THE INFLUENCE OF CLIMATE CHANGE ON AQUIFER THERMAL AND HYDRAULIC REGIMES: IMPLICATIONS FOR FISH HABITAT

by

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Abstract

Groundwater discharge provides critical thermal refuge for cold-water fishes in many river systems in North America, but the impact of future climate change on the timing, magnitude, and temperature of groundwater discharge has not been previously considered. This dissertation investigates the hydraulic and thermal influence of climate change on unconfined aquifers, particularly in the context of the climate and hydrology of the Little Southwest Miramichi River in New Brunswick, Canada. The thermal sensitivity of groundwater to climate change was investigated via (1) analytical, (2) empirical, and (3) numerical methods. Firstly, an analytical solution was developed to the transient, one-dimensional conduction-advection equation subject to nonlinear initial and boundary conditions. Secondly, an empirical ground surface temperature to groundwater temperature transfer function was developed from measured surface and groundwater temperature data and then explicitly coupled to a soil-vegetation-atmosphere-transfer (SVAT) model to investigate the subsurface thermal influence of climate change. Thirdly, a finite element model of coupled groundwater flow and energy transport with pore water phase change (SUTRA) was linked to surficial energy and hydrologic models to investigate the influence of changing precipitation and air temperature on the timing, magnitude, and temperature of groundwater discharge from unconfined aquifers. In general, the simulation results demonstrate that the hydraulic and thermal regimes of shallow aquifers are very sensitive to changing precipitation and air temperature, and that future climate change will likely limit the distribution of suitable cold-water thermal refugia in warming rivers.
Dedication

To my wife Judith who endured many lonely evenings and weekends as I progressed through my graduate studies and who kindly feigned interest as I daily expounded the complexities of the hydrological sciences.
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List of Symbols, Nomenclature and Abbreviations

a  Geothermal gradient
AT  Air temperature
b  Empirical fitting parameter for surface boundary condition
B  Empirical parameter for the GST to groundwater temperature function
BC  Boundary condition
c  Empirical fitting parameter for surface boundary condition
C  Volumetric heat capacity of soil-water matrix (Chapter 4)
cp  Volumetric heat capacity of soil-water matrix (Chapter 3)
cw  Specific heat of water (Chapter 3)
Cw  Volumetric heat capacity of water (Chapter 4)
cwpw  Volumetric heat capacity of water (Chapter 3)
CGCM3  Canadian Global Climate Model, version 3
CMIP3  Third Coupled Model Inter-Comparison Project Database of GCM’s
CO2  Carbon dioxide
CRCM  Canadian Regional Climate model
d  Empirical fitting parameter for initial conditions function
D  Thermal diffusivity (Ch. 3) and thermal damping factor (Ch. 4)
DEM  Digital elevation model
DM  Downscaling method
DOY  Day of year
DT  Daily translation climate series post-processing method
EC  Environment Canada
erf  Error function
erfc  Complementary error function
ES  Emission scenario
exp  Exponential function
EZD  Evaporative zone depth

1 Symbols, nomenclature, and abbreviations that only appear in appendices are only defined therein. Due to the articles format of this dissertation, nomenclature usage may not be fully consistent, thus instances are noted where a particular symbol has alternative meanings in different chapters.
ForHyM2: Forest Hydrology Model, version 2
GCM: Global climate model
GIS: Geographic information system
GPR: Ground-penetrating radar
GST: Ground surface temperature
GWDT: Groundwater discharge temperature
GWT: Groundwater table
HELP3: Hydrologic Evaluation of Landfill Performance Model, version 3
HMLR: Hybrid multi-linear regression downscaling method
IPCC: Intergovernmental Panel on Climate Change
L: Length (Ch. 3 and 5); Lag time for signal to reach subsurface (Ch. 4)
Ln: Natural logarithm
LSW: Little Southwest Miramichi River
m: Mass
MAGST: Mean annual ground surface temperature
NB: New Brunswick
NBADW: New Brunswick Aquatic Data Warehouse
NSERC: Natural Science and Engineering Research Council of Canada
p: Laplace transform of time (frequency domain)
P: Pore water pressure
PDE: Partial differential equation
Pe: Thermal Peclet number
q: Darcy velocity
Q: River discharge
R²: Coefficient of determination
RCM: Regional climate model
RCP: Representative concentration pathways
RMSE: Root mean square error
SUTRA: Saturated- Unsaturated Transport Model
SVAT: Soil-vegetation-atmosphere transfer model
t: Time
$T$  
Temperature
---
$\bar{T}$  
Laplace transform of temperature (frequency domain)
---
$T_0$  
Initial surface temperature
---
$T_1$  
Empirical fitting parameter for boundary conditions function
---
$T_i$  
Empirical fitting parameter for initial conditions function
---
$T_f$  
Freezing temperature for bulk water (0°C)
---
$T_{\text{wresi}}$  
Temperature at which residual liquid water content remains
---
UNB  
University of New Brunswick
---
v  
Thermal plume velocity $= q c_w \rho_w / (c \rho)$
---
z  
Depth
---
$\delta$  
Empirical fitting parameter for initial conditions function (Eq. 3.4)
---
$\Delta$  
Change in any parameter
---
$\lambda$  
Thermal conductivity of soil-water matrix
---
$\rho$  
Density of soil-water matrix
---
$\rho_w$  
Density of water
---
$\varphi$  
Rate of increase in surface temperature due to climate change
Abbreviated Units Used in Dissertation

mm, millimetre
cm, centimetre
m, metre
km, kilometre
s, second
min, minute
yr, year
g, gram
kg, kilogram
°C, degrees Celsius
J, joule
W, watt
Pa, pascal
CHAPTER 1: Introduction to Salmonid Thermal Refugia

1.1 Thermal Refugia and Climate Change

Many cold-water fishes, such as salmonids, are highly sensitive to water temperature variability (Johansson et al. 2009, Lund et al. 2002). A rise in river water temperature can increase their basal metabolic rate (Beauregard et al. 2013, Berman and Quinn 1991), lower oxygen solubility (Hayashi and Rosenberry 2002), and harm their sexual development (Pankhurst and King 2010). Thus, water temperature is an important factor that regulates salmonid populations in fresh water (Moore et al. 2012). Water temperatures are driven by spatiotemporally varying and diverse thermal inputs, including shortwave and longwave radiation, convective exchanges between the lower atmosphere and the water surface, and conductive and advective exchanges with surrounding sediment (Bogan et al. 2003, Caissie 2006a, Webb et al. 2008).

Discrete cold-water plumes within the mainstem of a river (i.e., riverine thermal diversity) can provide thermally suitable habitat for cold-water fishes when ambient water temperatures exceed critical thresholds (Ebersole et al. 2003a, Gibson 1966, Huntsman 1942, Sutton et al. 2007, Torgersen et al. 2012, Wilbur 2012). A proper understanding of the principal components of a fluvial system is required to recognize the factors and drivers of riverine thermal diversity.
Figure 1.1. Principal components of a fluvial system (after Alexander 2006, Findlay 1995, Poole and Berman 2001).

Essentially, a fluvial system can be broken up into three regions that continuously interact with each other: the riparian zone, the alluvial aquifer, and the fluvial channel (Figure 1.1, Hayashi and Rosenberry 2002, Poole and Berman 2001, Woessner 2000). The riparian zone acts as the transition between the terrestrial and riverine ecosystems. It typically hosts a variety of plant and animal life, protects the river from sediment runoff, and limits absorbed solar radiation (Allan 2004, Gregory et al. 1991, Hayashi and Rosenberry 2002, Tockner and Stanford 2002).

Groundwater from surrounding unconfined, alluvial aquifers continuously interacts with the fluvial channel. Hyporheic flow refers to hydraulic mixing at the alluvium-river interface and is typically differentiated from far-field groundwater.

elevations of their distributions, and reliance on available refugia is projected to increase in a warming climate (Brewer 2013, Mackenzie-Grieve and Post 2006).

The principal characteristics and generating mechanisms of thermal refugia (Figure 1.2) have been identified and categorized in previous studies (e.g., Chu et al. 2010, Ebersole et al. 2003a, Kaeding 1996, Nielsen et al. 1994, Olsen and Young 2009, Torgersen et al. 1999, 2012). Thermal refugia can be detected via airborne thermal infrared sensing (Deitchman and Loheide 2012, Dugdale et al. 2013, Torgersen et al. 2001) or in-channel temperature surveys. Breau et al. (2007) classified high stream temperature (i.e., when juvenile Atlantic salmon seek thermal refuge) as > 23°C and identified cold-water plumes with thermal differences from the ambient river temperature up to 11°C. Ebersole et al. (2003b) identified cool patches serving as thermal refugia as any region with a temperature at least 3°C less than the ambient river temperature, a surface area of at least 0.5 m², and a dissolved oxygen level of at least 3 ppm. They found that the thermal heterogeneities comprised 2% of the channel surface area and tended to be long and thin (median length to width ratio = 5.5).

Bilby (1984) classified thermal refugia as groundwater seeps, cold-water tributaries, emerging stream bed flow (hyporheic water), deep water impoundments, or shading. Ebersole et al. (2003b) characterized cold-water patches as cold alcoves (emergent floodplain/gravel bar groundwater), floodplain springbrooks, cold side channels, and lateral seeps. Most of these classifications of thermal refugia are groundwater-sourced, and many lotic ecology texts now emphasize the importance of discrete groundwater discharge for the generation of riverine thermal refugia (e.g., Allen and Castillo 2007, Cushing and Allan 2001, Jones and Mulholland 2000).
Figure 1.2. An overview of mechanisms and channel morphologies that induce thermal anomalies and thereby create suitable thermal refugia. Darker blue colors indicate colder water. An estimation of the maximum temperature difference between the temperature of a particular refugium and the ambient water temperature is given in brackets. These estimates are derived from other literature sources (Ebersole et al. 2003b, Nielsen et al. 1994) and extensive aerial infrared images and in-stream thermal surveys of the Little Southwest Miramichi River and other branches of the Miramichi River (e.g., Wilbur 2012).
Possible future climate scenarios are generated by performing simulations of atmospheric, oceanic, and surficial processes in global climate models (GCMs, Meehl et al. 2007, Solomon et al. 2007). There is international concern that projected future climate conditions will reduce available habitat for cold-water fishes in already threatened river systems. Many North American rivers have already exhibited decadal warming trends in seasonal or mean annual river temperature (e.g., Isaak et al. 2012, Kaushal et al. 2010, Swansburg et al. 2004), and changing river thermal regimes have already been linked to altered migratory patterns of salmonids (Kennedy and Crozier 2010). Numerous researchers have considered future climate change-induced increases in river temperature and the associated reduction in cold-water fish habitat (e.g., Chu et al. 2005, Eaton and Scheller 1996, Elliott and Elliott 2010, Ficke et al. 2007, Jones et al. 2014, Jonsson and Jonsson 2009, Mayer 2012, Meisner 1990, Mohseni et al. 2003, Moore et al. 2013, Schindler 2001, Sinokrot et al. 1995, van Vliet et al. 2011, Wu et al. 2012). These studies have generally indicated that cold-water fish species will experience a significant loss of thermally suitable habitat, particularly under the most extreme climate warming scenarios.

Models that investigate future trends in river temperatures should also consider trends in future groundwater temperature because thermal regimes of aquifers and hydraulically connected rivers are interrelated. However, previous studies that have simulated future river temperature response to climate change have either ignored changes to groundwater temperature or employed seemingly unrealistic simplifications, such as assuming decadal changes in groundwater temperature would track changes in air temperature. Furthermore, the sensitivity of the thermal regimes of shallow
groundwater and cold-water refugia to projected increases in air temperature has not been previously investigated. However, future states of thermal refugia are of critical importance and must therefore be considered in an effective river management strategy for cold-water fishes, particularly at the southern (latitude) or lower (elevation) limits of their distribution. For example, Berman and Quinn (1991) noted that the ‘the availability of suitable thermal refuges and appropriate holding habitat within mainstem rivers may affect long-term population survival’.

A number of researchers have acknowledged the lack of detailed analyses of the influence of climate change on shallow groundwater and the associated impact to fish habitat. Mohseni et al. (2003) called for a more thorough hydrogeological model of thermal refugia under the influence of climate warming. Chu et al. (2008) stated ‘climate change induced differences in precipitation and temperature that may influence the magnitude and timing of groundwater discharge should be addressed in future analyses’. Tanaka (2007) suggested that future research should investigate the hydrogeological processes that generate thermal refugia. Mayer (2012) observed that the lack of information regarding the future thermal state of groundwater posed a challenge to river temperature analysts. Recently, Kanno et al. (2013) noted that the ‘spatial variability in resiliency of groundwater temperature to air temperature is the missing piece to assess climate change impacts on headwater stream fish accurately’. Thus, the original research presented in this dissertation, which focuses on the hydraulic and thermal influences of climate change on shallow aquifers, helps to address knowledge gaps that have been identified by numerous researchers.
1.2 Study Site

1.2.1 General site: Miramichi River, New Brunswick

The Miramichi River is the second largest river in the Canadian Maritime provinces with a drainage basin of 14,000 km$^2$ (Caissie and El-Jabi 1995, Cunjak and Newbury 2005). It flows northeast from its headwaters in central New Brunswick and discharges into Miramichi Bay (Figure 1.3). Technically, the Miramichi River only begins downstream of the confluence of the Main Southwest Miramichi River and the Northwest Miramichi River (Marriner 1997), but herein ‘Miramichi River’ refers to the entire Miramichi River system, including its six major tributaries—Bartibog River, Napan River, Bay du Vin River, Black River, Northwest Miramichi River, and Main Southwest Miramichi River. The entire Miramichi River is unregulated and has been identified as the largest producer of Atlantic salmon (Salmo salar) in North America (Caissie et al. 2007, Cunjak et al. 2005). The Atlantic salmon is an anadromous fish that is coveted by anglers for its size and elusive nature (e.g., Wulff 1988), thus the Miramichi River is world-renowned angling destination.

Atlantic salmon are vital to the economic, cultural and ecological wellbeing of Atlantic Canada and, in particular, New Brunswick (Lund et al. 2002, Parenteau 1998, Whoriskey and Glebe 1998). For example, the total spending for wild Atlantic salmon related activities in Atlantic Canadian rivers in 2010 was estimated to be $166 million (Gardner Pinfold Consultants Inc. 2011). Unfortunately, the Atlantic salmon population is already in decline in New Brunswick compared to historical or recent numbers (Department of Fisheries and Oceans 2013). Atlantic salmon have been identified as an endangered species in the Bay of Fundy, and temperatures stress is one postulated reason
for the depletion in the stocks (Lund et al. 2002). Warming trends have been observed in portions of the Miramichi River basin (Swansburg et al. 2004), and these are expected to intensify in the coming decades (Huard 2011). These observed and projected temperature trends are generating concern regarding the future state of Atlantic salmon in the province New Brunswick.

Figure 1.3. A map of the Miramichi River and its location in New Brunswick, Canada (spatial data from, NBADW 2013).
1.2.2 Specific site: Little Southwest Miramichi River

The Little Southwest Miramichi River (LSW) is a sixth order river branch of the Miramichi River (Figures 1.3 and 1.4) that flows from the Christmas Mountains in central New Brunswick and eventually joins the Northwest Miramichi River near the community of Red Bank (NBADW 2013). The LSW has a mean annual flow of 32.5 m$^3$ s$^{-1}$ and drains a surface area of 1340 km$^2$ (Caissie et al. 2006b). The LSW catchment is primarily forested with a mixed coniferous (65%) and deciduous (35%) canopy (Cunjak et al. 1990). The region experiences a humid-continental climate characterized by arid, cold winters (Cunjak et al. 1993). The LSW has an average width and depth of 80 m and 0.55 m respectively, and is thus classified as a wide, shallow river (Caissie et al. 2006b). This fluvial geomorphology induces high thermal sensitivity to solar radiation (Caissie 2006a). Daily maximum LSW water temperatures have been previously documented in the range of 25-30°C (Breau et al. 2007, Caissie et al. 2007), which is well above the thermal preference (e.g., 22°C) of juvenile Atlantic salmon (Elliott and Elliott 2010, Johansson et al. 2009). Thus, thermal refugia within the LSW are critical for sustaining salmonid populations during high temperature events (Breau et al. 2011).
Watershed Sciences Inc. (2009) conducted an aerial infrared thermal survey of the LSW and identified numerous thermal anomalies. After examining the infrared imagery and conducting site investigations, two thermal refugia sites were selected for the present study (Figure 1.5). The first refugium selected for analysis is the cold-water plume formed at the mouth of Otter Brook. Otter Brook is a second-order tributary of the LSW that is predominantly sourced by groundwater discharge from the surrounding highly permeable glaciofluvial deposits (Allard 2008, Lavergne and Hunter 1982). The cold plume at the confluence of Otter Brook and the LSW continues downstream about 200 m along the south bank of the LSW (Figure 1.5). The other refugium is a lateral
groundwater seep (Figure 1.5), which is sourced by a coarse, glacial deposit on the north side of the LSW. This refugium is smaller and typically colder than the one at the mouth of Otter Brook.

Figure 1.5. Thermal infrared imagery of the LSW and the thermal refugia at the groundwater seep and the mouth of Otter Brook (spatial data from Wilbur and Curry 2011).
Figure 1.6. Images of (a) Otter Brook, (b) the groundwater (GW) seep, and the associated conceptual models for these refugia (c and d). $Q$ represents the river or groundwater discharge, and $T$ represents the river or groundwater temperature.

Both refugia are aquifer-sourced, and thus can be studied via a numerical model of groundwater flow and heat transport. However, these refugia also exhibit disparate features. Most notably, the Otter Brook refugium is sourced by indirect groundwater discharge (i.e., groundwater discharge to the brook, which then transports the water to the LSW), whereas the groundwater seep is sourced by direct groundwater seepage (Figure 1.6). Furthermore, due to the steep bank on the north side of the LSW (Figure 1.6b), groundwater discharge to the lateral groundwater seep is at a much greater depth.
relative to the ground surface than groundwater discharge to Otter Brook. This greater depth results in groundwater seep temperatures that are characterized by low seasonal variability. As discussed in detail in Chapter 5, this has profound implications for the resilience of this refugium to atmospheric climate change.

The summer of 2010 was unusually warm, and maximum daily LSW water temperature reached 31°C, which exceeds the 1000 minute upper lethal temperature limit (29.5°C) for Atlantic salmon parr (Lightfoot and Solomon 2008). During this high temperature event, approximately 6000-10,000 fish were observed to be simultaneously aggregating in the Otter Brook refugium (Figure 1.7, T. Linnansaari, UNB, pers. communication, 2013). It should be noted that although this research focuses on Atlantic salmon, brook trout (Salvelinus fontinalis) are also an important New Brunswick salmonid species that utilize thermal refugia within the Miramichi River system (Curry and Gautreau 2010, Wilbur 2012).

Figure 1.7. Photograph of an Atlantic salmon parr aggregation at the confluence of Otter Brook and the LSW (July 2010). Photo credit. Rick Cunjak’s lab, Stable Isotopes in Nature Laboratory, UNB Fredericton and Canadian Rivers Institute.
1.2.3 Thermal characterization of Otter Brook and the groundwater seep

To gain a preliminary understanding of the thermal dynamics of the refugia considered in this study and to inform the research described in subsequent chapters, VEMCO Minilog J-Series temperature loggers (VEMCO Limited, Nova Scotia) were installed at the mouth and east branch of Otter Brook and within the groundwater seep. Figure 1.8 illustrates the differences between the air temperature and the water temperature in the LSW, Otter Brook and groundwater seep in the spring-fall of 2012. Clearly the seep temperature is characterized by more damping and lagging in response to air temperature variations than the temperature of Otter Brook or the LSW. For example, the air or surface water temperatures peaked in July or August (day of year, DOY, 182-243), whereas, the seep temperatures were still increasing on the date the loggers were removed (DOY 270). Also, the seep temperature series does not exhibit the same daily fluctuations that characterize the other series. Differences between the seep temperature and the LSW water temperature approached 20°C on several occasions in 2010 (Figure 1.8). The temperature at the mouth of Otter Brook can also be much colder (up to 10°C) than the temperature of the LSW, which closely follows the air temperature (Figure 1.8). This extreme temperature difference is predominantly caused by the baseflow dominance and increased shading of Otter Brook and provides the mechanism for creating a large thermal refugium at this confluence.

Two longitudinal temperature surveys were conducted within Otter Brook to better characterize the thermal dynamics of this biologically significant tributary. The thermal surveys began at the mouth of Otter Brook and continued upstream along the east branch of the brook (Figure 1.9b) until the brook became hydraulically stagnant and
difficult to transverse due to dense cover. The brook temperature at the surface of the channel bed was recorded every 10 m using a YSI 85 meter (YSI Inc, Ohio, USA).

![Graph showing temperature data](image)

**Figure 1.8.** Air temperature and water temperature recorded in Otter Brook, the groundwater seep, and the LSW versus the day of the year for the spring-fall of 2012. The air temperature data were measured at the Environment Canada weather station in Miramichi (EC, 2012a), and the LSW river temperature data were provided by Emily Corey (Department of Biology, University of New Brunswick, Fredericton).

The mainstem of the brook (i.e., downstream of the confluence of the east and west branches, Figure 1.9) exhibits a relatively linear cooling trend toward the mouth, which is suggestive of spatially uniform groundwater discharge (Somers, 2013 and Figure 1.9a). At the confluence, rapid warming occurs in a deep stagnant pool exposed to solar radiation. The minimum water temperature occurs at the headwater (e.g., 1550 m upstream of the mouth), which is likely sourced by regional groundwater discharge. The information gained from the field analyses of the groundwater seep and Otter Brook helped form the basis for the modeling described in subsequent chapters.
Figure 1.9. (a) Otter Brook temperature vs. the distance upstream of the mouth for two thermal surveys, and (b) the location of the thermal surveys indicated on a map of the Otter Brook catchment. The temperature legend in (b) refers to the survey conducted on June 29, 2010. The green (Eastern Lowlands) and brown (Valley Lowlands) backgrounds in (b) indicate two separate New Brunswick ecoregions.

1.3 Objectives

The goal of this research is to explore the potential influence of climate change on aquifer thermal and hydraulic regimes and to apply this knowledge to quantitatively assess the sensitivity of cold-water thermal refugia to future climate change. The primary goal is accomplished via the following objectives:

1) Apply a daily soil water balance model to investigate the influence of climate change on surface and near-surface hydrological processes and to quantify the uncertainty when simulating future groundwater recharge (Chapter 2);

2) Improve on existing analytical solutions to the transient, one-dimensional conduction-advection equation by utilizing initial and boundary condition functions that more accurately reflect measured and projected data (Chapter 3);
3) Employ a validated, process-oriented model of surface energy transfer to investigate the role of climate change and associated snowpack evolution in altering future surface thermal regimes (Chapter 4);

4) Develop an empirical relationship between surface temperatures and depth-dependent groundwater temperatures using measured surface and groundwater temperature data, and apply the results from (3) in conjunction with this empirical relationship to predict future groundwater temperature trends (Chapter 4);

5) Perform numerical simulations of coupled groundwater flow and energy transport to investigate the impact of changing groundwater recharge and ground surface temperature on the timing, magnitude, and temperature of groundwater discharge (Chapter 5).

