>11.4</td>
</tr>
<tr>
<td></td>
<td>08/12/15</td>
<td>16.3</td>
<td>16.4</td>
</tr>
<tr>
<td>Cook’s Bay East</td>
<td>08/13/15</td>
<td>12.6</td>
<td>11.3</td>
</tr>
<tr>
<td></td>
<td>09/23/15</td>
<td>9.7</td>
<td>3.3</td>
</tr>
<tr>
<td></td>
<td>05/25/18</td>
<td>9.1</td>
<td>11.9</td>
</tr>
<tr>
<td></td>
<td>07/05/18</td>
<td>4.2, 4.1</td>
<td>5.7</td>
</tr>
<tr>
<td>Georgina</td>
<td>08/13/15</td>
<td>12.3</td>
<td>11.3</td>
</tr>
<tr>
<td></td>
<td>09/23/15</td>
<td>11.1</td>
<td>3.25</td>
</tr>
<tr>
<td></td>
<td>08/09/16</td>
<td>6.4</td>
<td>8.28</td>
</tr>
<tr>
<td></td>
<td>09/16/16</td>
<td>5.5</td>
<td>6.28</td>
</tr>
<tr>
<td></td>
<td>07/17/17</td>
<td>15.4, 15.5</td>
<td>7.4</td>
</tr>
<tr>
<td></td>
<td>07/19/17</td>
<td>9.0, 12.7</td>
<td>9.3</td>
</tr>
<tr>
<td></td>
<td>07/20/17</td>
<td>7.8</td>
<td>8.4</td>
</tr>
<tr>
<td>East Shore</td>
<td>08/09/16</td>
<td>12.0</td>
<td>7.2</td>
</tr>
<tr>
<td></td>
<td>08/28/17</td>
<td>7.0, 14.7, 6.7</td>
<td>16.7</td>
</tr>
<tr>
<td></td>
<td>09/01/17</td>
<td>11.0</td>
<td>11.8</td>
</tr>
<tr>
<td>Northshore</td>
<td>07/31/17</td>
<td>13.7</td>
<td>8.7</td>
</tr>
<tr>
<td></td>
<td>08/01/17</td>
<td>4.6</td>
<td>5.7</td>
</tr>
<tr>
<td></td>
<td>08/03/17</td>
<td>13.1</td>
<td>7.8</td>
</tr>
<tr>
<td></td>
<td>08/21/17</td>
<td>9.7</td>
<td>10.8</td>
</tr>
</tbody>
</table>
Figure A6.1: Data from individual surveys conducted for (a) the Oro North area. All (b) in-lake $^{222}$Rn concentrations and (c) corresponding calculated groundwater discharge values were plotted along the shoreline (shoreline area also shown in Figure 3.7).
Figure A6.2: Data from individual surveys conducted for (a) the Kempenfelt Bay North area. All (b) in-lake $^{222}\text{Rn}$ concentrations and (c) corresponding calculated groundwater discharge values were plotted along the shoreline (shoreline area also shown in Figure 3.7).
Figure A6.3: Data from individual surveys conducted for (a) the Cook’s Bay East area. All (b) in-lake $^{222}$Rn concentrations and (c) corresponding calculated groundwater discharge values were plotted along the shoreline (shoreline area also shown in Figure 3.7).
Figure A6.4: Data from individual surveys conducted for (a) the Georgina area. All (b) in-lake $^{222}$Rn concentrations and (c) corresponding calculated groundwater discharge values were plotted along the shoreline (shoreline area also shown in Figure 3.7).
Figure A6.5: Data from individual surveys conducted for (a) the East shore area. All (b) in-lake $^{222}$Rn concentrations and (c) corresponding calculated groundwater discharge values were plotted along the shoreline (shoreline area also shown in Figure 3.7).
Figure A6.6: Data from individual surveys conducted for (a) the Kempenfelt Bay South area. All (b) in-lake $^{222}$Rn concentrations and (c) corresponding calculated groundwater discharge values were plotted along the shoreline (shoreline area also shown in Figure 3.7).
Figure A6.7: Data from individual surveys conducted for (a) the Cook’s Bay West area. All (b) in-lake $^{222}$Rn concentrations and (c) corresponding calculated groundwater discharge values were plotted along the shoreline (shoreline area also shown in Figure 3.7).
Figure A6.8: Data from individual surveys conducted for (a) the Northshore area. All (b) in-lake $^{222}\text{Rn}$ concentrations and (c) corresponding calculated groundwater discharge values were plotted along the shoreline (shoreline area also shown in Figure 3.7).
Appendix 7: Estimating Offshore Mixing