1.4 Dissertation Description

This dissertation adheres to the article format with each of the chapters representing a self-contained article with its own independent introduction, methods, results, discussion, and conclusion. Chapter 2 describes the modeling of future groundwater recharge under multiple downscaled climate scenarios for the Little Southwest Miramichi River catchment. Changes to groundwater recharge can alter advective heat transport and thereby impact subsurface thermal regimes. In particular, this section focuses on the uncertainty in forming predictions regarding the future state of water resources. This chapter was recently published in the Journal of Hydrology. Chapter 3 details the development and application of a new analytical solution to the one-dimensional conduction-advection equation that accommodates non-linear boundary and
initial conditions. These modifications improve the fidelity of the initial and boundary conditions to measured temperature-depth profiles and projected air temperature trends. This chapter, which has been published in *Hydrological Processes*, does not discuss thermal refugia in detail, but it does add to the body of knowledge regarding the future state of groundwater temperature.

Chapter 4 addresses the future thermal state of surface and subsurface thermal regimes as a result of atmospheric climate change. Future ground surface temperature is simulated via a process-oriented surface energy transfer model. The simulations elucidate whether ground surface temperature will track changes in air temperature on a seasonal or mean annual basis. This chapter also describes the development and application of an empirically-based ground surface temperature-groundwater temperature transfer function. The results have been published in *Hydrology and Earth System Sciences*. Chapter 5 describes the application of a state of the art finite element model of groundwater flow and energy transport to examine the impacts of climate change on the timing, magnitude, and temperature of groundwater discharge. The sensitivity of groundwater-sourced thermal refugia to climate change and the associated implications for cold-water fish habitat are discussed. These results have been published in *Water Resources Research*. Chapter 6 discusses management strategies for thermal refugia based on the simulation results detailed in previous chapters. Finally, Chapter 7 restates the primary results of this research and the major scientific conclusions. The Candidate was the primary author for each of these journal submissions, and each coauthor provided permission to include these submissions in this dissertation.
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CHAPTER 2: The Uncertainty Associated With Estimating Future Groundwater Recharge

Abstract

Global climate models (GCMs) project significant changes to regional and globally-averaged precipitation and air temperature, and these changes will likely have an associated impact on groundwater recharge. A common approach in recent climate change-impact studies is to employ multiple downscaled climate change scenarios to drive a hydrological model and project an envelope of recharge possibilities. However, each step in this process introduces variability into the hydrological results, which translates to uncertainty in the future state of groundwater resources. In this contribution, seven downscaled future climate scenarios for a northern humid-continental climate in eastern Canada were generated from selected combinations of GCMs, emission scenarios, and downscaling approaches. Meteorological data from the climate scenarios and field data from a small unconfined aquifer were used to estimate groundwater recharge with the soil water balance model HELP3. HELP3 simulations for the period 2046–2065 indicated that projected recharge was most sensitive to the selected downscaling/debiasing algorithm and GCM. Projected changes in average annual recharge varied from an increase of 58% to a decrease of 6% relative to the 1961–2000 reference period. Such a large range in projected recharge provides very little useful information regarding the future state of groundwater resources. Additional results from recent comparable studies are compiled and discussed. Based on the results obtained from

the present case study and the other studies reviewed, the limitations of current approaches for projecting future recharge are identified, and several suggestions for research opportunities to advance this field are offered.

**Keywords**: groundwater recharge, climate change, climate scenarios, downscaling, snowmelt, uncertainty

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### 2.1 Introduction

Climate change has resulted in increases in globally-averaged mean annual air temperature and variations in regional precipitation, and these changes are expected to continue and intensify in the future (Solomon et al. 2007). Projected climate data are generated by simulating global atmospheric, oceanic, and surficial processes in global climate models (GCMs), which are driven by emission scenarios that require forecasts of future population growth and technology (Nakicenovic and Swart 2000). GCM simulations are performed using coarse computational grids, and the results should be downscaled to produce local climate conditions that may subsequently be used for hydrology applications (Wilby and Wigley 1997, Wilby et al. 2000).

The impact of climate change on the quantity and quality of groundwater resources is of global importance because 1.5 to 3 billion people rely on groundwater as a drinking water source (Kundzewicz and Döll 2009). Despite the importance of the relationship between climate conditions and groundwater reserves (Taylor et al. 2012), research examining the effects of future climate change on groundwater has lagged corresponding research for surface water resources (Green et al. 2011). The IPCC Fourth

Recently there has been a discernible shift in the approaches used to examine climate change impacts on groundwater recharge. Rather than simulating changes for a single climate scenario, researchers have been employing multiple climate change scenarios generated from a variety of methods to produce a range, or envelope, of projected changes in recharge. Holman et al. (2012) suggested that the best practice for using climate model projections to assess the impact on groundwater was to ‘use climate scenarios from multiple GCM or RCMs [regional climate models]. . .use multiple emission scenarios. . .[and] consider the implications of the choice of the downscaling method’. This approach introduces additional variability in the climate data, which translates into uncertainty in future groundwater recharge. For example, when more than 10 GCMs were employed for projecting future precipitation, it was found that less than 80% of the GCMs agreed ‘in whether annual precipitation will increase or decrease’ in
most regions other than at high northern latitudes and in the Mediterranean region (Döll 2009). The majority of uncertainty in the projected climate data (and consequently in the projected recharge) appears to stem from the selection of the GCM (Kay et al. 2009), although other factors, such as the emission scenarios, downscaling methods, or the hydrological model, can also contribute uncertainty (Crosbie et al. 2011b, Holman et al. 2009, Rowell 2006).

Several recent groundwater recharge studies employing multiple climate change scenarios have been conducted at a very large scale. Döll (2009) simulated the vulnerability of groundwater to climatic change at the global scale using the hydrology model WaterGAP driven by climate data from two GCMs and two emission scenarios, and concluded that the uncertainty in projected precipitation from the climate scenarios resulted in uncertainty in recharge estimates which was spatially heterogeneous (e.g., see Australia, Figure 1, Döll, 2009). Crosbie et al. (2013) simulated the changes in recharge for a 2050 climate for the entire continent of Australia using climate data from 16 GCMs and three emission scenarios to drive the WAVES hydrological model. Their study indicated that the range of projected changes in recharge was large and spatially variable. They also found that it was generally difficult to project the magnitude or even direction of future recharge changes, although in certain regions of southern Australia, all 48 climate variants projected a decrease in recharge.

Many more regional scale studies have been conducted to investigate the link between climate change and groundwater recharge. For example, Serrat-Capdevila et al. (2007) used climate data for the San Pedro Basin from 17 GCMs to estimate recharge from a simple empirical equation. In the case of the drier climate projections,
simulations indicated that groundwater recharge could cease completely. Holman et al. (2009) simulated future groundwater recharge using one GCM, two emission scenarios, and two downscaling methods (a stochastic weather generator and the change factor method) and found that the uncertainty due to the downscaling method was greater than the uncertainty associated with the emission scenario. Allen et al. (2010) used climate data from four GCMs, one emission scenario, and one downscaling algorithm to drive simulations within a hydrology model of the Abbotsford-Sumas aquifer. Crosbie et al. (2011b) simulated groundwater recharge changes at three locations in southern Australia using multiple GCMs, downscaling methods, and hydrology models and found that the highest uncertainty in modeling future recharge arose from the selection of the GCM. Dams et al. (2012) used 28 climate scenarios to simulate a range of changes in mean annual recharge for a catchment in Belgium. Table 2.1 gives a summary of the results from these and other recent groundwater recharge studies.

The purpose of this contribution is to provide a case study that adds to the recent body of literature by examining the uncertainty in projected recharge for a humid-continental climate in which snow accumulation and melt are important factors affecting groundwater recharge. Seven climate scenarios generated from multiple (1) GCMs, (2) emission scenarios, and (3) downscaling/debiasing methods are utilized to drive simulations of future (2046–2065) groundwater recharge for a small, shallow, unconfined aquifer in central New Brunswick, Canada. Others (e.g., Jackson et al. 2011, Serrat-Capdevila et al. 2007) have examined the uncertainty in groundwater recharge due to varying one or two of the climate modeling options noted above, but this is the first contribution to examine the effect of varying all three following the recommendations of
Holman et al. (2012). The uncertainty in recharge projections obtained in this study is also compared to the uncertainties reported in several recent groundwater recharge studies. Recommendations for future research opportunities are suggested based on the results obtained from the present case study and the studies summarized in Table 2.1.

Table 2.1. An overview of several recent studies that have employed multiple climate scenarios to examine the impact of climate change on groundwater recharge

<table>
<thead>
<tr>
<th>Study Reference</th>
<th>Number of GCMs</th>
<th>Number of ES(^1)</th>
<th>Number of DM(^2)</th>
<th>Scale of Studies</th>
<th>Max Changes in Avg. Recharge(^3)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Serrat-Capdevila et al. (2007)</td>
<td>17</td>
<td>4</td>
<td>1</td>
<td>Regional</td>
<td>-100% to ~+35%</td>
</tr>
<tr>
<td>Döll (2009)</td>
<td>2</td>
<td>2</td>
<td>NA</td>
<td>Global</td>
<td>~30 to +100%(^4)</td>
</tr>
<tr>
<td>Holman et al. (2009)</td>
<td>1</td>
<td>2</td>
<td>2</td>
<td>Regional</td>
<td>-14 to -37%(^5)</td>
</tr>
<tr>
<td>Allen et al. (2010)</td>
<td>4</td>
<td>1</td>
<td>1</td>
<td>Regional</td>
<td>-1.5 to +23%(^6)</td>
</tr>
<tr>
<td>Crosbie et al. (2010)</td>
<td>15</td>
<td>3</td>
<td>1</td>
<td>Regional</td>
<td>&lt;-50 to &gt;+50%</td>
</tr>
<tr>
<td>Crosbie et al. (2011b)</td>
<td>5</td>
<td>1</td>
<td>3</td>
<td>Regional</td>
<td>-83 to +447%</td>
</tr>
<tr>
<td>Jackson et al. (2011)</td>
<td>13</td>
<td>1</td>
<td>1</td>
<td>Regional</td>
<td>-26 to +31%</td>
</tr>
<tr>
<td>Dams et al. (2012)</td>
<td>5</td>
<td>2</td>
<td>1</td>
<td>Regional</td>
<td>-20 to +7%</td>
</tr>
<tr>
<td>Ali et al. (2012)</td>
<td>15</td>
<td>3</td>
<td>1</td>
<td>Regional</td>
<td>-33% to +28%(^7)</td>
</tr>
<tr>
<td>Crosbie et al. (2013)</td>
<td>16</td>
<td>3</td>
<td>1</td>
<td>Continental</td>
<td>+45% to +283%(^8)</td>
</tr>
</tbody>
</table>

\(^1\) ES= emission scenarios (A1F1, A2, A1B, B1, etc.).
\(^2\) DM= downscaling methods.
\(^3\) For studies with multiple locations this column lists the results from the locations with the highest uncertainty.
\(^4\) Estimated from the southwestern Australian region in Figure 1 of Döll (2009).
\(^5\) Taken from Table 1 of Holman et al. (2009) for loamy soil and the 2050’s climate scenarios.
\(^6\) Allen et al. (2010) have a discrepancy between the reported recharge ranges in their abstract (-10.5% to +23.2%) and in their Figure 13 (-1.5% to +23.2%). The text in the results seems to suggest that the latter is correct.
\(^7\) Taken from Table A1 of Ali et al. (2012), these results were from the Southern Perth Basin for the wet and dry simulations compared to the recent recharge.
\(^8\) Taken from Appendix C of Crosbie et al. (2011a), these results were for Brunswick Coastal Sands for the median dry climate and the median wet climate.

2.2 Methods

The approach for estimating recharge was based on techniques similar to those recently employed by others (e.g., Jackson et al. 2011, Jyrkama and Sykes 2007, Scibek and Allen 2006b, Toews and Allen 2009b). We first obtained an array of future climate
projections that were developed using several documented and established techniques. These climate scenarios were selected because they span the range of plausible future climatic conditions for the study location. The observed climate data and the projected climate series were then used to drive a simple water balance hydrology model to simulate historic and future groundwater recharge. In general, a parsimonious hydrological modeling approach was employed. For example, although it is known that increased CO$_2$ concentrations will affect canopy density and evapotranspiration and thereby impact groundwater recharge (Ficklin et al. 2010, Green et al. 2007), like many previous studies, the biophysical parameters (e.g., maximum leaf area index) were assumed to be temporally invariant. This approach isolates the effect of the driving climate data on groundwater recharge.

The approach outlined above was used to investigate the inherent uncertainty involved when using climate projections to drive simulations of groundwater recharge due to the uncertainty arising from the selection of the (1) GCM, (2) downscaling method, and (3) emission scenario. In this work, the uncertainty in projected groundwater recharge is defined as the magnitude of the range in changes to the projected mean annual groundwater recharge. The projected changes in mean annual groundwater recharge are quantified as the % difference from the simulation for the reference period (1961–2000). Uncertainties arising from the selection of (1), (2), and (3) are propagated through the climate and groundwater recharge modeling processes. However, following the approach of others who have demonstrated the uncertainty in groundwater recharge projections (e.g., Crosbie et al. 2011b, Jackson et al. 2011), we have not attempted to conduct a formal uncertainty propagation analysis for each step in the modeling process. Rather, the
uncertainty arising from the selection of (1), (2), and (3) was investigated by holding two climate simulation approaches constant while varying the third. For example, the effect of the downscaling algorithm was investigated by examining the difference in the resultant climate data and simulated recharge when the GCM and emission scenario were identical in two runs, but the downscaling method was varied.

2.2.1 Geographical setting

The geographic location for our simulations is the Otter Brook catchment in central New Brunswick, Canada (N46 52 W66 02). Otter Brook is a second order tributary of the Little Southwest Miramichi River (Figure 2.1) that is predominately fed by groundwater baseflow. Lavergne and Hunter (1982) mapped the surficial geology surrounding Otter Brook and determined that the brook lies within a glaciofluvial outwash deposit that mainly consists of sand and gravel. In a more recent investigation, multiple test holes were excavated within the Otter Brook catchment. Aerial photography and samples from these test holes indicate that the Otter Brook deposit is primarily composed of glaciofluvial outwash sediments varying from cross-bedded sand to thick-bedded coarse gravel (Allard 2008). The results from a ground penetrating radar survey indicated that the groundwater table is at a depth of about 7.5 m and that the surficial sand and gravel deposit is approximately 10 m thick (Allard 2008). The Otter Brook catchment has a land surface cover similar to the surrounding region, which is forested with a coniferous (65%) and deciduous (35%) canopy (Cunjak et al. 1990). The annual precipitation in the region is 1230 mm; with approximately 33% falling as snow (EC 2013a). The region experiences a humid-continental climate characterized by arid, cold
winters (Cunjak et al. 1993). This particular catchment was selected because it is part of a study area in which climate-induced thermal and hydrologic changes to salmonid habitat are being investigated.

Figure 2.1. The location of the Otter Brook catchment within the province of New Brunswick, Canada (data from, NBADW 2013).

2.2.2 Emission scenarios, downscaling algorithms, and GCMs

The three emission scenarios that we have utilized in this study are B1, A1B, and A2 (Nakicenovic and Swart 2000). Climate simulations driven by emission scenario A2 typically project more pronounced climatic changes than those driven by emission
scenarios A1B or B1; however, the relative effects of each emission scenario may not be
realized for several decades.

Downscaling approaches have been thoroughly reviewed in scientific literature
(e.g., Maraun et al. 2010, Wilby and Wigley 1997, Xu 1999). A simple downscaling
approach is the daily translation (DT) method, which is in the family of ‘statistical’ or
‘quantile–quantile mapping’ downscaling techniques (Teutschbein and Seibert 2012). In
the DT method, a GCM is initially run for a reference period containing local
observations. Scaling factors for precipitation and AT are then determined from the
distributions of the reference period simulation and the local observations using empirical
cumulative distribution functions. GCM simulations for a future time period/emission
scenario are then downscaled by applying the scaling factors. This approach differs from
the often criticized delta method by adopting variable scaling factors (Huard 2011).

Many more complex statistical downscaling methods have been developed; one of
these is the hybrid multivariate linear regression (HMLR) model (Jeong et al. 2012a,
Jeong et al. 2012b). Regression-based statistical downscaling techniques are predicated
on the assumption that local climate conditions can be determined from large-scale
climate variables using linear or non-linear transfer functions (Jeong et al. 2012a).
Because regression-based methods often have difficulty producing the observed
variability in local climate predictands, a stochastic generator is used to increase the
variance in the datasets. In the HMLR method, local climate variables are obtained from
GCM simulations using multiple regression functions determined from reference period
simulations.
The output from GCMs can also be dynamically downscaled by performing simulations with finer resolution RCMs, which are driven at the boundaries by the results from GCMs. However, RCMs tend to introduce additional biases, thus the results are often further debiased/downscaled by comparing simulations for a reference period to observations to determine scaling factors, and then making corrections to the generated dataset for a future period. For the present study, additional debiasing/downscaling was performed on the RCM climate series using the DT method (Huard 2011). Debiasing and downscaling are often collectively referred to as ‘post-processing’.

The HMLR downscaled climate data were contributed by the Université du Québec Institut National de la Recherche Scientifique (INRS) (D. Jeong, personal communication), while the other climate data series were produced from the third Coupled Model Intercomparison Project database of GCM output (CMIP3, Meehl et al. 2007) and dynamically downscaled using the Canadian Regional Climate Model (CRCM4.2.3, de Elia et al. 2008, Huard 2011) or statistically downscaled with the DT method (Huard 2011). In total, seven projected climate scenarios (Table 2.2) were produced for the period of 2046–2065 using five GCMs, two downscaling methods, and three emission scenarios. These climate data provide the basis for projecting future groundwater recharge.
Table 2.2. Details for the climate simulations utilized in this study

<table>
<thead>
<tr>
<th>Emission Scenario</th>
<th>Model Type</th>
<th>Model Name</th>
<th>GCM Driver</th>
<th>ID -</th>
<th>Post-processing Method</th>
<th>Contributor Organization</th>
</tr>
</thead>
<tbody>
<tr>
<td>A2</td>
<td>GCM</td>
<td>CGCM3</td>
<td>-</td>
<td>-</td>
<td>Statistical-HMLR</td>
<td>INRS (Jeong et al. 2012b)</td>
</tr>
<tr>
<td>A2</td>
<td>RCM</td>
<td>CRCM 4.2.3</td>
<td>CGCM3</td>
<td>Aev</td>
<td>Dynamical</td>
<td>Ouranos (Huard 2011)</td>
</tr>
<tr>
<td>A2</td>
<td>RCM</td>
<td>CRCM 4.2.3</td>
<td>Echam5</td>
<td>Agx</td>
<td>Dynamical</td>
<td>Ouranos (Huard 2011)</td>
</tr>
<tr>
<td>B1</td>
<td>GCM</td>
<td>CSIRO Mk3.0</td>
<td>-</td>
<td>-</td>
<td>Statistical-DT</td>
<td>Ouranos (Huard 2011)</td>
</tr>
<tr>
<td>B1</td>
<td>GCM</td>
<td>CSIRO Mk3.5</td>
<td>-</td>
<td>-</td>
<td>Statistical-DT</td>
<td>Ouranos (Huard 2011)</td>
</tr>
<tr>
<td>A1B</td>
<td>GCM</td>
<td>Miroc3.2 Hires</td>
<td>-</td>
<td>-</td>
<td>Statistical-DT</td>
<td>Ouranos (Huard 2011)</td>
</tr>
<tr>
<td>A1B</td>
<td>GCM</td>
<td>CGCM3</td>
<td>-</td>
<td>-</td>
<td>Statistical-HMLR</td>
<td>INRS (Jeong et al. 2012b)</td>
</tr>
</tbody>
</table>

Figure 2.2 shows the projected changes in mean annual precipitation and air temperature for each of the seven combinations given in Table 2.2 compared to reference period climate data obtained from Environment Canada (EC 2013b). All of the scenarios project an increase in air temperature (range of 0.4–3.9°C), but the projections for precipitation vary significantly in magnitude and direction (-12% to +49%).

Figure 2.2. Projected changes in mean annual air temperature and precipitation for the Otter Brook catchment for the period 2046-2065 (data provided by D. Huard of Ouranos and D. Jeong of INRS).
2.2.3 The hydrology model: HELP3

Downscaled climate data can be utilized to drive simulations within hydrology models. Kingston and Taylor (2010) determined that the selection of the GCM yielded far more uncertainty in their climate-hydrology simulations than their hydrological model parameterization. Crosbie et al. (2011b) found that the selection of the hydrology model contributed less uncertainty to recharge estimations than did the choice of the downscaling scenario or the GCM. Teng et al. (2012) simulated the impact of climate change on runoff and also found that the selection of the GCM contributed far more uncertainty to the hydrological simulation results than the selection of the hydrological model. In light of these three recent studies, only one hydrological model was employed for the present study.

We performed daily point simulations of recharge using the soil water balance hydrology model HELP3 (Hydrologic Evaluation of Landfill Performance, version 3), which simulates surface and shallow subsurface processes, including snow storage, snowmelt, interception, infiltration, runoff, evaporation, transpiration, and drainage (Schroeder et al. 1994). HELP3 is a one-dimensional model that simplifies horizontal lateral flow and the interactions between the hydrologic processes. For example, HELP3 does not allow water to be rerouted upwards once it has passed the evaporative zone depth (EZD). For the purpose of this study, water passing the EZD is assumed to result in groundwater recharge. HELP3 has been used in several recent studies (e.g., Allen et al. 2010, Crosbie et al. 2011b, Jyrkama and Sykes 2007, Liggett and Allen 2010, Scibek and Allen 2006a, Toews 2007) to project future groundwater recharge rates from climate scenarios. Our modeling approach in HELP3 follows the processes detailed by these
previous contributions, and additional details on the HELP3 model can be found in these studies. Limitations of the HELP3 model, particularly for application in arid regions, are discussed by Berger (2000) and Scanlon et al. (2002).

### 2.2.4 The hydrology model input data

HELP3 is driven by daily values of mean air temperature, precipitation, and solar radiation, and by annual average wind speed and quarterly relative humidity. The mean daily air temperature was determined by averaging the maximum and minimum daily air temperatures provided in the post-processed climate datasets. Quarterly relative humidity values were extracted from the Environment Canada database (EC 2013a) and held constant for each simulation. When needed, daily solar radiation data were generated using the methodology proposed by Hargreaves and Samani (1982) and expanded on by Allen (1997). This is a self-calibrated approach for determining solar radiation based on extraterrestrial radiation and the diurnal range in temperature.