Table A7.1: Summary of offshore $^{222}\text{Rn}$ transect data and corresponding horizontal offshore mixing coefficient ($K_h$; m$^2$ d$^{-1}$) and surface water advection ($\omega$; m d$^{-1}$) values for Kempenfelt Bay, Cook’s Bay, and the lake’s Main Basin; calculated using the iterative method outline in Santos et al. (2008).

<table>
<thead>
<tr>
<th>Lake Area</th>
<th>Offshore Transect</th>
<th>Calculated $K_h$ (m$^2$ d$^{-1}$)</th>
<th>Calculated $\omega$ (m d$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Distance Offshore (m)</td>
<td>In-lake $^{222}\text{Rn}$ Concentration (dpm L$^{-1}$)</td>
<td></td>
</tr>
<tr>
<td>Kempenfelt Bay</td>
<td>46</td>
<td>0.99 ± 0.23</td>
<td>2815</td>
</tr>
<tr>
<td></td>
<td>355</td>
<td>0.66 ± 0.15</td>
<td></td>
</tr>
<tr>
<td></td>
<td>483</td>
<td>0.00 ± 0.09</td>
<td></td>
</tr>
<tr>
<td>Cook’s Bay</td>
<td>136</td>
<td>3.30 ± 0.35</td>
<td>9032</td>
</tr>
<tr>
<td></td>
<td>481</td>
<td>1.48 ± 0.25</td>
<td></td>
</tr>
<tr>
<td></td>
<td>691</td>
<td>0.91 ± 0.22</td>
<td></td>
</tr>
<tr>
<td>Main Basin</td>
<td>171</td>
<td>1.32 ± 0.24</td>
<td>4325</td>
</tr>
<tr>
<td></td>
<td>281</td>
<td>0.78 ± 0.20</td>
<td></td>
</tr>
<tr>
<td></td>
<td>549</td>
<td>0.53 ± 0.18</td>
<td></td>
</tr>
</tbody>
</table>

Results of a comparison between the calculated groundwater discharge in each area of Lake Simcoe (Figure A7.1) show that the horizontal offshore mixing ($J_{mix}$), although a potentially important term for large lake environments, does not have a significant effect on the overall groundwater discharge value. The largest loss term in the mass balance is the loss due to atmospheric evasion.
Figure A7.1: Comparison of the average calculated groundwater discharge along the shoreline in the (a) Kempenfelt Bay North, (b) Cook’s Bay East, and (c) Georgina areas; with and without horizontal offshore mixing ($J_{mix}$) included as a loss term in the $^{222}$Rn mass balance.

Reference

<table>
<thead>
<tr>
<th><strong>Name:</strong></th>
<th>Hayley Wallace</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Post-secondary Education and Degrees:</strong></td>
<td>University of Western Ontario, London, Ontario, Canada, 2011-2016 BESc.</td>
</tr>
<tr>
<td><strong>Honours and Awards:</strong></td>
<td>Geotechnical Research Center R.M. Quigley Award, 2016-2017</td>
</tr>
<tr>
<td><strong>Related Work Experience:</strong></td>
<td>Teaching Assistant, The University of Western Ontario, 2016-2018</td>
</tr>
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