In addition to climate data, HELP3 also requires several soil and land cover parameters. Because the Otter Brook catchment is small (9.5 km²) and has a relatively homogeneous land cover, spatial variability in surface or subsurface characteristics was not considered in the present study. For the subsurface, HELP3 requires that the saturated hydraulic conductivity, field capacity, porosity, evaporative zone depth, and wilting point be specified. These subsurface properties were determined from a combination of site visits, soil gradation analyses (Allard 2008), soil pedotransfer functions (Balland et al. 2008), previous HELP3 recharge simulations studies (Toews 2007), and model-recommended values (Waterloo Hydrogeologic 2002). The surface parameters, including
the maximum leaf area index, and runoff curve number, were specified based on previous recharge studies (Liggett and Allen 2010, Toews and Allen 2009a), standards (United States Soil Conservation Service and Water Resources 1985), climate data, and model recommendations (Schroeder et al. 1994, Waterloo Hydrogeologic 2002). The growing season length was adjusted for each climate scenario based on the annual temperature cycle (i.e., when daily temperature exceeded 10°C). A summary of the key input data is given in Table 2.3.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Hydraulic cond.</td>
<td>0.25 cm/s</td>
<td>Obtain from Hazen’s (1911) method</td>
</tr>
<tr>
<td>Evaporative zone depth</td>
<td>150 cm</td>
<td>(Toews 2007)</td>
</tr>
<tr>
<td>Deposit thickness</td>
<td>9.5 m</td>
<td>(Allard 2008)</td>
</tr>
<tr>
<td>Depth to GWT</td>
<td>7.5 m</td>
<td>(Allard 2008)</td>
</tr>
<tr>
<td>Field capacity</td>
<td>0.031 (vol/vol)</td>
<td>(Balland et al. 2008)</td>
</tr>
<tr>
<td>Wilting point</td>
<td>0.019 (vol/vol)</td>
<td>(Balland et al. 2008)</td>
</tr>
<tr>
<td>Porosity</td>
<td>0.417</td>
<td>(Waterloo Hydrogeologic 2002)</td>
</tr>
<tr>
<td>Unfrozen curve number</td>
<td>50</td>
<td>(US Soil Conservation Service and Water Resources 1985)</td>
</tr>
<tr>
<td>Max leaf area index</td>
<td>4</td>
<td>(Toews 2007)</td>
</tr>
</tbody>
</table>

The flow of data from the GCMs through to the HELP3 model is indicated in Figure 2.3. Although not depicted in Figure 2.3, there are feedback loops between the land surface/subsurface characteristics and the climate data. For example, the maximum leaf area index is specified explicitly by the user, but HELP3 simulates an annual cycle for the vegetative density and leaf area index as a function of the air temperature and solar radiation (Schroeder et al. 1994). This will impact the timing and magnitude of the
evapotranspiration regime. During the winter when air temperature is low, recharge ceases as precipitation accumulates as snowpack. HELP3 also simulates a decrease in late fall and early spring recharge by increasing the runoff curve number for colder air temperatures.

Figure 2.3. Algorithm for translating global climate model results into projections of temporally varying groundwater recharge.

2.3 Results

In general, the post-processing of the GCM and RCM climate data had a significant impact on the resultant climate series. Figure 2.4 shows the change in mean annual precipitation and air temperature due to the downscaling (HMLR or DT) and debiasing processes (DT, for the RCM simulations). The results demonstrate that near-surface climate data (e.g., precipitation and air temperature) produced by a statistical downscaling method can deviate significantly from the climate data produced by the GCM, as many statistical downscaling methods (e.g., HMLR) are driven by upper-air field predictors from the GCM.
The HELP3 simulations were not calibrated, but the percent of annual precipitation simulated to result in groundwater recharge for the reference period (range = 32–56%, mean = 44%) generally concurs with previous hydrograph studies (45%, Noble and Bray 1995) and water balance techniques (48%, in 1994, Jones 1997) for unconfined aquifers in the Little Southwest Miramichi River catchment. Figure 2.5 presents the annual average groundwater recharge simulated for the reference period (1961–2000) and for each climate scenario given in Table 2.2 for the future period (2046–2065). The most pronounced increase (58%) in annual average recharge, compared to the reference period, was obtained for the CGCM3-A2 climate data, while the most pronounced decrease in the average annual recharge (-6%) was obtained for the MIROC 3.2 HIRES-A1B scenario.

<table>
<thead>
<tr>
<th>Climate Scenario</th>
<th>Change in Precipitation Due to Post-Processing (%)</th>
<th>Change in Air Temperature Due to Post-Processing (°C)</th>
</tr>
</thead>
<tbody>
<tr>
<td>CGCM3-A2</td>
<td>58</td>
<td>0</td>
</tr>
<tr>
<td>CRCM 4.2.3 aev-A2</td>
<td>-5</td>
<td>-2</td>
</tr>
<tr>
<td>CRCM 4.2.3 agx-A2</td>
<td>-4</td>
<td>-3</td>
</tr>
<tr>
<td>CSIRO MK 3.0-B1</td>
<td>-5</td>
<td>-2</td>
</tr>
<tr>
<td>CSIRO MK 3.5-B1</td>
<td>-4</td>
<td>-3</td>
</tr>
<tr>
<td>MIROC 3.2 HIRES-A1B</td>
<td>-6</td>
<td>0</td>
</tr>
<tr>
<td>CGCM3-A1B</td>
<td>0</td>
<td>0</td>
</tr>
</tbody>
</table>

Figure 2.4. Changes in (a) mean annual precipitation (%) and (b) air temperature (°C) due to the post-processing of the GCM/RCM data. The point of reference is the raw GCM/RCM data for each climate scenario not the reference period data.
Figure 2.5. Average annual groundwater recharge for each climate-hydrology simulation. The recharge results obtained using the reference climate data (1961-2000) are indicated by the far left data point. The error bars indicate one standard deviation in the annual average recharge results.

In addition to variations in annual recharge, the HELP3 hydrology simulation results also indicated changes to the timing of recharge. Figure 2.6 gives the average monthly distribution of simulated recharge for the observed climate data and the CGCM3-A1B, CSIRO Mk3.0-B1, and MIROC 3.2 HIRES-A1B climate simulations. Results for these four simulations are presented because they span the range in projected annual average recharge. The normalized December recharge increases for all three future climate scenarios while the normalized May recharge significantly decreases for two of the three climate scenarios. Little to no recharge occurs in January and February because precipitation during those months is mainly in the form of snow.
Figure 2.6. The monthly distribution of average annual groundwater recharge for the 1961-2000 period (reference) and three climate-hydrology simulations for 2046-2065.

2.4 Discussion

2.4.1 The impact of the GCM, emission scenario, and downscaling method on the magnitude and timing of groundwater recharge

As indicated in Figures 2.2, 2.4 and 2.5, the local projected climate data and the simulated groundwater recharge are dependent on the selected GCM, emission scenario, and downscaling method. For example, the CRCM 4.2.3 aev-A2 and CRCM 4.2.3 agx-A2 climate scenarios were both generated with the A2 emission scenario, dynamically downscaled with the CRCM 4.2.3 model, and further downscaled/debiased using the DT method. However, the CRCM 4.2.3 aev-A2 and CRCM 4.2.3 agx-A2 data produced projected changes in mean annual recharge of +3.4% and +14.7% respectively (Figure
Applying our definition of uncertainty yields a GCM-induced recharge uncertainty of 11.3% for these two climate–hydrology simulations. This significant difference can be attributed to the two GCMs, Echam5 and CGCM3 (Table 2.2), selected to drive the CRCM 4.2.3 simulations.

The effect of the choice of the emission scenario can be seen in the CGCM3-A2 and CGCM3-A1B climate data. These climate series were both generated with the same GCM and downscaled using the HMLR algorithm. Figures 2.2 and 2.5 illustrate that the emission scenario had very little impact on the resultant mean annual climate data and subsequent simulated recharge, but the CGCM3-A2 recharge results are characterized by more annual variability (higher standard deviation). Thus, in this case, the selection of the emission scenario contributed very little uncertainty in the climate–hydrology simulations, at least on an annual average basis. In this example, the CGCM3-A1B data actually displayed greater changes in the precipitation and temperature data than the CGCM3-A2 data. Although A2 is a higher emission scenario on a global scale, these effects may not be manifested at a local scale for a given GCM and time period. The effect of the emission scenario would be expected to be more pronounced in later decades (e.g., 2061–2100) due to the thermal inertia of the ocean, which creates a lag between a climate forcing and its realisation in local atmospheric conditions (Huard 2011). It cannot be concluded from such a limited sample that the emissions scenario will always have the least impact on simulated groundwater recharge.

The effect of the downscaling/debiasing method is demonstrated by differences between the CGCM3-A2 and CRCM 4.2.3 aev-A2 precipitation and air temperature data indicated in Figure 2.2. Both of these climate scenarios were generated with the same
GCM and emission scenario (Table 2.2), but one was dynamically downscaled with an RCM and debiased with the DT method, while the other was statistically downscaled using the HMLR algorithm. HELP3 simulated a 58% increase in annual average recharge for the CGCM3-A2 climate data and only a 3% increase for the CRCM 4.2.3 aev-A2 data (Figure 2.5). Thus, the differences in downscaling/debiasing techniques contributed significant recharge uncertainty (55%) when comparing these two climate–hydrology simulations. These results are predictable given the significant effect that different post-processing techniques have on the resultant climate series (Figure 2.4).

The changes to the timing of the recharge indicated in Figure 2.6 are a result of the projected changes to the timing and magnitude of precipitation and air temperature. On average, monthly precipitation remained relatively constant throughout the year for the reference period. However, in general, the projected climate scenarios were characterized by increased variability in the distribution of monthly precipitation. The change in available soil moisture during these periods will directly impact groundwater recharge for that season. The decreases in May recharge projected for several of the climate scenarios are a result of an increase in winter temperature and consequently an earlier shift in the timing of the snowmelt. This phenomenon is apparent in Figure 2.6 where the MIRCO 3.2 Hires-A1B normalized recharge increases in March and April and decreases in May compared to the simulations conducted for the reference period. However, this effect is not as apparent for the CSIRO Mk 3.0-B1 data (Figure 2.6), which is likely due to the relatively small increase in annual average air temperature when compared to the MIRCO 3.2 Hires-A1B data (Figure 2.2). Thus, at northern latitudes, the selection of the GCM, downscaling/debiasing algorithm, and emission scenario may all
have a significant effect on the timing of the simulated snowmelt and consequently, the timing of simulated groundwater recharge.

2.4.2 Comparison to other recharge studies with multiple climate scenarios

The uncertainty in the projected annual average recharge for the present study (range = -6% to +58%, uncertainty = 64%) is larger than the uncertainty in recharge simulated by Allen et al. (2010) and Dams et al. (2012) (Table 2.1), and this likely arises because the present study considered multiple downscaling/debiasing methods in addition to multiple GCMs. However, as shown in Table 2.1, the uncertainty in the current study is approximately the same as ranges simulated by Holman et al. (2009), Jackson et al. (2011) and Ali et al. (2012) and considerably less than those simulated by Serrat-Capdevila et al. (2007), Döll (2009), and Crosbie et al. (2010, 2011b, 2013).

These uncertainties primarily arose from the approaches used to generate the climate data that drive the hydrologic models. Holman et al. (2009) indicated that more uncertainty arose from the choice of the downscaling method than the choice of the emission scenario. Crosbie et al. (2011b) found that the largest source of uncertainty could be attributed to the GCM, while the choice of the downscaling method was of secondary importance. The information given in Table 2.2, Figure 2.5, and the discussion above indicates that the variability in simulated future recharge for the present study arose primarily from the downscaling method, secondly from the GCM, and thirdly from the emission scenario. These findings generally agree with those of Holman et al. (2009), although that study only employed one GCM. However, these findings contrast with the study by Crosbie et al. (2011b) by suggesting that the downscaling/debiasing method has
more impact on the resultant climate data than the choice of the GCM. This difference likely arises from the more limited suite of GCMs utilized in the present study. Thus, this study suggests that multiple downscaling methods should be employed when projecting future recharge if a full range of uncertainty is to be determined.

For all of the studies summarized in Table 2.1, the magnitude of the change in future groundwater recharge was difficult to forecast given the uncertainty in climate–hydrology modeling. For example, Döll (2009) stated: ‘climate change scenarios cannot be used to quantitatively project the future development of groundwater resources’. Furthermore, many of these studies produced ranges that made it difficult to project the future trajectory of groundwater recharge. For example, Crosbie et al. (2010) concluded, ‘it [is] difficult to project the direction of the change in recharge . . . let alone the magnitude’ and ‘such variability in recharge estimates using different climate sequences means that making recommendations for water-resources management . . . is highly uncertain’. Crosbie et al. (2013) stated: ‘for most of [Australia] there is no consensus amongst the models on the direction of the change in recharge’. The present study produced an average change in mean annual recharge of +18% (Figure 2.5), but similar to what others have concluded, it would seem presumptuous to suggest that future recharge will increase given that three of the seven climate scenarios resulted in a decrease in recharge.

Several authors have attempted to address the uncertainty in recharge projections by applying a probabilistic approach or averaging the results (e.g., Ali et al. 2012, Crosbie et al. 2013, Jackson et al. 2011). This approach may assist water resource managers in understanding the uncertainty in future recharge projections; however, the findings are
prone to being statistically insignificant. For example, Jackson et al. (2011) found that the sign of the change in potential groundwater recharge could not be determined at the 95% confidence interval. Additionally, a problem arises when assigning probabilities to distinct climate scenarios, each of which should not necessarily be considered equally likely to occur.

A number of recent groundwater recharge studies have maintained that considerable value remains in projecting future recharge, despite the uncertainty in the simulated results. Serrat-Capdevila et al. (2007) suggested that their study could benefit policy makers because almost three quarters of their GCM simulations indicated a decreasing trend in precipitation; however, the present study suggests that their results would likely exhibit more uncertainty if they had employed more than one downscaling method. Döll (2009) suggested that valuable information can still be obtained from these studies to demonstrate the possible ranges in future groundwater resources that will require an adaptive response. Crosbie et al. (2013) stated that their study demonstrated that water resources managers should understand the uncertainty involved with making future groundwater resource decisions. In general, we agree that the major contribution of the studies summarized in Table 2.1 has been to demonstrate that future groundwater resources could potentially change significantly, but that the magnitude and trajectory of this change is uncertain. These results are not surprising considering that precipitation is poorly resolved and inconsistent in GCMs (Döll, 2009), and that variations in recharge can be at least 2–3 times greater than variations in precipitation (Ali et al., 2012).
2.4.3 Suggestions for future climate change-groundwater recharge research

Given the number of recent studies, including the present example, that explicitly or implicitly suggest it is difficult to predict the magnitude and direction of groundwater recharge, perhaps it is time that we refocus our efforts rather than continue with similar simulation approaches. The following suggestions for future research are offered based on the findings of the illustrative case study detailed above and the gaps identified in the literature reviewed:

1) To date, groundwater recharge projections have exhibited significant uncertainty. Thus, there should be an exploration of new techniques for (a) better quantifying and communicating the uncertainty in projected recharge, such as employing probabilistic approaches that recognize that projected emission scenarios may not have equal likelihood, and (b) reducing the uncertainty in the driving climate data and resultant recharge projections. Because it appears that hydrologic models used to estimate groundwater recharge contribute relatively little to the overall uncertainty (Crosbie et al. 2011b, Kingston and Taylor 2010, Teng et al. 2012), the focus for improvements should be the GCMs, post-processing methods, and emission scenarios. For example, uncertainty in the driving data could be reduced by abandoning climate modeling techniques (e.g., coupled GCM-downscaling method simulations) that do not adequately reproduce recent climate.

2) Studies investigating the impacts of climate change on groundwater recharge should employ multiple post-processing methods in addition to multiple GCMs.
and emission scenarios. There has been a tendency to deemphasize the impact of the downscaling/debiasing methods, but recent studies, including the present one, have demonstrated that downscaling may contribute significant uncertainty to the simulated groundwater recharge. In accordance with the first suggestion, only downscaling methods that have been shown to perform well for the study region in question should be employed.

3) There remains a lack of data on the relationship between long-term climate change and groundwater recharge, although several studies (e.g., Chen et al. 2004, Hughes et al. 2012, Rivard et al. 2009) have examined the relationship between seasonal or decadal climate variations and groundwater levels. Kundzewicz et al. (2007, Table 3.1) found ‘no evidence for [a] ubiquitous climate-related trend’ for groundwater resources. Thus, there should be an increase in field-based studies that track climate change-induced impacts on groundwater recharge.

4) If recharge projections are to be obtained for a period only a few decades into the future, it may be reasonable to assume constant land cover conditions. However, for recharge projections for time periods in the more distant future there should be an increased effort to address simultaneous changes to the socioeconomic environment, land cover, and climate conditions. It seems unreasonable to project climate change impacts on groundwater recharge for a century into the future and assume that water supply needs and land cover conditions will be intransient (Döll 2009, Holman 2006, Holman et al. 2012).
5) Climate–groundwater interactions are currently simulated by explicitly coupling GCMs with hydrological models. Because of the interdependence of groundwater and land-energy feedbacks (Maxwell and Kollet 2008) it may be more appropriate to directly simulate the groundwater recharge response to climate change within the land surface model of the GCM. More hydrogeologists should be researching and improving the groundwater components of existing GCM land surface models (Gulden et al. 2007).

2.5 Conclusions

Downscaled climate scenarios for central New Brunswick, Canada were used to drive HELP3 and simulate future groundwater recharge. The simulated data exhibited uncertainty both in the direction and magnitude of future changes in mean annual recharge (-6 to +58%). The variations arose from the selection of the GCMs, emission scenarios, and downscaling algorithms employed to generate the climate data. For the particular combinations examined, the largest variations resulted from the choice of the post-processing approach (i.e., statistical downscaling or dynamical downscaling coupled with statistical debiasing).

This study has demonstrated the limitations inherent in predicting future changes in groundwater recharge using downscaled climate change scenarios. For example, a single projection of climate based on one emission scenario, simulated with one GCM, and downscaled using only one approach, will provide limited insight into potential changes in recharge. Although significant efforts are required to produce downscaled
climate data using a variety of GCMs, emission scenarios and downscaling methods, it is concluded that this approach will provide a more honest representation of the uncertainty involved in assessing the hydrogeological impacts of climate change.

The IPCC correctly identified a gap in the knowledge of the impact of climate change on groundwater resources (Kundzewicz et al. 2007), and numerous recent studies have attempted to bridge this gap. However, these studies have demonstrated that we do not currently have the ability to quantitatively predict the magnitude or direction of the impact of climate change on groundwater resources with a high degree of confidence. This does not imply that we should abandon the modeling of projected recharge; indeed there are many opportunities for advancing this field, including constraining climate projections; collecting extensive time series of recharge and climate data; simulating simultaneous changes in land cover, water withdrawals and climate change; and developing increasingly complex land surface models within GCMs.

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Abstract

Groundwater temperature is an important water quality parameter that affects species distributions in subsurface and surface environments. To investigate the response of subsurface temperature to atmospheric climate change, an analytical solution is derived for a one-dimensional, transient conduction–advection equation and verified with numerical methods using the finite element code SUTRA. The solution can be directly applied to forward model the impact of future climate change on subsurface temperature profiles or inversely applied to produce a surface temperature history from measured borehole profiles. The initial conditions are represented using superimposed linear and exponential functions, and the boundary condition is expressed as an exponential function. This solution expands on a classic solution in which the initial and boundary conditions were restricted to linear functions. The exponential functions allow more flexibility in matching climate model projections (boundary conditions) and measured temperature–depth profiles (initial conditions). For example, measured borehole temperature data from the Sendai Plain and Tokyo, Japan, were used to demonstrate the improved accuracy of the exponential function for replicating temperature–depth profiles. Also, the improved accuracy of the exponential boundary condition was demonstrated using air temperature anomaly data from the Intergovernmental Panel on Climate Change.

These air temperature anomalies were then used to forward model the effect of surficial thermal perturbations in subsurface environments with significant groundwater flow. The simulation results indicate that recharge can accelerate shallow subsurface warming, whereas upward groundwater discharge can enhance deeper subsurface warming. Additionally, the simulation results demonstrate that future groundwater temperatures obtained from the proposed analytical solution can deviate significantly from those produced with the classic solution.

**Keywords:** groundwater temperature, subsurface warming, thermal regime, climate change, soil temperature

### 3.1 Introduction

#### 3.1.1 Importance of groundwater temperature

Groundwater temperature is less variable than surface water temperature; thus, groundwater discharge can stabilize river temperatures during winter and summer months (Caissie 2006, Hayashi and Rosenberry 2002, Webb et al. 2008). Discrete cold-water plumes formed by groundwater–surface water interactions have also been shown to provide critical thermal relief for cold-water fish during high-temperature events (Breau et al. 2011, Cunjak et al. 2005, Torgersen et al. 2012). In a warming climate, the ecology of riverine systems may become increasingly dependent on groundwater discharge. Groundwater temperature can also affect subsurface biogeochemical processes and influence the distribution of unicellular and multicellular microorganisms in the subsurface and in surface water at points of groundwater discharge (Andrushchysyn et

Recent climate change effects on surface and shallow subsurface temperature have already been observed (e.g., Qian et al. 2011), and climate model projections for the coming decades indicate that significant warming will occur on a regional and global scale (Solomon et al. 2007). Groundwater temperature will respond to a warming climate as rising ground surface temperature trends will be propagated through the subsurface (Taylor and Stefan 2009); however, very little research has been conducted on the impact of future climate change on groundwater temperature. Because aquifer and river thermal regimes are interrelated, there is a need for further studies examining the effect of climate change on groundwater temperature. For example, Mayer (2012) has recently noted: ‘The future warming of shallow groundwater temperature...could affect stream temperatures...One challenge to assessing this is that most studies of climate change impacts to groundwater have focused on changes to recharge and subsurface flow rather than effects to groundwater temperature.’

### 3.1.2 History and limitations of conduction heat transport mathematics

The subsurface thermal regime is driven by energy and water exchanges at the ground surface and the geothermal heat flux from the Earth’s interior. Ground surface temperature signals are transported and retained in the subsurface, but the amplitude of the temperature variation (e.g., daily or seasonal temperature fluctuations) exponentially attenuates with depth because of the process of heat diffusion and the ability of the soil
and water to retain heat (Carslaw and Jaeger 1959, Lunardini 1981, Williams and Smith 1989). High-frequency ground surface temperature signals (e.g., monthly) are only retained at shallow depths, but lower frequency signals (e.g., decadal climate change) may be retained at great depths (Lesperance et al. 2010).

Early subsurface heat transport analyses were predicated on the assumption that conduction was the dominant form of heat transport. Birch (1948) proposed that the distribution of subsurface temperature could be used to infer past climatic conditions by analyzing a transient form of the conduction equation. His theory has since been expanded to reproduce paleoclimates by inverting temperature–depth profiles (e.g., Beltrami et al. 1995, Harris and Chapman 1997, Lachenbruch and Marshall 1986, Lesperance et al. 2010, Mareschal and Beltrami 1992, Pollack et al. 1998). These results have been used to assess the validity of global climate model (GCM) simulations of past climates. Conduction-based analytical solutions have also been used to forward model the effects of past climate change by using measured or GCM-simulated air temperature history to perturb an assumed initial geothermal profile (Beltrami et al. 2005, Gonzalez-Rouco et al. 2006). Other solutions to the transient heat conduction equation under a variety of initial and boundary conditions have been compiled by Carslaw and Jaeger (1959), Özışık (1968), and Crank (1980).

Conduction-based techniques for paleoclimate inversion may be invalid in regions of significant groundwater flow because the subsurface thermal signature used to infer surficial climate change may have resulted from advective heat transport (Bodri and Cermak 2005, Ferguson et al. 2006, Kukkonen et al. 1994, Lewis and Wang 1992). For example, both ground surface temperature increases and downward groundwater flow can
induce a concave upward temperature distribution (Figure 3.1). Bodri and Cermak (2005) observed that the effects of advection were discernible in their simulated one-dimensional temperature profiles when Darcy flux exceeded $1 \times 10^{-9}$ m s$^{-1}$ (0.032 m yr$^{-1}$). Thus, the assumption that conduction is the only significant form of heat transport may be invalid for regions where water or vapor transport is significant (Hinkel and Outcalt 1993, Kane et al. 2001, McKenzie and Voss 2013, van der Kamp and Bachu 1989, Woodbury and Smith 1985).

Figure 3.1. Temperature depth profiles for a stable climate as a function of the groundwater flow direction. Below the seasonal penetration depth, temperature–depth profiles are concave upward in groundwater recharge areas (vertical downward flow), linear in lateral flow areas (with no horizontal thermal gradient) and convex upward in discharge areas (vertical upward flow) (after Anderson 2005, Saar 2011, Taniguchi et al. 1999).
3.1.3 Groundwater temperature as a tracer for groundwater velocity

Suzuki (1960) first proposed that temperature–depth profiles could be used to estimate rates of vertical groundwater flow by solving the following one-dimensional conduction–advection equation for semi-infinite media:

$$\lambda \frac{\partial^2 T}{\partial z^2} - q c_w \rho_w \frac{\partial T}{\partial z} = c \rho \frac{\partial T}{\partial t}$$

where $\lambda$ is the thermal conductivity of the medium (M L t$^{-3}$ T$^{-1}$), $T$ is temperature, $z$ is depth below ground surface (L), $q$ is the vertical Darcy flux (positive downwards, L t$^{-1}$), $c_w \rho_w$ is the volumetric heat capacity of water (M L$^{-1}$ t$^{-2}$ T$^{-1}$), $c \rho$ is the volumetric heat capacity of the medium (M L$^{-1}$ t$^{-2}$ T$^{-1}$), and $t$ is time. Suzuki (1960) solved this equation by applying a periodic first-type boundary condition to represent the annual cycle of ground surface temperature. Stallman (1963, 1965) later provided a more detailed derivation of Equation 3.1 and the solution for determining groundwater velocity on the basis of the subsurface temperature distribution. The methods for using groundwater temperature to estimate groundwater velocity have been comprehensively reviewed by Anderson (2005) and Saar (2011).

3.1.4 Combined effects of groundwater flow and ground surface temperature rise

A temperature–depth profile that is assumed to have been induced by vertical groundwater flow may have actually arisen from long-term climate change (Ferguson and Woodbury 2005, Reiter 2005); this is the converse of the problem facing paleoclimatologists. To address this issue, Taniguchi et al. (1999) modified an analytical
solution by Carslaw and Jaeger (1959, p. 388) to account for both surface temperature changes and vertical groundwater flow. The governing equation is of the same form as Equation 3.1; the initial conditions, boundary condition and solution are given as follows:

**Initial Conditions:** \[ T(z, t = 0) = T_0 + az \]

**Boundary Condition:** \[ T(z = 0, t) = T_0 + \varphi t \]

**Solution:** \[ T(z, t) = T_0 + a(z - vt) + \frac{1}{2v} (\varphi + va) \times \left[ (vt - z) \times \text{erfc} \left( \frac{z - Vt}{2\sqrt{Dt}} \right) + (vt + z) \exp \left( \frac{vz}{\alpha} \right) \text{erfc} \left( \frac{z + vt}{2\sqrt{Dt}} \right) \right] \] (3.2)

where \( T_0 \) is the initial surface temperature, \( a \) is the geothermal gradient (\( T \) \( L^{-1} \)), \( \text{erfc} \) is the complementary error function, \( \varphi \) is the slope of the linear ground surface temperature rise (\( T \) \( t^{-1} \)), \( D \) is the thermal diffusivity of the medium = \( \lambda(c\rho)^{-1} \) (\( L^2 \) \( t^{-1} \)), and \( v = qc_w\rho_w(c\rho)^{-1} \) (\( L \) \( t^{-1} \)).

Taniguchi et al. (1999) proposed that this analytical solution could be used to determine groundwater fluxes in regions of measured surface temperature rise due to urbanization or climate change; conversely, if the groundwater flux was known, the solution could be inverted to reproduce a linear ground surface temperature history from borehole temperature data (Miyakoshi et al. 2003, Taniguchi 2006, Taniguchi et al. 2003, Taniguchi and Uemura 2005, Taniguchi et al. 2009, Uchida and Hayashi 2005, Uchida et al. 2003). This solution has also been recently applied to forward model future groundwater temperature based on downscaled climate change projections for the Sendai Plain, Japan (Gunawardhana and Kazama 2011, Gunawardhana et al. 2011).
3.1.5 Improved boundary and initial conditions using exponential functions

Global climate models simulate atmospheric, surficial and oceanic processes to project future climate regimes due to increased atmospheric greenhouse gas concentrations (Solomon et al. 2007). The Intergovernmental Panel on Climate Change (IPCC 2007) compiled multi-model averages for global air temperature anomalies for 2000–2099 from 24 GCMs driven by emission scenarios B1, A1B and A2. The multi-model A2 emission scenario results are characterized by significant concavity that may be better represented by boundary conditions containing an exponential function, rather than the linear boundary conditions available in existing conduction–advection solutions. Some previous solutions to the transient conduction equation have employed a series of superimposed step functions as the surface boundary condition (e.g., Beltrami et al. 2005). This boundary condition can mimic surface temperature history or projections better than a linear function, but these solutions have limited applicability for simulating subsurface temperature evolution when advection is significant.

Although we utilized air temperature anomalies to form our boundary condition, we acknowledge that the ground surface temperature, not the air temperature, drives the subsurface thermal regime (Lunardini 1981). It is well established that high-frequency (e.g., daily or seasonal) air temperature and ground surface temperature variations are decoupled because of snow cover insulating effects, latent heat arising from freeze-thaw phase changes, shading from vegetation and other factors (e.g., Beltrami et al. 2005, Beltrami and Kellman 2003, Smerdon et al. 2006, Zhang 2005). This short-term decoupling does not necessitate a decoupling between low-frequency (e.g., decadal) variations in air temperature and ground surface temperature, and there remains an
ongoing debate regarding the coupling of air temperature and ground surface temperature anomalies in past or projected long-term climate change (e.g., Beltrami et al. 2005, Gonzalez-Rouco et al. 2006, Mann and Schmidt 2003, Pollack et al. 2005, Smerdon et al. 2004, Smerdon et al. 2006, Stieglitz and Smerdon 2007). It is possible that the significant rate of climate change projected for the coming decades may reduce the average length of the snow-covered period and the mean annual snowpack depth at some latitudes. This would reduce the mean annual thermal conductivity between the lower atmosphere and the ground surface and could result in differences between projected changes in decadal air temperature and ground surface temperature trends (Kurylyk et al. 2013, Mellander et al. 2007). However, following the approach of others listed earlier, we assume that long-term ground surface temperature changes will closely follow air temperature changes.

Measured temperature–depth profiles are also typically non-linear. This is particularly noticeable in the shallow subsurface because of the effects of vertical groundwater fluxes or recent warming due to climate change or urbanization (Ferguson and Woodbury 2004, 2005, 2007, Kukkonen et al. 1994, Mareschal and Beltrami 1992). The present borehole temperature profile represents the initial conditions for forward modelling the subsurface thermal response to climate change; thus, initial conditions functions should be capable of matching non-linear temperature profiles.
3.1.6 Research objectives

The intent of this contribution is to address the limitations associated with the initial and boundary conditions employed in the analytical solution developed by Carslaw and Jaeger (1959) and modified by Taniguchi et al. (1999). Our first objective is to demonstrate that exponential functions, which are still amenable to analytical solutions, are better representations for surface boundary conditions and initial temperature–depth profiles. Our second objective is to develop an analytical solution for which exponential terms are included in both the boundary and initial conditions and to apply this solution in several illustrative examples.

Although numerical solution methods can be employed to simulate subsurface temperature evolution; in some cases, analytical solutions offer important advantages. Javandel et al. (1984) noted four advantages that are relevant to the current topic: (1) analytical methods are more efficient than numerical methods when the system is poorly defined or uncertain; (2) analytical methods are more economical (computationally) than numerical methods; (3) analytical methods are useful for initial estimations; and (4) analytical methods remove the need for experienced modellers or complex numerical codes. In addition to these advantages, analytical methods can allow the user to quickly simulate the effect of varying the governing equation parameters (e.g., soil thermal properties or groundwater velocity) and the initial and boundary conditions. Furthermore, analytical solutions are more stable than numerical methods and are not subject to the spatiotemporal discretization problems that may arise when employing numerical methods.
3.2 Methods

3.2.1 Development of exponential initial and boundary conditions

As previously discussed, the surface temperature projections produced by GCMs for some emission scenarios may be more appropriately simulated with an exponential Dirichlet boundary condition (Equation 3.3) rather than a linear surface boundary condition (Equation 3.2):

\[ T(z = 0, t) = T_i + b \exp(ct) \]  

(3.3)

where \( T_i \) (T), \( b \) (T), and \( c \) (t\(^{-1}\)) are fitting parameters. The boundary conditions in Equations 3.2 and 3.3 indicate that \( T_0 \) should equal \( T_i + b \) for both boundary conditions to have the same value at \( t = 0 \). The ability of Equation 3.3 to represent projected surface temperature trends will be demonstrated in the results.

Also, existing temperature–depth profiles may be better represented with the superposition of exponential and linear functions (Equation 3.4), rather than with a simple geothermal gradient (Equation 3.2):

\[ T(z, t = 0) = T_i + a z + \delta \exp(dz) \]  

(3.4)

where \( T_i \) is a fitting parameter for the initial conditions (T), \( a \) is the general geothermal gradient (T L\(^{-1}\)), and \( \delta \) (T) and \( d \) (L\(^{-1}\)) are fitting parameters to account for recent surface temperature changes or vertical groundwater flow. The superposition of linear and exponential functions to form the initial conditions shall hereafter be referred to as the ‘exponential initial conditions’. Equation 3.4 is flexible and capable of producing initial conditions that mimic a variety of measured temperature–depth profiles. If \( d \) is negative, the last term in Equation 3.4 exponentially decays with depth and reproduces the shallow
thermal signature seen in the profiles given in Figure 3.1 arising from surficial warming and/or recharge (+\(\delta\) for concave upwards) or surficial cooling and/or discharge (-\(\delta\) for convex upwards). The ability of Equation 3.4 to reproduce existing temperature–depth profiles will also be demonstrated in the results.

### 3.2.2 Development of the modified analytical solution

Equation 3.1 was manipulated and expressed in the standard form employed by Carslaw and Jaeger (1959):

\[
D \frac{\partial^2 T}{\partial z^2} - v \frac{\partial T}{\partial z} = \frac{\partial T}{\partial t}
\]

(3.5)

where \(D\) is the thermal diffusivity of the medium \(\lambda(c\rho)^{-1}\) (L\(^2\) t\(^{-1}\)), and \(v = qc_w\rho_w(c\rho)^{-1}\) (L t\(^{-1}\)). The initial and boundary conditions were taken as Equations 3.4 and 3.3, respectively. Our formulation for thermal conductivity/diffusivity ignores the effects of hydrodynamic dispersion, although several researchers have suggested alternative formulations that account for these effects (Ferguson 2007, Molina-Giraldo et al. 2011, Sauty et al. 1982).

Typically, a partial differential equation (PDE) of this form would be first reduced to a pure diffusion equation by means of a transform of the independent variables \(z\) and \(t\) (Farlow 1982) or the dependent variable \(T\) (Ogata and Banks 1961). However, these techniques tend to cause the boundary condition to become overly complex or spatially transferred so that it is no longer a true boundary condition. In this case, a more appropriate technique is to first reduce the PDE to an ordinary differential equation by
means of the Laplace integral transform (Farlow 1982, Trim 1990). The subsidiary
equation for this governing equation and the associated initial conditions is then:

\[ D \frac{\partial^2 \tilde{T}}{\partial z^2} - v \frac{\partial \tilde{T}}{\partial z} - p \tilde{T} = -T_i - a z - \delta \exp(dz) \]  

(3.6)

The boundary condition (Equation 3.3) was used to find the solution to the
subsidiary equation according to the method of undetermined coefficients (Zill 2005):

\[ \tilde{T} = \left( \frac{a z + T_i}{p} - \frac{va}{p^2} + \frac{\delta}{-Dd^2 + vd + p} \exp(dz) \right) + \exp \left( \frac{vz}{2D} \right) \times \]

\[ \left( \frac{T_i - T_i}{p} + \frac{va}{p^2} + \frac{b}{p - c} \right) \exp \left( -z \sqrt{\frac{v^2}{4D^2} + \frac{p}{D}} \right) \]  

(3.7)

where \( \tilde{T}(z, p) \) is the Laplace transform of the temperature \( T(z, t) \), and \( p \) is the Laplace
transform of \( t \) or the time expressed in the frequency domain. By first employing the shift
theorem (Trim 1990), then applying appropriate tabulated inverse Laplace transforms
(Carslaw and Jaeger 1959, Roberts and Kaufman 1966) and finally making the
appropriate simplifications, the following solution was obtained (see Appendix 1):

\[ T = az + T_i - vat + \delta \exp(Dd^2 t + dz - vd t) + \frac{(T_i - T_i)}{2} \left[ \text{erfc} \left( \frac{z}{2\sqrt{D}t} - \frac{\nu}{2} \sqrt{\frac{T}{D}} \right) + \exp \left( \frac{vz}{2D} \right) \text{erfc} \left( \frac{z}{2\sqrt{D}t} + \frac{\nu}{2} \sqrt{\frac{T}{D}} \right) \right] \]

\[ + \frac{va}{2} \left[ t - \frac{z}{v} \right] \text{erfc} \left[ \frac{z}{2\sqrt{D}t} - \frac{\nu}{2} \sqrt{\frac{T}{D}} \right] + \left( t + \frac{z}{v} \right) \times \text{erfc} \left( \frac{z}{2\sqrt{D}t} + \frac{\nu}{2} \sqrt{\frac{T}{D}} \right) \]

\[ + \frac{b}{2} \exp \left( \frac{vz}{2D} + t \right) \left[ \text{erfc} \left( -z \sqrt{\frac{v^2}{4D^2} + \frac{v^2}{4D^2} + c} \right) \right] \times \exp \left( \frac{vz}{2D} + \frac{v^2}{4D^2} + c \right) \text{erfc} \left( \frac{z}{2\sqrt{D}t} - \frac{\nu}{2} \sqrt{\frac{T}{D}} \right) \]

\[ - \frac{\delta}{2} \exp \left( \frac{vz}{2D} + Dd^2 - vd \right) t \left[ \text{erfc} \left( z \right) \text{erfc} \left( \frac{z}{2\sqrt{D}t} - \frac{\nu}{2} \sqrt{\frac{T}{D}} \right) \right] \]

(3.8)
If desired, Equation 3.8 can be solved for an initial linear temperature–depth profile by setting the $\delta$ term in Equations 3.4 and 3.8 to zero.

This solution can be evaluated in a spreadsheet; however, the product term involving the exponential function and the complementary error function can be problematic. Warrick (2003) gives the following approximation that was used to evaluate Equation 3.8:

$$\exp(x) \times \text{erfc}(y) = \begin{cases} y < 3, & \exp(x) \times \text{erfc}(y), \\ \frac{\exp\left(x - y^2\right)}{\pi^{0.5} y} \times \left(1 - \frac{0.5}{y^2} + \frac{0.75}{y^4}\right) & \end{cases}$$ (3.9)

### 3.2.3 Verification of the analytical solutions with numerical methods

Equation 3.8 was verified by comparisons to simulations performed within the groundwater flow and heat transport model SUTRA (Voss and Provost 2010). SUTRA is a robust finite element model that accommodates variably saturated, multi-dimensional groundwater flow and heat transport. For the present study, the SUTRA parameters were set so that the simulations were performed for one-dimensional flow and heat transport, fully saturated conditions, and spatiotemporally-constant groundwater velocity. In this case, the SUTRA governing equations reduce to Equation 3.1. Figure 3.2 shows the simplified physical scenario and numerical modeling setup.
Figure 3.2. The assumed (a) physical scenario and (b) numerical model domain used for performing SUTRA simulations. Appropriate initial and boundary conditions were imposed in SUTRA to constrain the heat and fluid flow to the vertical dimension.

3.3 Results

3.3.1 Assessment of the exponential boundary condition

Two synthetic curves were generated to represent the IPCC (2007) projected temperature anomalies for each emission scenario for 2000–2099. The first was generated using Equation 3.3 with the fitting parameters optimized to minimize the root-mean-square error (RMSE) between the synthetic data and the IPCC-projected data. The second function was developed by connecting the IPCC initial and final temperature anomalies for 2000–2099 with a straight line (boundary condition, Equation 3.2) (e.g.,
Gunawardhana et al. 2011, Gunawardhana and Kazama 2011). These curves are shown in Figure 3.3. It is evident that, particularly for the higher emission scenario A2, the exponential and linear curves deviate significantly from each other at mid-century (maximum difference ~ 0.5 °C). The RMSE between each curve and the associated IPCC GCM ensemble indicated that the exponential function (RMSE= 0.032, 0.053 and 0.039 for the B1, A1B and A2 IPCC data, respectively) provided a better fit to the AT projections than did the linear function (RMSE= 0.115, 0.081 and 0.346 for the B1, A1B and A2 data, respectively).

Figure 3.3. IPCC multi-model, globally averaged air temperature anomalies for 2000–2099 relative to 1980–1999 for emission scenarios B1, A1B and A2 (data from, IPCC 2007) and the linear (boundary condition, Equation 3.2) and exponential (Equation 3.3) functions generated to match the projections. The equation parameters are indicated.
3.3.2 Assessment of the exponential initial conditions

Measured subsurface temperature data were obtained for the Sendai Plain, Japan (L. Gunawardhana, personal communication, 2012), and for Tokyo, Japan (M. Taniguchi, personal communication, 2012) to demonstrate the ability of Equation 3.4 to match observed, present day temperature–depth profiles. Figure 3.4 shows the temperature data for depths below the seasonal penetration depth (>10 m). A linear function (initial conditions, Equation 3.2) and an exponential function (Equation 3.4) were used to match the measured profiles by minimizing the RMSE between the measured and calculated temperature–depth profiles (Figure 3.4).

![Image of temperature data for Sendai Plain and Tokyo](image)

**Figure 3.4.** Measured borehole temperature data from (a) the Sendai Plain, Japan (data from Gunawardhana et al. 2011) and (b) Tokyo, Japan (data from Taniguchi et al. 1999). The associated linear and exponential are also indicated.
The fitting parameters for the initial condition functions (Figure 3.4) are indicated in Table 3.1 along with the associated RMSE and correlation coefficient (R values). It is evident that the exponential function does a far superior job of replicating the measured profiles in all five instances. It should be noted that the solution presented by Taniguchi et al. (1999) can be utilized to produce present-day, non-linear temperature-depth profiles, but doing so requires several intermediate calibration and simulation steps and invokes assumptions about initial groundwater flow rates (e.g., Gunawardhana et al. 2011).

Table 3.1. Fitting parameters for each curve shown in Figure 3.4 and the associated RMSE and R values for (a) the Sendai Plain and (b) Tokyo.

<table>
<thead>
<tr>
<th>Curves</th>
<th>Fitting Parameters</th>
<th>Fit to Data</th>
</tr>
</thead>
<tbody>
<tr>
<td>Description</td>
<td>$T_0$ (°C)</td>
<td>$a$ (°C m$^{-1}$)</td>
</tr>
</tbody>
</table>
| (a). Sendai Plain Data
| Well 1-Linear | 13.261             | 0.0309       |             |             |             |             | 0.0965 | 0.9790 |
| Well 4-Linear | 13.501             | 0.0128       |             |             |             |             | 0.0817 | 0.9080 |
| Well 5-Linear | 14.095             | 0.0259       |             |             |             |             | 0.1921 | 0.9237 |
| Well 1-Expo   | 7.804 -0.0596      | 5.889 0.01054|             |             |             |             | 0.0141 | 0.9996 |
| Well 4-Expo   | 7.0197 -0.0802     | 6.856 0.00977|             |             |             |             | 0.0121 | 0.9981 |
| Well 5-Expo   | 7.5245 0.0949      | 7.5656 -0.0185|             |             |             |             | 0.0607 | 0.9926 |
| (b). Tokyo Data
| Well 1-Linear | 15.382             | 0.0137       |             |             |             |             | 0.3270 | 0.9117 |
| Well 74-Linear| 16.328             | 0.0214       |             |             |             |             | 0.2595 | 0.9542 |
| Well 1-Expo   | -2.886 0.0701      | 19.513 -0.0046|             |             |             |             | 0.0257 | 0.9994 |
| Well 74-Expo  | 15.603 0.0283      | 3.352 -0.0670|             |             |             |             | 0.0440 | 0.9987 |
3.3.3 Results for illustrative examples with future surface warming

Temperature–depth profiles were generated from Equation 3.8 for hypothetical surface warming and groundwater flow scenarios having the thermal properties, initial conditions and boundary conditions described below. These scenarios were also simulated with the SUTRA numerical code and thus served as verification problems. The thermal properties were obtained from Bonan (2008) and Oke (1978). The values assumed for the thermal conductivity of the medium $\lambda$, volumetric heat capacity of the medium $c_\rho$, and heat capacity of water $c_w\rho_w$ were 2.20 W m$^{-1}$ °C$^{-1}$, $2.96 \times 10^6$ J m$^{-3}$ °C$^{-1}$, and $4.18 \times 10^6$ J m$^{-3}$ °C$^{-1}$, respectively. These yield an effective heat diffusivity $D$ for the medium of $7.43 \times 10^{-7}$ m$^2$ s$^{-1}$. An initial surface temperature of 14°C was assumed for the simulations; this is approximately the current average global surface air temperature (Jones et al. 1999). In all cases, the initial conditions (black series) were assumed to be concave upwards due to recent warming. A typical geothermal gradient $a$ of 0.025 °C m$^{-1}$ was imposed below the near surface concavity (Equation 3.4). The exponential boundary condition used to match the A2 projections (Figure 3.3) was applied for the surficial boundary condition, as the A2 projections are often considered to be the most realistic of the IPCC emission scenarios. In its upcoming Fifth Assessment Report, the IPCC is adopting ‘representative concentration pathways’ (RCP, van Vuuren et al. 2011); the highest of these RCPs generally predicts more extreme warming than the A2 emission scenario. The exponential curve fitted to the A2 projections was shifted upwards by 14°C to represent the actual surface temperature rather than the temperature anomalies. The intent of using globally averaged current and projected surface
temperatures for the boundary condition is to simulate the subsurface response to climate change at a representative location.

The temperature–depth results shown in Figure 3.5 indicate that the analytical results (Equation 3.8) are in excellent agreement with the SUTRA simulations. It is quite clear that at this groundwater flux (0.5 m yr\(^{-1}\), Figure 3.5a and 3.5b), advection is a significant heat transport mechanism. Figure 3.5a indicates that groundwater recharge could accelerate the transport of warming surface temperatures into the shallow subsurface. Additionally, Figure 3.5b demonstrates that groundwater flow in discharge regions could exacerbate the increase in subsurface temperature, because the subsurface is being warmed from above (climate change) and from below (advective heat transport from deeper geothermal regions). Thus, it appears that the direction of the groundwater velocity will play an important role in determining the subsurface thermal response to climate change. Simulations for other groundwater fluxes were performed, and our results generally concurred with those of Bodri and Cermak (2005) in that groundwater fluxes exceeding 0.03 m yr\(^{-1}\) discernibly perturbed the temperature profiles from a purely conductive regime. The temperature–depth profiles in Figure 3.5c exhibit a typical conductive response to a surficial thermal perturbation (Carslaw and Jaeger 1959, Taniguchi et al. 1999).
Figure 3.5. Temperature–depth profiles generated with SUTRA and the analytical solution (Equation 3.8) for the case of (a) recharge (downward flow), (b) discharge (upward flow) and (c) negligible flow. The equation parameters used to generate the initial conditions (IC) and boundary conditions (BC) are given in the inset. The thermal properties are provided in the text.

3.3.4 Comparison of Equation 3.8 and the Taniguchi et al. (1999) solution

Previously compiled borehole data from the Sendai Plain were utilized to demonstrate the potential differences that may arise from the selection of the initial and boundary conditions (Equation 3.8 vs. Equation 3.2). These data were chosen as recent research has demonstrated that groundwater flow and heat transport in the Sendai Plain is primarily one-dimensional (Gunawardhana et al. 2011). The linear and exponential
functions fitted to the temperature–depth profile in borehole 5 (Figure 3.4 and Table 3.1) were used as initial conditions for the simulations. Borehole 5 was chosen because it contained the deepest temperature profile. For both sets of initial conditions (i.e., linear and exponential), the initial surface temperatures were found by extrapolating the curves up to the ground surface. The boundary conditions were then developed by adding the initial surface temperature to the exponential and linear A2 global surface temperature anomaly functions (Figure 3.3).

A vertical groundwater recharge flux of 0.130 m yr\(^{-1}\) was imposed to match the value suggested by Gunawardhana et al. (2011, Table 3). The thermal properties listed earlier were also utilized for these simulations. The analytical solution presented by Taniguchi et al. (1999) was applied to determine temperature–depth profiles for the case of linear initial conditions (from Borehole 5, Figure 3.4a and Table 3.1), and a linear boundary condition (Figure 3.3). Equation 3.8 was used to determine temperature–depth profiles for the case of exponential initial and boundary conditions.

Figure 6 demonstrates the significant thermal effects of the different initial conditions (black series) and boundary conditions. The initial condition has an inevitable impact on the boundary condition because the two conditions (initial and boundary) should coincide at \( t = 0, z = 0 \). In this case, the subsurface temperatures produced by the two solutions differ on the order of 1°C in the shallow (upper 20 m) zone and 2°C in the deeper (e.g., 100 m) zone.
Figure 3.6. Temperature–depth profiles generated from Equation 3.8 and the solution by Taniguchi et al. (1999) (Equation 3.2) when the exponential and linear temperature profiles matched to Borehole 5 data (Figure 3.4a and Table 3.1) are used for initial conditions, and the surface is perturbed by the linear and exponential functions matched to the A2 global temperature anomaly projections (Figure 3.3).

3.4 Discussion

The preceding examples illustrated the application of the analytical solution to forward model the subsurface thermal impacts of climate change. However, the analytical solution can also be inverted to estimate past climate change trends from borehole temperature profiles. Other researchers have demonstrated the application of inverted forms of other analytical solutions to the conduction or conduction–advection equation to
infer ground surface temperature history (e.g., Lachenbruch and Marshall 1986, Beltrami et al. 1995, Taniguchi 2006). In this process, initial conditions are assumed for the temperature profile at some time before the present, and the boundary condition is adjusted until the simulated present-day temperature profile concurs with the measured profile. Various formal inversion techniques have been utilized to obtain the best fit between the simulated and measured borehole data (e.g., Mareschal and Beltrami 1992, and references therein). These approaches have limitations because of the following: (1) the conduction-based solutions cannot be inverted to infer ground surface temperature in subsurface environments where advection is significant, and (2) the conduction–advection solution (i.e., Equation 3.2) that has been inverted restricts the surface temperature history to a linear function.

The analytical solution proposed in Equation 3.8 can be inverted without the two limitations noted earlier because it can accommodate groundwater flow and non-linear ground surface temperature changes. One question to be addressed in future research is how to establish the initial conditions for this inversion problem. Generally, for inversion purposes, geothermal (i.e., linear) temperature profiles are assumed for initial conditions. Following the approach of others, the $\delta$ term in Equation 3.4 can be set to zero to yield geothermal initial conditions for Equation 3.8. However, even in the absence of climate change, the initial conditions may deviate from the geothermal profile if groundwater flow is significant. A potential solution to this problem is to generate appropriate initial conditions by first running the solution for a longer period (e.g., 1000 years) with a constant boundary condition ($b=0$ in Equation 3.3) and the presumed groundwater flux. Equation 3.4 could then be fit to the results of such a simulation and used as the initial
conditions for the inversion of a present-day borehole temperature profile. The formal inversion of this solution would not be onerous as the new boundary condition (Equation 3.3) only contains one additional fitting parameter compared to the linear boundary condition of Taniguchi et al. (1999).

It should be noted that care must be employed when selecting an appropriate initial condition for the proposed solution for the purpose of forward modelling. If Equation 3.4 is used to generate an initial condition by simply minimizing the RMSE between the function and measured borehole temperature data, very high temperatures may result at great depth, and these temperatures will eventually be conducted upwards. This effect will be exacerbated in regions of discharge, as advective heat flux will propagate the unreasonably high temperatures from below. It is recommended that the generated initial conditions be plotted at depths much deeper than the region of interest to ensure that unreasonable results are not obtained. For this reason, results that extend far below measured borehole profiles should be considered with caution.

Finally, it should be recognized that the proposed analytical solution has inherent simplifications associated with the governing equation (e.g., one-dimensional water and heat flow, spatiotemporally constant groundwater velocity, no phase change in pore water, fully saturated conditions, and constant soil grain and soil water thermal properties). Additionally, the boundary condition cannot account for seasonal variations in temperature, and thus, the shallow temperature–depth profiles produced by the solution represent annual average values.
3.5 Conclusions

An analytical solution to the transient one-dimensional conduction–advection equation was developed that can better accommodate measured temperature–depth profiles for initial conditions and air temperature projections for boundary conditions. For example, field data from the Sendai Plain and Tokyo, Japan, were used to demonstrate the flexibility and improved accuracy of the initial conditions, and IPCC multi-GCM air temperature projections were used to demonstrate the improved accuracy of the boundary condition. The solution was verified with numerical methods and applied to investigate the effect of surface warming and groundwater flow in idealized subsurface environments. Results from these simulations indicated that groundwater recharge can exacerbate the rate of shallow subsurface warming, whereas groundwater discharge can transport heat from deeper geothermal zones and thereby accelerate deeper subsurface warming.

To compare potential differences between the proposed analytical solution and the solution by Taniguchi et al. (1999), simulations were performed using the temperature profiles from the Sendai Plain, Japan, as initial conditions and globally averaged air temperature anomalies from the IPCC as the boundary condition. The results indicated that the differences between the two solutions can be significant (e.g., 1–2°C) in both the shallow and deeper subsurface thermal regimes. The potential of this solution to be applied to infer past ground surface temperate changes has also been discussed.
References


CHAPTER 4: Potential Surface Temperature and Shallow Groundwater Temperature Response to Climate Change* 

Abstract

Global climate models project significant changes to air temperature and precipitation regimes in many regions of the Northern Hemisphere. These meteorological changes will have associated impacts to surface and shallow subsurface thermal regimes, which are of interest to practitioners and researchers in many disciplines of the natural sciences. For example, groundwater temperature is critical for providing and sustaining suitable thermal habitat for cold-water salmonids. To investigate the surface and subsurface thermal effects of atmospheric climate change, seven downscaled climate scenarios (2046–2065) for a small forested catchment in New Brunswick, Canada were employed to drive the surface energy and moisture flux model, ForHyM2. Results from these seven simulations indicate that climate change-induced increases in air temperature and changes in snow cover could increase summer surface temperatures (range −0.30 to +3.49°C, mean +1.49°C), but decrease winter surface temperatures (range −1.12 to +0.08°C, mean −0.53°C) compared to the reference period simulation. Thus, changes to the timing and duration of snow cover will likely decouple changes in mean annual air temperature (mean +2.11°C) and mean annual ground surface temperature (mean +1.06°C).

Projected surface temperature data were then used to drive an empirical surface to groundwater temperature transfer function developed from measured surface and groundwater temperature. Results from the empirical transfer function suggest that changes in groundwater temperature will exhibit seasonality at shallow depths (1.5 m), but be seasonally constant and approximately equivalent to the change in the mean annual surface temperature at deeper depths (8.75 m). The simulated increases in future groundwater temperature suggest that the thermal sensitivity of baseflow-dominated streams to decadal climate change may be greater than previous studies have indicated.

**Keywords:** ground surface temperature, climate change, groundwater temperature, baseflow-dominated rivers, snowpack evolution, thermal regime, river temperature

### 4.1 Introduction

#### 4.1.1 Drivers and importance of ground surface temperature

The impact of climate change on ground surface temperature (GST) is of interest to a diversity of scientific disciplines. For example, hydrologists are concerned with the influence of surface freezing and thawing on infiltration and runoff rates (Williams and Smith 1989), agricultural scientists have shown that seed germination is affected by surface and near-surface temperature (Mondoni et al. 2012), and geotechnical engineers have linked soil strength properties to surface/subsurface temperature (Andersland and Ladanyi 1994). Increased GST could also enhance decay rates and CO$_2$ release from soils and thereby act as a positive feedback mechanism to climate change (Eliasson et al. 2005). Potential effects of changes in winter GST include: altered nutrient concentrations
in soil water, enhanced winter root mortality, and decreased runoff quality (Mellander et al. 2007).

Increases in mean annual air temperature will not necessarily result in equivalent changes in mean annual GST (Mann and Schmidt 2003, Mellander et al. 2007, Zhang et al. 2005). Other physical processes must be considered to predict the associated increase in mean annual GST. For example, the duration of the snow-covered period is expected to decrease in northern latitudes due to increases in late fall and early spring air temperature (AT), which would alter the dynamics of atmosphere–surface heat exchange by reducing the insulating effect of the snowpack (Zhang 2005). A reduction in the snow-covered period would also lead to an increase in the amount of radiation absorbed by the ground (Bonan 2008). Furthermore, the length of the growing season is expected to increase with an upward shift in the AT regime. Under a deciduous canopy, an increase in early-spring and late-fall foliation/defoliation could affect both thermal and hydrological processes by altering the amount of net radiation and evapotranspiration at the land surface (Bonan 2008). The net effect of these positive and negative climate change feedback signals can be studied with a process-oriented model capable of simulating surficial thermal and hydrological processes.

Very few local-scale studies have investigated the impact of future climate change on GST regimes. Mellander et al. (2007) used two climate scenarios for 2091–2100, generated with the Hadley global climate model (GCM) and downscaled with a regional climate model (RCM), to predict changes to soil temperature in northern Sweden. They simulated a decrease in the period of persistent snowpack of 73–93 days, an increase in annual soil temperatures of 0.9–1.5°C at 10 cm depth, and an advance in spring soil
warming of 15–19 days. Salzmann et al. (2007) used the data from ten RCM-generated
and two incremental climate scenarios to drive the surface energy balance model TEBAL
and predicted a potential range of increased GST (mean +3.5°C) for a mountainous
permafrost region in Switzerland. Each of their RCM simulations were driven with the
HadAm3H GCM forced by the A2 or B2 emission scenarios. They suggested that their
study should be expanded by using multiple GCMs.

4.1.2 Drivers and importance of shallow groundwater temperature

The subsurface thermal regime is driven by water and energy fluxes across the
ground surface and the geothermal flux from the Earth’s interior. Seasonal and decadal
variations in GST can be propagated downwards via conduction and advection and
thereby perturb the temperature of shallow (i.e., <10 m) groundwater (Taylor and Stefan
2009). Thus, atmospheric climate change may result in changes to seasonal and mean
annual GST (as previously discussed), which could translate to shifts in the timing and
magnitude of the seasonal groundwater temperature cycle.

Because groundwater temperature is less variable than surface water temperature,
groundwater discharge provides a thermal buffer for riverine systems during the summer
and winter months (Caissie 2006, Hayashi and Rosenberry 2002, Webb et al. 2008). In
eastern Canada, summer river temperatures are approaching the critical threshold for
salmonids (Breau et al. 2007, 2011, Swansburg et al. 2004). Discrete cold-water plumes
formed by groundwater–surface water interactions have been shown to provide critical
thermal relief for stressed cold-water fishes during high temperature events (Breau et al.
warsms, cold-water fishes may become increasingly dependent on these groundwater-sourced thermal refugia.

Recently, there have been a number of publications investigating the thermal sensitivity of rivers to environmental conditions (Kelleher et al. 2012, Mayer 2012, O'Driscoll and DeWalle 2006, Tague et al. 2007). Many of these studies have demonstrated that groundwater-dominated streams are less sensitive to AT variability on a seasonal basis; however, the response of groundwater temperature (and consequently the temperature of baseflow-dominated streams) to decadal climate change has not been well-studied. For example, Chu et al. (2008) stated ‘climate change induced differences in precipitation and temperature that may influence the magnitude and timing of groundwater discharge should be addressed in future analyses [of stream temperature and fish habitat]’. Mayer (2012) acknowledged the dearth of information regarding the response of groundwater temperature to climate change and suggested that it posed a challenge for modelling future river temperatures.

Several researchers have attempted to address the relationship between atmospheric climate change and shallow groundwater temperature. Taylor and Stefan (2009) employed an analytical solution to a simple conduction equation with a sinusoidal boundary condition to infer that the change in the mean annual groundwater temperature would be equivalent to the change in mean annual GST for Minnesota, USA. They suggested that the current seasonal amplitude of the groundwater temperature cycle would be relatively unchanged in a warmer climate. Gunawardhana et al. (2011), Gunawardhana and Kazama (2011), and Kurylyk and MacQuarrie (2013a) employed analytical solutions to a one-dimensional conduction-advection equation with an
increasing GST to investigate the subsurface response to climate change in the Sendai Plain, Japan. This type of boundary condition does not allow for an investigation of seasonal trends in groundwater temperature. Gunawardhana and Kazama (2012) also used downscaled data from 15 GCMs for the Sendai Plain to drive a numerical model (VS2DH) of groundwater flow and heat transport to investigate subsurface thermal perturbations due to climate change. These simulations were performed on a coarse temporal resolution (1 year) and also did not address seasonal effects. Others (e.g., Bense et al. 2009, Ge et al. 2011) have simulated the impact of rising surface temperatures on permafrost degradation, but the focus of these contributions was primarily the hydraulic evolution of supra-permafrost aquifers, rather than the thermal evolution.

4.1.3 Research objectives

The objective of this contribution is to provide answers to the following scientific questions:

1. Will changes in mean annual and seasonal GST closely mimic climate change-induced changes in mean annual and seasonal AT?

2. Can monthly GST data be utilized to predict monthly groundwater temperature in shallow aquifers by adopting an empirical approach?

3. How will shallow groundwater temperature respond to climate change on a monthly basis and what are the implications for salmonid thermal refugia?

These questions will be answered in reference to a small forested catchment with available field data (AT, GST, and groundwater temperature), and in which cold
groundwater discharge has been observed to provide thermal refuge for salmonids. Thus, the answers to these scientific questions will primarily be valid for our particular study site. This work differs from previous future GST studies (Question 1) by employing multiple GCMs and by investigating climate change impacts on surface processes in a warmer climate, albeit still with seasonal snow cover. The answer to the second question will provide a simplistic alternative to implementing analytical solutions, which are replete with restrictions, particularly for the GST boundary condition. Finally, the answer to the third question should provide surface water hydrologists and ecologists with valuable information concerning the thermal sensitivity of shallow aquifers and baseflow-dominated rivers and the resilience or sensitivity of cold-water fish habitat to a warming climate.

4.2 Study Site

The geographic location selected for this study is the Catamaran Brook catchment (~53 km$^2$) in east-central NB, Canada (46°52’N latitude, 66°06’W longitude). Catamaran Brook is a third-order tributary of the Little Southwest Miramichi River (Figure 4.1). The Catamaran Brook catchment has a mixed coniferous (65%) and deciduous (35%) Acadian forest cover (Cunjak et al. 1990). Portions of the catchment were clear-cut in 1996 (Alexander 2006). The annual average precipitation in the region is 1140 mm, with approximately 33% falling as snow (Cunjak et al. 1993). The region experiences a humid continental climate with arid, cold winters and warm, humid summers (Cunjak et al. 1990).
As indicated in Figure 4.2, the Catamaran Brook catchment is covered with a thin blanket of coarse sandy till, which is underlain by a dense clay till and Silurian and Devonian bedrock (Alexander 2006). The water table is shallow and lies within the surficial sand and gravel deposit (Alexander 2006). Discharge in Catamaran Brook remains primarily groundwater-sourced throughout the year (Caissie et al. 1996). Thus, the summer water temperature of Catamaran Brook is generally considerably lower than that of the main stem of the Little Southwest Miramichi River, which has a high width to depth ratio and responds rapidly to solar radiation (Caissie et al. 2007). Juvenile Atlantic salmon (*Salmo salar*) have been documented seeking thermal refuge in the cold-water plume at the mouth of Catamaran Brook during high temperature events (Breau et al. 2011). During a heat wave in 2010, very dense fish aggregations (6000-10,000 fish) were
observed in nearby thermal refugia within the Little Southwest Miramichi River (T. Linnansaari, personal communication, 2012).

![Diagram of the Catamaran Brook catchment](image)

Figure 4.2. Cross-sectional view of the portion of the Catamaran Brook catchment containing the monitoring well and the temperature loggers (adapted from Alexander 2006).

4.3 Methods

4.3.1 Reference period and projected climate data

Future climate scenarios are generated by GCMs driven by emission scenarios that invoke assumptions about future population growth, technology, and economic development (Nakicenovic and Swart 2000). GCM simulations are performed on coarse computational grids (e.g., 250 km by 250 km), thus their results should be downscaled to translate the data from the coarse grid to local climate conditions. Downscaling approaches have been thoroughly reviewed in past contributions (e.g., Wilby and Wigley...
A simple downscaling approach is the daily translation (DT) method, which is in the family of ‘statistical’ or ‘quantile-quantile mapping’ downscaling techniques (Teutschbein and Seibert 2012). Statistical downscaling is predicated on the assumption that local climate conditions can be determined from large-scale climate variables using linear or non-linear transfer functions (Jeong et al. 2012a). In the DT method, a GCM is initially run for a reference period containing local observations. Scaling factors for precipitation and AT are then determined from the reference period simulations and local observations using empirical cumulative distribution functions. GCM simulations for a future time period (emission scenario) are then downscaled by applying the daily scaling factors.

Many more complex statistical downscaling methods have been developed; one of these is the hybrid multivariate linear regression (HMLR) method (Jeong et al. 2012a, 2012b). In this method, the local climate variables are obtained from the GCM simulations using multiple regression functions determined from reference period simulations. Because regression-based methods often have difficulty producing the observed variability in local climate predictands, a stochastic generator is used to increase the variance in the datasets. These two methods (i.e., DT and HMLR) were employed in this study.

Output from GCMs can also be dynamically downscaled with RCMs. RCMs produce results on a finer computational grid than GCMs, using GCM output as boundary values. However, RCMs tend to introduce additional biases (Wilby et al. 2000). For this reason, results from RCMs are often bias-corrected. This can be accomplished through techniques similar to those used for downscaling. In the present study, both of the
dynamically downscaled climate series (CRCM 4.2.3 aev-A2 and CRCM 4.2.3 agx-A2, Table 4.1) were bias-corrected using the DT method addressed earlier. This present study is part of a multi-disciplinary initiative investigating salmonid thermal refugia and their sensitivity to climate change. The climate scenarios utilized in this study (Table 4.1) were provided by collaborating climatologists. The HMLR-downscaled data were contributed by the Université du Québec Institut National de la Recherche Scientifique (D. Jeong, INRS, personal communication, 2011), while the other climate data series were produced from the third Coupled Model Inter-comparison Project database of GCM output (CMIP3, Meehl et al. 2007) and dynamically downscaled using the Canadian Regional Climate Model (CRCM4.2.3, de Elia et al. 2008, Huard 2011) or statistically downscaled with the DT method (Huard 2011). In total, seven projected climate scenarios (Table 4.1) were produced for the period of 2046–2065 using six GCMs, one RCM, two downscaling methods, and three emission scenarios. This period was selected for the present study due to the number of available downscaled climate scenarios. These particular climate scenarios were selected because they span the range of plausible future climatic conditions for east-central New Brunswick. These climate data provide the basis for predicting future GST and groundwater temperature. The term ‘Run ID’ (Table 4.1) refers to a particular simulation performed within the RCM. In this case, the primary difference between the two RCM runs (Agx and Aev) is the GCM driver (CGCM3 and Echam5).
Table 4.1. Details for each of the climate scenarios employed in the current study.

<table>
<thead>
<tr>
<th>GCM (RCM)</th>
<th>Resolution</th>
<th>Run ID</th>
<th>Emission Scenario</th>
<th>Downscaling Approach</th>
<th>Contributor (Reference)</th>
</tr>
</thead>
<tbody>
<tr>
<td>CGCM3</td>
<td>3.75 x 3.71°</td>
<td>-</td>
<td>A2</td>
<td>Statistical-HMLR</td>
<td>INRS (Jeong et al. 2012a)</td>
</tr>
<tr>
<td>CGCM3 (CRCM4.2.3)</td>
<td>3.75 x 3.71° (45x45 km)</td>
<td>Aev</td>
<td>A2</td>
<td>Dynamical</td>
<td>Ouranos (Huard 2011)</td>
</tr>
<tr>
<td>ECHAM5 (CRCM4.2.3)</td>
<td>1.88 x 1.87° (45x45 km)</td>
<td>Agx</td>
<td>A2</td>
<td>Dynamical</td>
<td>Ouranos (Huard 2011)</td>
</tr>
<tr>
<td>CSIRO Mk3.0</td>
<td>1.9 x 1.9°</td>
<td>-</td>
<td>B1</td>
<td>Statistical-DT</td>
<td>Ouranos (Huard 2011)</td>
</tr>
<tr>
<td>CSIRO Mk3.5</td>
<td>1.9 x 1.9°</td>
<td>-</td>
<td>B1</td>
<td>Statistical-DT</td>
<td>Ouranos (Huard 2011)</td>
</tr>
<tr>
<td>MIROC 3.2 HIRES</td>
<td>1.1 x 1.1°</td>
<td>-</td>
<td>A1B</td>
<td>Statistical-DT</td>
<td>Ouranos (Huard 2011)</td>
</tr>
<tr>
<td>CGCM3</td>
<td>3.75 x 3.71°</td>
<td>-</td>
<td>A1B</td>
<td>Statistical-HMLR</td>
<td>INRS (Jeong et al. 2012a)</td>
</tr>
</tbody>
</table>

Figure 4.3 provides the projected changes in mean annual precipitation and mean annual AT for each of the seven scenarios listed in Table 4.1. The observed/reference climate data (1961–2000) were taken from the Environment Canada (EC) database of adjusted and homogenised Canadian climate data (EC 2013a). All of the scenarios predict a rise in mean annual AT (with range of +0.4 to +3.9°C for the 2046–2065 period compared to the 1961–2000 period), but the projections for annual precipitation vary significantly in magnitude and direction (−12 to +49%). It should be noted that the intent of this contribution is not to provide an in-depth analysis of the climate data but rather to drive the models with multiple plausible climate scenarios produced with a variety of established methods. Thus, the intent is to examine the sensitivity of GST and groundwater temperature to external forcing rather than to make an assessment of the accuracy of predictions regarding future surface and subsurface thermal regimes.
Figure 4.3. Projected changes in average annual precipitation vs. projected changes in mean annual AT for east-central New Brunswick. The climate data from the observed period (1961-2000) were chosen for a baseline for which to compare the future period (2046-2065) simulations. The 1961-2000 observed data series and each projected climate series (2046-2065) have their own distinct set of statistics (e.g., annual mean and standard deviation).

4.3.2 Physically based model of surface temperature

The physically based surface energy balance model ForHyM2 (Forest Hydrology Model v.2, Arp and Yin 1992, Yin and Arp 1993) was selected to simulate daily GST from the reference and projected climate data. This physically based model is capable of simulating the complex relationship between future climate change and the lower atmosphere–surface energy exchange. ForHyM was originally developed for simulating water fluxes through shallow forest soils; this model successfully reproduced data from a deciduous forest in Ontario, Canada, and a coniferous forest in Quebec, Canada (Arp and Yin 1992). Yin and Arp (1993) later developed ForSTeM, to simulate soil temperature. ForSTeM was created to be coupled to ForHyM, and these two models with subsequent revisions are collectively referred to as ForHyM2. ForHyM2 has been applied at many
other sites (e.g., Balland et al. 2006, Bhatti et al. 2000, Houle et al. 2002, Meng et al. 1995, Oja et al. 1995) to simulate hydrologic fluxes and/or soil temperatures; at all of these sites, model simulations closely agreed with field observations. ForHyM2 has also performed well at simulating snowpack depths under various forest canopies (Balland et al. 2006, Houle et al. 2002).

ForHyM2 requires daily AT, precipitation, and relatively few site characteristics for input conditions. Thus, simulations can be performed without extensive field work or site parameterisation. GST simulations can be started without initial conditions, provided that the first time step is not during a period with snow on the ground. ForHyM2 simulations were conducted in a simplified manner (one-dimensional) with the entire catchment represented as a point; thus surface heterogeneities were ignored. The observed climate data and downscaled projected climate data were utilized to drive the ForHyM2 simulations on a daily time step. The GST was taken as the temperature of the forest floor surface. A more detailed description of the ForHyM2 model mechanics and input parameters is included in Appendix 2.

4.3.3 Comparison of ForHyM2 simulations with measured GST

To test the accuracy of performing uncalibrated GST simulations in ForHyM2, simulated GST results were generated for a reference period containing GST measurements. Alexander (2006) installed temperature loggers (VEMCO Minilog) in the Catamaran Brook catchment to record hourly AT, GST, and groundwater temperature (see temperature loggers, Figure 4.2). AT observations were recorded in shelters at a height of 1.5m above the ground surface. Groundwater temperature data were recorded in
the monitoring well indicated in Figure 4.1 at four depths: 1.5, 2.75, 5.25, and 8.75m (Figure 4.2). GST data were recorded near the monitoring well. The temperature data utilized in the present study were collected during 1 October 2001–30 September 2003. Additionally, measured precipitation data from the EC historical weather database for the Miramichi weather station (EC 2013b) were used as input for the ForHyM2 simulations. In general, the ForHyM2-simulated GST time series were in agreement with the GST observations (Figure 4.4). The associated coefficient of determination ($R^2$) was 0.98, thus ForHyM2 was judged to have performed favourably for these site conditions and meteorological data considering that no model calibration was undertaken.

Figure 4.4. Measured and ForHyM2-simulated GST and ForHyM2-simulated snowpack for the Catamaran Brook catchment for the October 2001-September 2003 period. GST data are relatively constant during the simulated snow-covered period.
4.3.4 Empirically-based estimation of groundwater temperature

Several previous studies investigating the impact of climate change on groundwater temperature have employed an analytical solution to the following governing equation for one-dimensional heat transport in regions of significant groundwater flow (Domenico and Schwartz 1990):

\[
\lambda \frac{\partial^2 T}{\partial z^2} - q C_w \frac{\partial T}{\partial z} = C \frac{\partial T}{\partial t}
\]

(4.1)

where \( \lambda \) is the thermal conductivity of the subsurface \((\text{M} \ \text{L} \ \text{t}^3 \ \text{T}^{-1})\), \( T \) is the spatiotemporally varying groundwater temperature, \( q \) is the vertical Darcy flux \((\text{L} \ \text{t}^{-1})\), \( C \) and \( C_w \) are the volumetric heat capacities of the medium and water, respectively \((\text{M} \ \text{L}^{-1} \ \text{t}^{-2} \ \text{T}^{-1})\), \( z \) is depth \((\text{L})\), and \( t \) is time. Analytical solutions to Equation 4.1 have been derived for GST that is linearly or exponentially increasing on a decadal basis (Kurylyk and MacQuarrie 2013a, Taniguchi et al. 1999) or periodically varying on a seasonal basis (Goto et al. 2005, Stallman 1965). However, these boundary conditions are inappropriate for the present study. A linearly or exponentially increasing GST boundary condition ignores seasonal variations in temperature. Furthermore, a periodically varying (i.e., sinusoidal) GST boundary condition is a poor approximation of the annual GST cycle in seasonally snow-covered catchments due to the insulating effect of the winter snowpack (Lapham 1989, Zhang 2005).

There are many other limitations associated with Equation 4.1, including spatiotemporally constant groundwater flux and thermally homogeneous subsurface properties. Taylor and Stefan (2009) utilized a solution to a simplified form of Equation 4.1 that ignored advective heat transport due to groundwater flow. However, advective
heat transport could be significant in the Catamaran Brook basin, particularly in the shallow aquifer. The subsurface heat transport modelling capabilities in ForHyM2 (Appendix 2) were also not utilized for the present study because the model equations ignore advective heat transport due to groundwater flow.

In light of the limitations associated with ForHyM2 and previously published analytical solutions, an empirical GST-to-groundwater temperature transfer function was adopted for simulating the monthly groundwater temperature response to rising GST:

$$ GWT_i = MAGST + D\left(GST_{(i-L)} - MAGST\right) + B $$

where $MAGST$ is the mean annual GST, $GWT_i$ is the groundwater temperature for month $i$, $GST_{(i-L)}$ is the GST for month $(i-L)$, and $D$, $L$, and $B$ are empirical parameters that are temporally constant, but depth-dependent. Although this equation is empirically based, the equation parameters relate to physical processes. Analytical solutions to the transient conduction equation have demonstrated that the seasonal groundwater temperature cycle is damped and lagged with respect to the seasonal GST cycle. These damping and lagging effects increase with depth (Bonan 2008, p. 134, Taylor and Stefan 2009). The unitless $D$ parameter produces the damping effect of the subsurface thermal diffusivity, while the $L$ parameter (units of months) produces the lagging effect between a GST change and its subsurface realization. The $B$ term ($^\circ\text{C}$) accounts for shallow heat transfer phenomena that may not be included in the other two parameters, which are conduction based. These phenomena include: latent heat effects due to freezing and thawing, vadose zone heat transfer processes in the vapour phase, and groundwater advection.
There are admittedly limitations to adopting an empirical approach for relating GST and groundwater temperature. However, linear and non-linear regression-based AT-stream temperature transfer functions have been developed for examining the response of surface water temperature to future warming climates (Kelleher et al. 2012, Mayer 2012, Mohseni et al. 2003, and references therein). Furthermore, GST is the actual driver for shallow groundwater temperature, whereas AT is merely used as a surrogate for radiation, the primary driver of river temperature (Allen and Castillo 2007). Thus, the application of this GST to groundwater temperature transfer function should be at least as insightful as its surface water counterparts.

The measured groundwater temperature and GST collected by Alexander (2006) were used to estimate the depth-dependent values of $D$, $L$, and $B$ by minimising the root-mean-square-error (RMSE) between the measured and simulated monthly groundwater temperatures for each depth. It should be noted that two of the data loggers were installed in the sand and gravel aquifer, while the other two were installed in the clay aquitard (Figure 4.2). Future projections of groundwater temperature at each depth were obtained by driving the GST-groundwater temperature transfer function with the future projections of GST obtained from ForHyM2. The flow of data from the GCMs to the GST-groundwater temperature transfer function is illustrated in Figure 4.5.
4.4 Results

4.4.1 Projected climate change impacts on GST and snow cover

The most pronounced increase (2.64°C) in mean annual GST, compared to the reference period simulated GST (1961–2000), was for the MIROC 3.2 HIRES-A1B climate data, while the only projected decrease in the mean annual GST (−0.15°C) was obtained for the CSIRO Mk3.0-B1 climate data (Figure 4.6). It should be noted that changes in mean annual AT for these particular scenarios were +3.96 and +0.39°C, respectively (Figure 4.3). Thus, it appears that the projected increases in mean annual AT are damped at the ground surface.
**Figure 4.6. ForHyM2-simulated mean annual GST for each climate scenario (2046-2065).** Simulated mean annual GST based on the historical climate data (1961-2000) is indicated by the data point to the far left. Error bars indicate one standard deviation in mean annual GST.

In addition to the projected variations in mean annual GST, the ForHyM2 results also suggest that changes will occur to seasonal GST. As indicated in Figures 4.6 and 4.7, the trends in seasonal GST do not necessarily reflect the trends in mean annual GST. For example, the projected changes in mean annual GST are generally positive, whereas the trends in seasonal GST are typically positive for the spring, summer, and fall, but negative for the winter. Additionally, Table 4.2 indicates that the changes in seasonal GST will not necessarily closely follow the changes in seasonal AT.
Figure 4.7. ForHyM2-simulated average seasonal GST for each climate scenario (2046-2065) and the reference period (1961-2000) for (a) winter, (b) spring, (c) summer, and (d) fall. The y axes show various ranges in GST but are drawn to the same scale. Winter = December-February, spring = March-May, summer = June-August, and fall = September-November.

Table 4.2. Changes in seasonal AT and ForHyM2-simulated GST for each scenario.

<table>
<thead>
<tr>
<th>Climate Scenario</th>
<th>Winter (°C)(^1)</th>
<th>Spring (°C)</th>
<th>Summer (°C)</th>
<th>Fall (°C)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>∆AT</td>
<td>∆GST</td>
<td>∆AT</td>
<td>∆GST</td>
</tr>
<tr>
<td>CGCM3-A2</td>
<td>3.31</td>
<td>-0.33</td>
<td>2.07</td>
<td>1.21</td>
</tr>
<tr>
<td>CRCM 4.2.3 aev-A2</td>
<td>4.34</td>
<td>-0.63</td>
<td>2.40</td>
<td>1.65</td>
</tr>
<tr>
<td>CRCM 4.2.3 agx-A2</td>
<td>2.05</td>
<td>-0.77</td>
<td>1.19</td>
<td>0.86</td>
</tr>
<tr>
<td>CSIRO MK3.0 -B1</td>
<td>1.37</td>
<td>-0.71</td>
<td>0.32</td>
<td>0.28</td>
</tr>
<tr>
<td>CSIRO MK3.5-B1</td>
<td>1.81</td>
<td>-1.12</td>
<td>1.73</td>
<td>1.19</td>
</tr>
<tr>
<td>MIROC 3.2-HIRES-A1B</td>
<td>4.58</td>
<td>0.08</td>
<td>4.16</td>
<td>3.54</td>
</tr>
<tr>
<td>CGCM3-A1B</td>
<td>3.80</td>
<td>-0.21</td>
<td>2.05</td>
<td>1.29</td>
</tr>
</tbody>
</table>

\(^1\)Winter = Dec-Feb, Spring = Mar-May, Summer = Jun-Aug, and Fall = Sep-Nov
Figure 4.8 indicates that the snow-covered period is expected to decrease under the various climate scenarios. The predicted reduction in the average snow-covered period could potentially range from 13 (CSIRO Mk3.0-B1) to 49 days (CRCM 4.2.3 aev-A2). Additionally, Figure 4.8 suggests that the mean annual snowpack depth will also likely decrease. Estimated reductions in mean annual snowpack depth range from 35% (CSIRO Mk3.0-B1) to 77% (CRCM 4.2.3 aev-A2). The simulated snowpack depths for the other five climate scenarios are within the range of the CSIRO Mk3.0-B1 and CRCM 4.2.3 aev-A2 results.

4.4.2 GST to groundwater temperature transfer function

The best fits for the $L$, $D$, and $B$ parameters in the GST-groundwater temperature transfer function for each of the four depths are presented in Table 4.3. Figure 4.9 gives plots for the measured and simulated monthly groundwater temperature at each of the four depths. The resultant correlation coefficient (R value) and RMSE between the measured and simulated data are also indicated.
Figure 4.8. ForHyM2-simulated snowpack depth (averaged over the duration of the simulation period) vs. the date. Simulated snowpack depths for the other five climate scenarios are within the range set by the CSIRO Mk3.0B1 and CRCM 4.2.3 aev-A2 results.

Table 4.3. Depth-dependent parameter values for the GST to groundwater temperature transfer function

<table>
<thead>
<tr>
<th>Depth (m)</th>
<th>Lag $L$ (months)</th>
<th>Damping Term $D$</th>
<th>Empirical $B$ ($\degree$C)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.50</td>
<td>1.29</td>
<td>0.467</td>
<td>1.575</td>
</tr>
<tr>
<td>2.75</td>
<td>1.81</td>
<td>0.308</td>
<td>1.237</td>
</tr>
<tr>
<td>5.25</td>
<td>3.30</td>
<td>0.127</td>
<td>0.571</td>
</tr>
<tr>
<td>8.75</td>
<td>5.70</td>
<td>0.045</td>
<td>0.274</td>
</tr>
</tbody>
</table>
Figure 4.9. Comparison of observed and simulated average monthly groundwater temperature at different depths: (a) 1.5 m, (b) 2.75 m, (c) 5.25 m, and (d) 8.75 m. The $y$ axes are not all drawn to the same scale.

According to the analytical solution employed by Taylor and Stefan (2009), the lag term ($L$) and the natural logarithm of the damping term ($D$) should be linearly related to the depth. Figure 4.10 indicates that the relationship between the depth and the $L$ term or the natural logarithm of the $D$ term can be reasonably approximated by a line with a zero intercept.
Figure 4.10. The lag factor ($L$) and natural logarithm of the damping factor ($D$) plotted vs. depth ($z$). The $L$ and $D$ parameters were obtained from fitting Equation 4.2 to measured groundwater temperature.

4.4.3 Projected climate change impacts on groundwater temperature

Figure 4.11 presents the simulated groundwater temperature for the reference and future periods. For the sake of clarity, only the groundwater temperature results from the reference period simulation (1961–2000) and the maximum (MIROC 3.2 HIRES-A1B) and minimum (CSIRO-Mk3.0-B1) simulated groundwater temperature are shown. The monthly groundwater temperatures simulated from the MIROC 3.2 HIRES-A1B climate data exceed the reference period groundwater temperature by approximately 1–3.5°C depending on the month and depth, whereas the groundwater temperature simulated from the CSIRO-Mk3.0-B1 climate data is relatively unchanged from the reference period.
Figure 4.11. Results from the empirical GST-groundwater temperature transfer function driven by the ForHyM2-produced GST for the reference period and two projected climate scenarios for depths of (a) 1.5 m, (b) 2.75 m, (c) 5.25 m, and (d) 8.75 m. Of the seven climate scenarios, the CSIRO Mk3.0-B1 and the MIROC 3.2-HIRES-A1B climate scenarios consistently produced the lowest and highest monthly groundwater temperature data, respectively.

Figure 4.12 shows the change in average monthly groundwater temperature projected for each of the climate scenarios. As expected, the variability in the changes of monthly groundwater temperature decreases with depth for each of the climate scenarios.
Figure 4.12. Changes in monthly groundwater temperature produced by driving the GST-groundwater temperature transfer function with the projected GST data simulated for each of the seven climate scenarios (change = the simulated average monthly groundwater temperature for 2046-2065 minus the simulated average monthly groundwater temperature for the 1961-2000 period) for each depth.

4.5 Discussion

4.5.1 Relationship between climate-induced changes in AT and GST

Figures 4.3 and 4.6 demonstrate that the magnitude of the increases in projected mean annual AT for the seven climate scenarios for this study site (+0.4 to +3.9°C) are larger than the simulated changes in mean annual GST (−0.15 to +2.64°C). Thus, the findings of the present study concur with Mann and Schmidt (2003), who proposed the debated notion that decadal mean annual GST changes will not necessarily track mean
annual AT changes (see Chapman et al. 2004, Schmidt and Mann 2004). Decadal snow-cover evolution can decouple mean annual GST and mean annual AT trends by altering the winter thermal resistance between the lower atmosphere and the ground surface.

Figure 4.4 indicates that the variability in simulated and measured GST is considerably less in winter than in the other seasons due to the insulating effect of the snowpack. This effect is simulated in ForHyM2 by the inclusion of a thermal resistance layer during snow-covered periods (Appendix 2). Figure 4.8 suggests that the snowpack insulating effect will likely be reduced in the coming decades due to a reduction in the snow-covered period and the average snowpack depth. Thus, even when snow cover exists in a warmer climate, the equivalent thermal resistance will be limited by snowpack thinning. This will increase the winter heat transfer between the lower atmosphere and the ground surface and result in colder winter GST as indicated in Figure 4.7a. This decrease in simulated winter GST limits increases in mean annual GST and may be an important negative climate change feedback mechanism for the subsurface thermal regime.

Table 4.2 indicates that changes in average summer GST (−0.30 to +3.49°C) will closely replicate changes in average summer AT (−0.30 to +3.49°C) for this location. Table 4.2 also suggests that increases in fall GST will be slightly less than increases in fall AT (average fall ratio, ΔGST/ΔAT = 93%) and that increases in spring AT will be noticeably damped at the ground surface (average spring ratio, ΔGST/ΔAT = 72%). The spring damping effect is caused by a lingering snowpack. Thus, changes in seasonal GST will likely follow changes in seasonal AT during the warmer period, but not during the colder period.
4.5.2 The empirical GST to groundwater temperature function

The results presented in Figure 4.9 indicate that an empirical relationship of the form of Equation 4.2 is flexible enough to reproduce groundwater temperature at coarse spatial (~2 m) and temporal (monthly) resolutions. As expected, the RMSE values were higher at shallow depths where the seasonal variability in groundwater temperature is greater. However, the relatively constant correlation coefficients (R values) suggest that the function is consistent in addressing the variability in measured groundwater temperature to depths of approximately 9 m. Figure 4.9 suggests that the function has a problem with simulating groundwater temperature in April. We expect that this error arises due to advective heat transport associated with the spring snowmelt and major recharge event during this period and/or with latent heat absorbed during the thawing of the upper few cm of soil.

The coefficients of determination ($R^2$) in Figure 4.10 illustrate that the lag term ($L$) and the damping factor ($D$) generally varied with depth as expected. The slope of the natural logarithm of the damping factor vs. depth relationship can be used to infer a soil thermal diffusivity of $7.1 \times 10^{-7}$ m$^2$ s$^{-1}$ (see Equation 4 in Taylor and Stefan 2009), which is in agreement with the typical saturated soil thermal diffusivity ($7.0 \times 10^{-7}$ m$^2$ s$^{-1}$) reported by Bonan (2008). Thus, the behaviour of the $D$ and $L$ terms concurred with our expectations based on classic heat conduction theory. As discussed, $B$ is an encompassing parameter to account for any heat transfer process other than saturated zone conduction. Near-surface phenomena (e.g., freezing, evaporative losses to the atmosphere, thermal conductivity heterogeneities due to variable saturation) and groundwater flow can significantly perturb a temperature-depth profile. These phenomena exhibit seasonality,
which would contribute to the errors in the estimated groundwater temperature for particular months. Table 4.3 indicates that the values of $B$ decreased with increasing depth, which we expected given our proposition that $B$ accounts for near-surface phenomena and advective heat transport. Clearly, the effects of near-surface phenomenon would decrease with depth. Advective heat transport would also decrease with depth given the presence of the clay aquitard (Figure 4.2). In general, the behaviour of the equation parameters (i.e., $D$, $L$, and $B$) is suggestive that they produce the physical effects that we postulated.

### 4.5.3 Shallow groundwater temperature response to climate change

Figures 4.11 and 4.12 indicate that the MIROC 3.2 HiresA1B and CSIRO Mk3.0-B1 climate data resulted in the highest and lowest groundwater temperatures, respectively. These results were consistent for each month and depth. This is not surprising given that these two climate scenarios also resulted in the highest and lowest mean annual AT (Figure 4.3) and mean annual GST (Figure 4.6). Thus in general, increased mean annual AT will result in increased mean annual GST and mean annual groundwater temperature; however, the response of monthly groundwater temperature to rising AT (and consequently rising GST) may be complex. For example, Figure 4.12 indicates that at 1.5 m depth the maximum changes in the groundwater temperature for the MIROC 3.2 Hires-A1B scenario (+3.43°C) will occur in May, and the minimum changes in groundwater temperature (+1.26°C) will occur in March. Thus, climate change-induced increases in shallow groundwater temperature in the Catamaran Brook catchment will likely exhibit seasonality. Figure 4.12 demonstrates that this seasonality
will decrease with depth. At a depth of 8.75m, the increase in monthly groundwater temperature is approximately constant and equivalent to the rise in mean annual GST.

This study has also demonstrated that, due to increased summer GST and decreased winter GST (Figure 4.7), the range in the annual cycle of groundwater temperature will increase. For example, the difference between the maximum (September) and minimum (February) average monthly groundwater temperature at a depth of 1.5m was 13.8°C for the reference period simulation, and the differences projected for the seven climate scenarios ranged from 14.0 to 17.2°C. These findings contrast with those of Taylor and Stefan (2009) who proposed that the amplitude of shallow groundwater temperature would not be significantly affected by a warming climate.

One motivation for this contribution was to examine the effect of rising groundwater temperature on groundwater-sourced salmonid thermal refugia. Stream temperature researchers have defined the thermal sensitivity of a stream as the slope of the AT–stream temperature relationship (e.g., Mayer, 2012). Following this approach, we define the subsurface thermal sensitivity as the ratio of the climate change-induced increase in summer (June–August) groundwater temperature to the increase in summer AT. We are primarily concerned with summer temperatures as groundwater discharge is most critical for the preservation of salmonid populations during this time of the year (Breau et al. 2007, Cunjak et al. 2005). Table 4.4 presents the simulated subsurface thermal sensitivities for each climate scenario at each of the four depths. The sensitivity values are negatively correlated with depth because simulated summer GST increases closely followed summer AT increases (Table 4.2), but simulated changes in mean annual
GST were significantly damped compared to changes in mean annual AT. With increased depth, the subsurface thermal effects of seasonal GST are reduced, and the mean annual GST becomes the primary driver of groundwater temperature.

The mean summer subsurface thermal sensitivity values for Catamaran Brook (0.66–0.88, Table 4.4) are somewhat higher than stream temperature sensitivities obtained by previous researchers examining AT–surface water temperature relationships for shorter periods (Kelleher et al. 2012, Mayer 2012, O’Driscoll and DeWalle 2006). For example, Kelleher et al. (2012) found that the average thermal sensitivity of smaller streams (0.52) was less than that of larger (stream order > 3) streams (0.83) for stream systems in Pennsylvania, USA. Mayer (2012) used weekly AT and stream temperature data and found that the summer thermal sensitivities varied from 0.25 to 0.58 for six streams in Oregon, USA. Kelleher et al. (2012, and references therein) proposed that thermal sensitivities of stream water temperature to AT variations derived from high-frequency temperature data (e.g., seasonal, monthly, weekly, or daily) could be employed to estimate stream temperature sensitivity to climate change. This approach may not be valid in groundwater-dominated streams, because it is fundamentally based on the premise that groundwater temperature will respond in the same manner to both high-frequency and low-frequency AT variations. Groundwater temperature may be insensitive to daily or even seasonal AT variations, but it is sensitive to decadal AT variations, as we have demonstrated by the subsurface thermal sensitivity values in Table 4.4. Thus extrapolated seasonally derived thermal sensitivities obtained for baseflow-dominated streams may underestimate their true thermal sensitivity to decadal climate change.
Table 4.4. Simulated summer subsurface thermal sensitivities for each depth averaged for the seven climate scenarios

<table>
<thead>
<tr>
<th>Depth (m)</th>
<th>Average ΔGWT$_s^1$</th>
<th>Average Summer Subsurface Thermal Sensitivity (STS)$^2$</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.5</td>
<td>1.31°C</td>
<td>0.88</td>
</tr>
<tr>
<td>2.75</td>
<td>1.26°C</td>
<td>0.81</td>
</tr>
<tr>
<td>5.25</td>
<td>1.07°C</td>
<td>0.72</td>
</tr>
<tr>
<td>8.75</td>
<td>0.98°C</td>
<td>0.66</td>
</tr>
</tbody>
</table>

$^1$ΔGWT$_s$ refers to the average change in summer groundwater temperature (°C)

$^2$STS refers to the summer subsurface thermal sensitivity: STS = ΔGWT$_s$/ΔAT$_s^3$

$^3$ΔAT$_s$ refers to the average change in summer AT: ΔAT$_s$ (°C) = 1.49°C

The changes in shallow (1.5 or 2.75 m) summer groundwater temperature projected for the MIROC 3.2 HIRES-A1B data are approximately 3°C (Figure 4.12). Changes to the temperature of groundwater discharge on this order would negatively impact riverine ecosystems by increasing the temperature of groundwater-sourced thermal refugia and increasing local surface water temperature in groundwater-dominated streams and rivers by increasing the heat flux at the river bed (see Caissie et al. 2007, Hebert et al. 2011). Although groundwater temperature will respond to decadal climate change, groundwater-dominated tributaries will generally continue to remain colder than the main stems and thereby induce riverine thermal heterogeneity. Catchments for tributaries that provide thermal refugia should be protected from deforestation (Alexander 2006, Bourque and Pomeroy 2001) and aggregate extraction (Markle and Schincariol 2007), which have been shown to increase shallow groundwater and river temperatures.
4.5.4 Limitations of the approach

Our modelling approach for estimating future groundwater temperature from GST assumes that the empirical parameters \( (B, D, \text{ and } L) \) will be unaffected by a warming climate. It seems reasonable to assume that this will be the case for \( D \) and \( L \) based on the physical processes they represent. The subsurface damping and lagging effects are primarily controlled by the period of the seasonal GST cycle, the soil diffusivity, and depth (Bonan 2008), and none of those controls will likely be significantly affected by atmospheric or surficial climate change in this catchment. Climate change-induced variations in the timing or magnitude of precipitation may impact soil moisture and consequently thermal diffusivity; however, these effects would be more noticeable in a catchment with a deeper water table. It is more likely that the \( B \) parameter would be affected by a warming climate. For example, increased precipitation and recharge rates in this catchment (Kurylyk and MacQuarrie 2013b) could potentially lead to accelerated advective heat transport due to increased groundwater flow. Increases in extreme precipitation events could potentially alter the \( B \) parameter due to changes in hydrological processes such as intensified subsurface lateral flow (horizontal advection) and heterogeneous surface ponding. Furthermore, at the latitude considered for this study, latent heat effects arising from pore-water phase changes may actually increase in a warmer climate, as winter GST are projected to decrease. As previously discussed, the importance of \( B \) decreases with depth, thus the results for deeper depths (e.g., 8.75 m) may be more reliable. Also, this empirical function assumes that the subsurface is in dynamic thermal equilibrium with the surface; however, it could potentially take years for the rise in mean annual GST to be fully realised in deeper (> 10 m) aquifer systems.
These complex subsurface climate change feedbacks could be simulated with physically based groundwater flow and heat transport models that can accommodate freezing and thawing effects. We anticipate that this research will motivate others to perform additional studies in other aquifer-river systems.

4.6 Conclusions

To investigate the relationships between climate change and GST in the Catamaran Brook catchment, simulations were conducted using the physically based surface flux balance model ForHyM2. In general, the ForHyM2 results indicate that the changes in summer GST will mimic changes in summer AT for this catchment. However, the model results also indicate that, due to complex snow-cover evolution effects, winter GST will likely decrease (−1.12 to +0.08°C). This will result in mean annual GST changes (−0.15 to +2.64°C) that are significantly damped with respect to mean annual AT changes (+0.39 to +3.96°C).

Measured groundwater temperature and GST were utilized to develop an empirical relationship between the surface and subsurface thermal regimes for this catchment. This empirical function was then driven with ForHyM2-produced GST projections to estimate the impact of atmospheric climate change on groundwater temperature. Results from these simulations indicated that shallow (1.5 m) groundwater temperature in this catchment is very sensitive to increases in summer AT. High-frequency GST signals, such as seasonal variations in GST, are damped at greater depths; however, low frequency GST signals, such as decadal climate change, are retained at these depths. Thus, even groundwater temperature at deeper depths will respond to increasing AT and GST. The results presented in this contribution are site-specific;
however, our approach could be employed to investigate surface and subsurface thermal processes in other catchments with seasonal snow cover.

This study has also demonstrated the limitations inherent in predicting future climate change impacts using a single projected climate series based on one emission scenario, simulated with one GCM, and downscaled using only one approach. Climate modelling involves many assumptions and, as such, an array of climate scenarios should be considered. We have also demonstrated that baseflow-dominated streams may exhibit more sensitivity to climate change than previous contributions have indicated. Salmonids are threatened by rising river temperatures in eastern North America, and the ability of groundwater to buffer rising river temperatures may be overestimated. Future physically based groundwater flow and heat transport modelling will be conducted to further investigate the potential increase in the temperature of groundwater discharge and the subsequent impact to cold-water fish habitat.

References


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Huard, D. 2011. Climate change scenarios: Thermal refuge for salmonids. Report # 011-02, Ouranos Consortium on Regional Climatology and Adaption to Climate Change, Quebec, Canada.


CHAPTER 5: Climate change impacts on the temperature and magnitude of groundwater discharge from shallow, unconfined aquifers

Abstract

Cold groundwater discharge to streams and rivers can provide critical thermal refuge for threatened salmonids and other aquatic species during warm summer periods. Climate change may influence groundwater temperature and flow rates, which may in turn impact riverine ecosystems. This study evaluates the potential impact of climate change on the timing, magnitude, and temperature of groundwater discharge from small, unconfined aquifers that undergo seasonal freezing and thawing.

Seven downscaled climate scenarios for 2046-2065 were utilized to drive surficial water and energy balance models (HELP3 and ForHyM2) to obtain future projections for daily ground surface temperature and groundwater recharge. These future surface conditions were then applied as boundary conditions to drive subsurface simulations of variably-saturated groundwater flow and energy transport. The subsurface simulations were performed with the U.S. Geological Survey finite element model SUTRA that was recently modified to include the dynamic freeze-thaw process. The SUTRA simulations indicate a potential rise in the magnitude (up to 34\%) and temperature (up to 3.6°C) of groundwater discharge to the adjacent river during the summer months due to projected increases in air temperature and precipitation. The thermal response of groundwater to climate change is shown to be strongly dependent on the aquifer dimensions. Thus, the

\(^1\)Kurylyk, B.L., MacQuarrie, K.T.B., and Voss, C.I. 2014. Climate change impacts on the temperature and magnitude of groundwater discharge from shallow, unconfined aquifers. Water Resources Research, 50(4): 3253-3274.
simulations demonstrate that the thermal sensitivity of aquifers and baseflow-dominated streams to decadal climate change may be more complex than previously thought. Furthermore, the results indicate that the probability of exceeding critical temperature thresholds within groundwater-sourced thermal refugia may significantly increase under the most extreme climate scenarios.

**Keywords**: river temperature, aquifer thermal regimes, thermal sensitivity, groundwater dependent ecosystems, fish habitat, thermal refugia

### 5.1 Introduction

Spatially discrete groundwater discharge can induce riverine thermal heterogeneity and thereby create microhabitats for cold water fishes during high temperature events (e.g., Ebersole et al. 2003, Goniea et al. 2006, Torgersen et al. 1999, Torgersen et al. 2012). These cold-water plumes, known as thermal refugia, enable fish to survive in river reaches that would otherwise be thermally uninhabitable (Sutton et al. 2007). Thermal refugia generated by groundwater discharge from adjacent/underlying aquifers have been observed at geographically diverse locations and for various aquatic species (e.g., Bilby 1984, Befus et al. 2013, Biro 1998, Briggs et al. 2013, Bunt et al. 2013, Dugdale et al. 2013, Olsen and Young 2009, Nielsen et al. 1994). Furthermore, the thermal regimes of shallow, unconfined aquifers are of ecological importance because diffuse groundwater discharge can buffer seasonal or diel variability in ambient river water temperatures (Caissie 2006, Hayashi and Rosenberry 2002, Story et al. 2003).
Buffering occurs because groundwater temperature exhibits less seasonal variability than surface water temperature due to the heat capacity of the overlying soil and the insulation from solar radiation (Bonan 2008).

Global climate models (GCMs) project significant shifts in global and regional air temperature (AT) and precipitation regimes (Meehl et al. 2007). These changing meteorological conditions will likely result in rising surface water temperatures and a corresponding loss of habitat for cold-water fish species (Chu et al. 2005, Jones et al. 2014, Jonsson and Jonsson 2009, Moore et al. 2013, van Vliet et al. 2011, 2013, Webb et al. 2008, Wu et al. 2012). The increased thermal stress on lotic ecosystems is expected to magnify salmonid reliance on groundwater-sourced thermal refugia (Brewer 2013), particularly at the lower elevational or latitudinal limits of their distributions. Several researchers have acknowledged the need for further studies investigating the influence of climate change on groundwater temperature and/or groundwater discharge rates (e.g., Chu et al. 2008, Mayer 2012, Mohseni et al. 2003). Kanno et al. (2013) noted that the ‘spatial variability in resiliency of groundwater temperature in response to air temperature is the critical missing piece to assess climate change impacts on headwater stream fish accurately.’

Research examining relationships between climate change and groundwater temperature and flow rates is limited, and knowledge gaps remain. Firstly, previous studies investigating the effects of climate change on groundwater were primarily focused on groundwater resources, largely ignoring subsurface thermal response to climate change (see reviews by Green et al. 2011, Kurylyk and MacQuarrie 2013, Taylor et al. 2013). Secondly, most studies that have considered the thermal evolution of aquifers due
to future climate change employed simplified one-dimensional analytical solutions to conduction-advection heat transport equations (e.g., Gunawardhana et al. 2011, Kurylyk and MacQuarrie 2014). One-dimensional solutions have limited ability to simulate groundwater flow and heat transport in shallow aquifers because both thermal and hydraulic processes are typically multi-dimensional. Also, existing solutions to the analytical solutions to the conduction-advection equation that account for climate change ignore seasonal changes in surface and subsurface temperature. Other analytical solution studies have assumed that conduction is the only significant heat transport process controlling the thermal response of groundwater to climate change (e.g., Taylor and Stefan 2009); however, heat advection due to groundwater flow can significantly perturb the conductive thermal regime in some subsurface environments (Woodbury and Smith 1985). Thirdly, the few climate change-groundwater temperature studies that have utilized multi-dimensional models were focused on relationships between permafrost degradation and aquifer reactivation (e.g., Bense et al. 2009, Frampton et al. 2013, McKenzie and Voss 2013, Painter 2011). These were conducted for high latitude/altitude climates, and did not detail effects of aquifer warming on riverine ecosystems. Finally, previous studies typically employed simplified climate change scenarios (e.g., a linear increase in surface temperature) rather than downscaled GCM output. Simplified scenarios may overlook complex relationships between atmospheric climate change and groundwater temperature, such as the subsurface thermal influence of changing groundwater recharge patterns and snowpack evolution.
The objective of this paper is to investigate the influence of climate change on the characteristics of groundwater discharge from shallow, unconfined aquifers to streams or rivers. In particular, we consider changes to the timing, magnitude, and temperature of groundwater discharge. The investigations are conducted in the context of the climate, hydrology, and hydrogeology of central New Brunswick, Canada. The hydraulic and thermal sensitivities of unconfined aquifers to climate change are simulated by driving surface energy and water balance models with downscaled climate model output and then applying the output from these surface models as boundary conditions for a groundwater flow and heat transport model. Unlike most previous studies, the groundwater discharge and temperature projections are investigated on both a mean annual and seasonal basis. The ecological implications are evaluated by considering simulated changes to the timing, magnitude, and temperature of groundwater discharge in the context of existing thermal refugia conditions and previously-established salmonid thermal tolerances. Particular emphasis is placed on the summer period when the existence of thermal refugia can be critical for riverine ecosystem complexity.

To our knowledge, this is the first study to utilize daily downscaled climate projections to investigate climate change impacts to aquifer thermal and hydraulic regimes. The study of future climate change impacts is becoming increasingly multidisciplinary (Meehl et al. 2007a), and it is important that groundwater hydrologists also progress towards the integration of GCM simulations with groundwater flow and energy transport modeling. Thus, we improve on the conventional practice of driving subsurface energy transport models with climate trends that are only loosely derived from GCM output and align our methodology with river temperature analysts who utilize
actual downscaled GCM output. This contribution, which employs a process-oriented groundwater flow and energy transport model, expands on a recent study that utilized an empirical groundwater temperature model to investigate shallow groundwater temperature response to climate change (Kurylyk et al. 2013).

5.2 Site Description and Conceptual Models

The present study is focused on groundwater-sourced thermal refugia within the Miramichi River system in New Brunswick, Canada, which is the largest producer of wild Atlantic salmon (*Salmo salar*) in North America (Caissie et al. 2007). In New Brunswick, summer river temperatures are currently approaching the critical threshold for salmonids (Breau et al. 2007, Cunjak et al. 2013), and discrete cold water plumes formed by groundwater-surface water interactions have been shown to provide critical thermal refuge for salmonids in the Miramichi River system (Breau et al. 2007, 2011, Cunjak et al. 2005, Monk et al. 2013). The Little Southwest Miramichi River (LSW) is a sixth order branch of the Miramichi River located in central New Brunswick, Canada (Figure 5.1) that experiences a humid-continental climate characterized by arid, cold winters (Cunjak et al. 1993). Annual precipitation is 1230 mm, with approximately 33% falling as snow (EC 2013). The LSW is a relatively wide, shallow river with a width to depth ratio of approximately 150. Thus it responds rapidly to summer radiation, with water temperatures occasionally exceeding 30°C.

An airborne infrared thermal survey of the LSW (Wilbur and Curry 2011) identified numerous potential thermal anomalies, including one on the south side of the LSW at the mouth of a cold, baseflow-dominated tributary (Otter Brook, N46 52 W66 02)
and one on the north side of the LSW at a lateral groundwater seep (Figure 5.1). A thermal infrared image of the LSW and the thermal refugia at the groundwater seep and the mouth of Otter Brook is included in Chapter 1 (Figure 1.5). These identified refugia are biologically important. For example, the refuge at the mouth of Otter Brook (Figure 5.1) accommodated approximately 6000-10,000 juvenile Atlantic salmon simultaneously in July 2010 when LSW temperatures exceeded 30°C (Linnansaari, UNB, pers. communication, 2013). These field observations of unconfined, alluvial aquifers generating thermal refugia form the motivation and conceptual models for this study.

![Figure 5.1. Study site. The locations of the Little Southwest Miramichi River, Otter Brook and the groundwater seep (b, c, and d) are shown with respect to their location in New Brunswick, Canada (a) (map data from, NBADW 2011). The coordinates for Otter Brook are N46 52 W66 02.](image)
The unconfined aquifers that provide groundwater discharge to Otter Brook and the lateral groundwater seep were conceptualized as being two-dimensional (cross-sections) and homogeneous. The thermal and hydraulic properties of the aquifers were derived from field observations of surface and subsurface conditions in these catchments. The Otter Brook and groundwater seep deposits are primarily composed of highly permeable glaciofluvial outwash sediments, varying from cross-bedded sand to thick-bedded coarse gravel (Allard 2008). Aquifer hydraulic properties were established from grain size analyses conducted by Allard (2008) in conjunction with the methodology proposed by Hazen (1911). More details are provided in Appendix 3.

Dimensions of the aquifers contributing to the groundwater seep and Otter Brook were estimated from a digital elevation model (DEM) (NBADW 2011) and/or GPR surveys (Allard, 2008). Although Otter Brook diverges into two branches approximately 650 m upstream of its mouth (Figure 5.1b), the brook was conceptualized as a single symmetrical branch with an outwash deposit on either side. Section A-A’(Figure 5.1b) was chosen as the simulation cross section for the Otter Brook deposit because surface water thermal surveys indicated significant groundwater discharge in this region. Because the Otter Brook catchment is a marshy environment (Figure 5.1d), the depth of groundwater discharge to Otter Brook is near the ground surface. The aquifer discharging to the groundwater seep is underlain by a grey clay aquitard (Figure 5.1c) that forces the groundwater seep to discharge at an elevation above the LSW water surface. Due to the steep vertical bank on the north side of the LSW, the depth of the seep discharge is on average 7 m below the ground surface. These two distinct aquifer configurations were selected to investigate the influence of an aquifer’s dimensions on its hydraulic and...
thermal response to atmospheric climate change. The idealized aquifers representing the hydrogeological units discharging to Otter Brook and the groundwater seep are hereafter referred to as ‘Configuration 1’ and ‘Configuration 2’, respectively. Our intent is not to make predictions regarding the future states of these two particular aquifers, but rather to assess the sensitivity of thermal refugia-generating aquifers to external climatic forcing.

5.3 Numerical Simulation Approach

The physical processes that generate groundwater-sourced thermal refugia and the associated overall modeling sequence employed in the present study are depicted in Figure 5.2. Atmospheric processes drive hydraulic and thermal exchanges across the ground surface, and these surface processes in turn control subsurface conditions. The combined effects of these physical processes can be simulated in a sequential manner by linking physically-based models for the lower atmosphere, ground surface, and shallow subsurface (Figure 5.2). In the present study, surface simulations were performed independently of subsurface simulations. The modeling processes shown in steps 2a, 2b, and 2c of Figure 5.2 are described in subsequent sections.

5.3.1 Climate data

Daily observed climate data (AT and precipitation) for 1961-2000 were obtained for the LSW region from the Environment Canada (EC) adjusted and homogenized daily climate database (EC 2011). These observed climate data were utilized to drive reference period simulations of surface processes to form a datum from which to evaluate the sensitivity to future climate change (Figure 2a and 2b).
Figure 5.2. The physical processes generating groundwater-sourced thermal refugia (left) and the associated modeling sequence (right). Climate data are utilized to drive surface simulations of the energy and water budget, and the results from the surface simulations form the boundary conditions for the variably-saturated groundwater flow and heat transport model.

Coarse-resolution GCM projections are often statistically downscaled (Chen et al. 2012, Jeong et al. 2012b) or dynamically downscaled with regional climate models (RCMs, Boe et al. 2007, de Elia et al. 2008) and further debiased (Teutschbein and Seibert 2012) to produce projections of local climate data and hydrology. For the present study, seven projected climate scenarios (Table 5.1) were produced for the period of 2046-2065 using six GCMs, two downscaling methods, and three emission scenarios. Two downscaled climate scenarios were obtained via hybrid multiple regression statistical downscaling approaches (HMR, Jeong et al. 2012a, Jeong et al. 2012b). These climate series were contributed by the Université du Québec Institut National de la Recherche Scientifique (INRS) (D. Jeong, personal communication, 2011). The other
climate data series were produced from the third Coupled Model Inter-comparison Project database of GCM output (CMIP3, Meehl et al. 2007b) and statistically downscaled with the daily translation (DT) method or dynamically downscaled using the Canadian Regional Climate Model (CRCM4.2.3 de Elia et al., 2008) and further processed (Huard, 2011). The ‘ID’ in Table 5.1 refers to a particular RCM simulation and GCM driver. These climate series are further described in Kurylyk and MacQuarrie (2013).

The climate series span the range of plausible projected changes in local mean annual AT (range among climate scenarios of +0.4 to +3.9°C, mean +2.1°C) and precipitation (range of -12% to +49%, mean +8%) for 2046-2065 compared to the 1961-2000 observed climate data (Figure 5.3a).

### Table 5.1. Details for the climate scenarios utilized in this study

<table>
<thead>
<tr>
<th>Emission Scenario</th>
<th>Model Type</th>
<th>Model Name</th>
<th>GCM Driver</th>
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<td>-</td>
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#### 5.3.2 Surface and near-surface modeling of recharge and ground surface temperature

The observed climate data from 1961-2000 (EC 2011) and the climate scenarios given in Table 5.1 were utilized to drive simulations within the hydrological model HELP3 (Hydrologic Evaluation of Landfill Performance, Schroeder et al. 1994) to obtain
future projections of daily groundwater recharge (Step 2a, Figure 5.2). HELP3 is a quasi-two-dimensional daily soil water balance model capable of simulating the influence of changing climatic conditions on surface and shallow subsurface hydrological processes, including snowpack accumulation and melt (e.g., Allen et al. 2010, Crosbie et al. 2011). In HELP3, precipitation accumulates as snowpack during the winter months when AT is low causing a seasonal cessation of recharge. HELP3 also simulates a decrease in late fall and early spring recharge by increasing the runoff curve number during periods when air temperatures are low and soil is presumed to be frozen. Additional information describing the HELP3 modeling process can be found in Kurylyk and MacQuarrie (2013). The HELP3 simulations indicate that projected changes to future mean annual groundwater recharge for each climate series vary significantly (range of -6% to +58%, mean +18% for 2046-2065 compared to the recharge simulation using observed climate data, Figure 5.3b). These simulations also indicate that spring snowmelt and the consequent major recharge event will likely occur earlier in the year due to increases in winter and early spring AT (Kurylyk and MacQuarrie 2013).

The observed and projected climate series (Table 5.1) were also utilized to drive simulations within the surface energy flux balance model ForHyM2 (Forest Hydrology Model, Yin and Arp 1993) to obtain future projections of daily ground surface temperature (GST, Step 2b, Figure 5.2). ForHyM2 is a water and energy transport model that simulates surficial energy fluxes (e.g., shortwave and longwave radiation and convective exchanges) and near-surface heat transport via conduction through multiple layers (forest canopy, snowpack, forest floor, soil, and subsoil) (Balland et al. 2006, Yin and Arp 1993). ForHyM2 was only applied to obtain GST for the present study, as the
model is one-dimensional and does not accommodate subsurface advective heat transport. ForHyM2-simulated daily GST agree closely (mean error = 0.02°C) with field observations of GST during both snow-covered and snow-free periods (Figure 4 of Kurylyk et al. 2013). ForHyM2 simulations for the projected climate scenarios suggest that changes to average summer GST (mean among climate scenarios of +1.49°C compared to the 1961-2000 simulation) will generally follow average summer AT changes. However, projected increases in winter AT (mean +3.04°C) will result in decreased winter GST (mean -0.53°C) due to a reduction in the winter snowpack and its insulating capabilities. As Figure 5.3b illustrates, the decrease in simulated winter GST results in simulated average annual GST changes (mean +1.06°C) that are damped compared to the projected increases in average annual AT (mean +2.1°C). Further details related to the ForHyM2 modeling process can be found in Kurylyk et al. (2013).

Note that the details given by Kurylyk et al. (2013) relate to simulations performed for the Catamaran Brook catchment, which is a small catchment adjacent to Otter Brook. Because of the geographical proximity and similar surface and shallow subsurface conditions, the GST simulations performed for the Catamaran Brook catchment were assumed to also represent current and projected GST trends in the Otter Brook catchment and the lateral groundwater seep catchment (Figure 5.1). Similarly, the current and projected recharge simulated for the Otter Brook catchment (Kurylyk and MacQuarrie 2013) are also assumed to be representative of the current and future recharge regimes in the groundwater seep catchment due to the similar surface and shallow subsurface conditions (Allard 2008).
Figure 5.3. (a) The observed/reference (1961-2000) and projected (2046-2065) mean annual AT and precipitation (see Table 5.1) and (b) the mean annual groundwater recharge and GST simulated by driving surface models HELP3 and ForhyM2 with the climate data shown in part (a).

5.3.3 Groundwater flow and heat transport model selection

Due to the noted limitations inherent in analytical solutions, a numerical model was employed to simulate the impacts of atmospheric and surficial climate change on the timing, magnitude, and temperature of groundwater discharge from shallow, unconfined aquifers. Pore ice formation can impact groundwater discharge conditions where significant seasonal freezing occurs. For example, the latent heat released or absorbed during pore water freeze-thaw usually dominates conductive and advective heat transport in the zone of freezing and significantly retards the propagation of surface temperature signals during the early winter freeze and spring thaw (Woo 2012). Furthermore, pore ice formation reduces hydraulic conductivity and impedes groundwater flow (Kurylyk and Watanabe 2013, Watanabe and Wake 2008). The model selected for the present study, SUTRA, is a multi-dimensional finite element model of coupled groundwater flow and heat transport (Voss and Provost 2002) that has been modified by McKenzie et al. (2007)
to accommodate freezing and thawing processes for saturated conditions, and then recently enhanced to allow for variably-saturated conditions during freezing and thawing. In addition to accommodating the dynamic freeze-thaw process, SUTRA simulates subsurface heat transport via conduction, advection, and thermal dispersion, allowing pore water flow due to gradients in pressure, elevation, and water density (Voss and Provost 2002). The governing water flow and heat transport equations, model parameterization (thermal and hydraulic properties, relative permeability function, soil drying curve, and soil freezing curve) and model controls (mesh density, time step size, and solver settings) are given in Appendix 3.

5.3.4 SUTRA boundary conditions and run information

Figure 5.4 depicts the boundary conditions and domains for the SUTRA simulations. Both aquifers were assigned no-flow hydraulic boundaries and perfectly-insulating thermal boundaries at the groundwater divide (left vertical boundaries, Figure 5.4). The bottom boundary was specified as a no-flow boundary condition for both configurations. A specified heat flux (0.060 W m\(^{-2}\)) was also assigned to this boundary to represent the geothermal heat flux from greater depths (e.g., Bense et al. 2009). The discharge location for both configurations was represented as a specified pressure boundary condition. For Configuration 1, this pressure increased linearly with depth (1 m) to represent hydrostatic pressure within Otter Brook.
Figure 5.4. SUTRA aquifer configurations and boundary conditions for (a) Configuration 1 and (b) Configuration 2. Vertical exaggeration is approximately 25:1. The finite element meshes employed in SUTRA were much denser than those shown above (see Appendix 3).

The surface boundaries for both aquifers were assigned specified temperature (GST) and specified fluid flux (groundwater recharge) boundary conditions, with recharging water having the same temperature as the ground surface. By directly specifying the GST output from ForHyM2, it was possible to simulate subsurface responses to complex atmosphere-surface interactions (e.g., snowpack accumulation, insulation, and ablation). A specified groundwater recharge boundary condition is preferred to a specified-pressure boundary condition if, as in this case, the recharge is climate-controlled (Sanford 2002). This approach allows for a natural groundwater table and unsaturated zone to develop. Recharge, rather than infiltration, was specified along...
the ground surface boundary because this version of SUTRA (4.0 beta) does not include evapotranspiration.

SUTRA simulations were performed with fine temporal and spatial scales (constant time step = 4.8 hours (0.2 days), minimum element height = 0.03 m). The model predicts the volumetric liquid and ice saturations, temperature, and pore water pressure at every time step for each node in the model domain (85,170 and 62,300 nodes for Configurations 1 and 2, respectively) and groundwater velocity for every element. Table 5.2 lists the details of each simulation performed.

### Table 5.2. Details for each simulation performed in SUTRA

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1 The number following the decimal point in the Run ID refers to the aquifer configuration. For example, Run 1.1 is Run 1 (i.e., the CGCM3-A2 climate scenario with advection and freezing on) for Config. 1.
2 ‘Freezing’ refers to the subroutine in SUTRA to accommodate pore water freezing
3 ‘Init. Cond.’ refers to initial conditions, which were generated by first running simulations for 40 years with observed climate data (1961-2000) and then using those conditions to start a subsequent run driven by 1961-1980 climate data. The 1961-1980 simulation results formed the initial conditions for each of the other runs indicated in Table 5.2.
4 The period for each of the future climate scenarios was 2046-2065 (see Table 5.1).
Reference period simulations (runs R.1 and R.2) were performed to establish a baseline from which to measure projected future changes in the timing, magnitude, and temperature of groundwater discharge (runs 1.1 to 7.2). Runs 8.1-9.2 were performed to investigate the thermal influence of advection and pore water phase change and thereby determine the suitability of employing simplified models (i.e., conduction-only) to predict the response of shallow groundwater to climate change.

5.4 Results

5.4.1 Seasonal and long term changes to groundwater discharge temperature

Figure 5.5 shows measured AT and simulated GST and groundwater discharge temperature for both aquifer configurations averaged for each calendar day over the 20 year reference period (1981-2000, SUTRA runs R.1 and R.2 in Table 5.2). The presence of the insulating snowpack clearly decouples winter AT and GST trends. The groundwater discharge temperatures presented in Figure 5.5 and subsequent figures were obtained from the bottom node of the discharge boundary (specified pressure boundary, Figure 5.4). Temperatures at the discharge boundary were shown to be relatively uniform due to high groundwater flows (advection) and concomitant high dispersion.

The groundwater discharge temperatures for Configuration 2 are characterized by more damping and lagging than the groundwater discharge temperatures for Configuration 1 due to the increased depth of the discharge point (Figure 5.4). A groundwater discharge thermal damping factor can be defined as the ratio of the amplitude of the annual groundwater discharge temperature cycle to the amplitude of the
annual GST cycle. For the reference period simulation, the average amplitudes of the annual groundwater discharge temperature cycles are 8.3°C and 2.1°C for Configurations 1 and 2, respectively, whereas the amplitude of the annual GST cycle is 11.6°C (Figure 5.5). Thus, the attenuation of the seasonal GST signal for Configuration 2 (damping factor = 0.18) is approximately four times that of Configuration 1 (damping factor = 0.72). Figures 5.5 and 5.6 also indicate that the maximum groundwater discharge temperature for the reference period simulations occurs on days of the year (DOY) 223 and 303 for Configurations 1 and 2, respectively. This indicates that the groundwater discharge temperature signal is lagged an additional 80 days for Configuration 2.

Figure 5.5. Measured AT, ForHyM2-simulated GST, and SUTRA-simulated groundwater discharge temperature for both aquifer configurations averaged for each day of the year during the reference period (1981-2000, Runs R.1 and R.2, Table 5.2). Although not depicted in these results, the AT and GST exhibit considerably diel variability, whereas the groundwater discharge temperature, even in Configuration 1, is constant throughout the day.
Figure 5.6 depicts the simulated groundwater discharge temperatures for the reference period (SUTRA runs R.1 and R.2, Table 5.2) and the future period (SUTRA runs 1.1-7.2). The future period simulations shown in Figure 5.6 are generally characterized by higher groundwater discharge temperatures that are shifted slightly earlier in the year. For example, the warmest climate scenario (MIROC-HIRES-A1B) results in a maximum groundwater discharge temperature for Configuration 1 that is 3.3°C higher than that simulated for the reference period simulation.

Figure 5.6. Simulated groundwater discharge temperature for (a) Configuration 1 and (b) Configuration 2. Results are shown for the reference period (runs R.1 and R.2, Table 5.2) and for the last ten years (2056-2065) of data for each of the climate scenarios given in Table 5.1 (runs 1.1-7.2, Table 5.2). The results are averaged over the simulation period for each day of the year. The plots are drawn to different scales.
The climate change-induced increases in groundwater discharge temperature simulated for each calendar day for Configuration 2 are relatively constant and approximate the increase in mean annual GST for each climate scenario (see Figure A4.3, Appendix 4). However, the simulated changes in daily groundwater discharge temperatures for Configuration 1 are characterized by considerable daily variability.

5.4.2 Influence of advection and pore water phase change

Heat advection via groundwater flow can significantly perturb subsurface thermal environments (Kurylyk and MacQuarrie 2014, McKenzie and Voss 2013). Also, the latent heat released or absorbed during freezing or thawing greatly increases the subsurface effective heat capacity and thereby attenuates high frequency GST signals. In addition to releasing or absorbing latent heat and impeding heat advection, the dynamic freeze-thaw process can also influence subsurface thermal regimes by altering the bulk thermal diffusivity (thermal conductivity/heat capacity) of the porous medium, as the thermal diffusivity of ice is approximately eight times that of water (Bonan 2008). To test the subsurface thermal influence of groundwater flow and pore water phase change, separate reference period simulations were performed for both configurations without considering the effects of advection (SUTRA runs 8.1 and 8.2, Table 5.2) or the dynamic freeze-thaw process (SUTRA runs 9.1 and 9.2).

Figure 5.7 demonstrates the thermal influence of groundwater flow in Configuration 1 for the reference period. The temperatures at the specified pressure boundary condition (i.e., the location of groundwater discharge when flow is activated)
were compared for the simulations including (run R.1) and neglecting (run 8.1) the thermal influence of advection. Because groundwater flow primarily occurs during the warmer months and abates during the winter months, the outlet temperatures for the simulation neglecting advection are generally colder (maximum difference = 3.1°C) than the groundwater discharge temperatures for the simulation with advection considered.

Figure 5.7. Simulated groundwater discharge temperature vs. the day of the year for Configuration 1. Results are presented for the default simulation for the reference period (run R.1, Table 5.2) and a simulation with no recharge (and thus no advection) applied (run 8.1). Results are averaged for each day of the year over the reference period.
Figure 5.8 shows the results for the reference period Configuration 1 simulations performed with (run R.1) and without (run 9.1) the influence of pore water phase change considered. The results demonstrate that the inclusion of the freeze-thaw process can significantly impact the groundwater discharge temperature in a given year, but that its influence is reduced in this aquifer when the results are averaged over the 20 year simulation. For example, the maximum difference between the two series representing the minimum groundwater discharge temperature for each calendar day is 2.71°C, but this difference decreases to 0.81°C when the results are averaged for each calendar day of the simulation.

Temperature profiles were also analysed halfway up the hillslope in Configuration 1 (i.e., 500 m from the outlet, Figure 5.4a) to investigate the impact of the dynamic freeze-thaw process at locations other than the discharge location. The results, which are shown in Figure A4.3 (Appendix 4), indicate that the maximum difference between the simulated groundwater temperatures at 1 m depth with and without the effects of freeze-thaw considered (i.e., runs R.1 and 9.1) was 4.03°C. However, the thermal influence of freeze-thaw is quickly attenuated with depth and is barely discernible at depths greater than 4 m in this aquifer. In general, the results presented in Figures 5.7 and 5.8 and the auxiliary material indicate that properly accounting for the thermal influence of advection and pore water phase change is most important in the shallow subsurface of the LSW catchment.
Figure 5.8. Configuration 1 groundwater discharge temperature vs. the day of the year for the reference period simulation (1981-2000). Solid lines indicate the results for when the effects of ice were considered (run R.1, Table 5.2), and the dashed lines indicate the simulation results when the effects of ice were not included (run 9.1). Results are presented for the discharge temperature averaged for each day of the year over the 20 year simulation and for the minimum discharge temperature simulated for each day of the year.

5.4.3 Simulated groundwater discharge magnitude and timing

Figure 5.9 presents simulated reference period daily groundwater recharge and discharge for both aquifer configurations. In general, the groundwater discharge per unit width from Configuration 1 exceeds the groundwater discharge per unit width from Configuration 2 due to their differing aquifer lengths (1000 and 600 m, Figure 5.4). Figure 5.9 indicates that the magnitude and timing of groundwater discharge from these highly permeable aquifers is strongly controlled by intra-annual groundwater recharge. For example, the timing of the maximum groundwater discharge rate for the reference period simulation lags the timing of the maximum recharge rate by only 13 and 14 days for Configurations 1 and 2, respectively. The average annual total recharge masses per
metre aquifer width (534,000 kg and 320,000 kg for Configurations 1 and 2 per meter aquifer width) are approximately equal to the average annual total discharge masses (538,000 kg and 322,000 kg for Configurations 1 and 2 per meter aquifer width). There are, however, discernible differences between the seasonal recharge and discharge rates shown in Figure 5.9. Due to the subsurface hydraulic storage properties, the temporal variability of groundwater discharge is damped in comparison to the temporal variability of groundwater recharge. For example, groundwater recharge exceeds discharge during the spring months, and this accumulated water is slowly released from the aquifers during the summer months when discharge typically exceeds recharge (Figure 5.9). Similarly, late fall recharge exceeds discharge, and this water is released during the winter when recharge ceases during snowpack accumulation (Kurylyk and MacQuarrie 2013).

![Graph](image)

**Figure 5.9.** Simulated groundwater recharge and discharge per meter aquifer width for Configuration 1 (SUTRA run R.1, Table 5.2) and Configuration 2 (run R.2) averaged for each day of the year for the reference period.
Figure 5.10 shows the simulated groundwater rates for both aquifer configurations for the reference period and future climate series. In general, the future period simulations exhibit increases or decreases in average groundwater discharge rates that correspond to the simulated increases or decreases in average annual recharge (Figure 5.3b). Some similarities exist between reference period and future period simulations. For example, Figure 5.10 shows that a major discharge event occurs in the spring as a result of snowmelt. These peak daily discharges range from 3855 kg (MIROC-HIRES-A1B) to 6846 kg (CGCM3-A1B) per meter width for Configuration 1, and 2843 kg (CRCM-AEV-A2) to 4668 kg (CGCM3-A1B) for Configuration 2. Each discharge series in Figure 5.10 exhibits a distinct decreasing trend in discharge during the summer months (June-August) arising from the abatement of summer recharge due to reduced precipitation and increased evapotranspiration (Kurylyk and MacQuarrie, 2013). Each simulation also includes a smaller groundwater discharge event corresponding to the late-fall rainy season and a reduction in groundwater discharge during the winter months due to recharge cessation.
5.5 Discussion

5.5.1 Changes to the mean and amplitude of the groundwater discharge temperature cycle

The simulated mean annual groundwater discharge temperature trends for each climate scenario generally follow the trends in projected mean annual GST (Figures 5.3b and 5.6). For example, the MIROC3.2-HIRES-A1B climate series simulations (run 6.1) are characterized by the largest increases in mean annual AT (+3.96°C, Figure 5.3a) and
mean annual GST (+2.65°C, Figure 5.3b) compared to the reference period, and this climate scenario is also characterized by the most pronounced increases in mean annual groundwater discharge temperature (+2.85°C and +2.70°C for Configurations 1 and 2, respectively). The SUTRA simulation results for the CSIRO Mk3-B1 scenario exhibit slight decreases in both the mean annual groundwater discharge temperature (-0.45°C and -0.33°C for Configurations 1 and 2, respectively) and summer groundwater discharge temperature (-0.73°C and -0.53°C for Configurations 1 and 2, respectively), but all other climate scenarios result in an increase in mean annual and summer groundwater discharge temperature (Figure 5.6). It should be noted that actual trends in global CO₂ emissions and concentrations have exceeded emission scenario B1 projections (Raupach et al. 2007), thus the CSIRO Mk3-B1 climate scenario likely underestimates future changes in climate.

The simulations also suggest that the amplitude of the seasonal groundwater temperature cycle may increase in shallow aquifers. For example, the increase in average winter groundwater discharge temperature (Dec-Feb, +1.41°C) for the MIROC3.2-HIRES-A1B scenario (Configuration 1, Figure 5.6a) was not as pronounced as the simulated increase in average summer groundwater discharge temperature (Jun-Aug, +3.37°C). This amplitude increase arises from the projected changes in seasonal GST, which were typically positive in the summer and negative in the winter (Kurylyk et al. 2013). These results differ from those of Taylor and Stefan (2009), who found that the minimum and maximum groundwater temperature would increase by approximately the same amount.
5.5.2 Aquifer summer thermal sensitivities

Changes to the temperature of summer groundwater discharge are of particular interest to stream ecologists as this is the period most critical for the generation of thermal refugia. Figure 5.11 presents the simulated changes in mean annual and mean summer groundwater discharge temperature versus the changes in mean annual and mean summer AT. This information can be utilized to generally assess the thermal response of each aquifer to decadal climate change. The groundwater discharge temperature data in Figure 5.11 display a relatively consistent trend amongst climate scenarios (relatively high $R^2$ for trend lines) for each aquifer configuration; however, the slopes of the best fit lines indicate that the thermal regime of Configuration 1 is more sensitive to decadal summer AT increases than the thermal regime of Configuration 2. This difference arises because the summer thermal regime of Configuration 1 is driven by summer GST trends, which closely follow summer AT increases. Conversely, the summer thermal regime of Configuration 2 is primarily driven by changes in mean annual GST, which were damped compared to mean annual or seasonal AT changes due to snowpack thinning or removal (Kurylyk et al. 2013).

The decrease in mean annual GST changes compared to mean annual AT changes also accounts for the higher slope through the mean summer data compared to the mean annual data (Figure 5.11). Configuration 2 exhibits less sensitivity to seasonal GST variability than Configuration 1 (Figure 5.5), thus the difference between the slopes representing mean annual and mean summer data in Figure 5.11 is not as pronounced for Configuration 2 (0.60-0.51=0.09) as it is for Configuration 1 (0.95-0.57=0.38). Figure
5.11 also indicates that, contrary to the assumptions made in other approaches (e.g., Deitchman and Loheide 2012, MacDonald et al. 2014, Meisner et al. 1988), climate-induced changes to groundwater discharge temperature will not necessarily follow AT changes on a mean annual or seasonal basis.

Figure 5.11. Simulated changes in mean annual and mean summer (June 1- August 31) groundwater discharge temperature for the last ten years of each future period simulation (2056-2065, runs 1.1-7.2, Table 5.2) and the reference period simulation (1981-2000, runs R.1 and R.2) vs. the change in mean annual and mean summer AT for (a) Configuration 1 and (b) Configuration 2. The mean annual results are depicted by the symbols with fill. The mean summer results for each climate scenario are represented by symbols with the same shape and color but with no fill. The best fit lines were constrained to have a zero intercept.

Several recent studies have investigated the sensitivity of stream and river temperature to AT variations (e.g., Bogan et al. 2003, Kelleher et al. 2012, Mayer 2012). Kelleher et al. (2012) and Mayer (2012) specifically defined the thermal sensitivity of surface water as the slope of the surface water temperature-AT plot. These plots are typically derived from high-frequency (e.g., weekly or seasonal) AT and surface water temperature data, but can be employed to estimate the thermal response of rivers to
decadal climate change (e.g., Mayer 2012, Kelleher et al. 2012). The summer aquifer thermal sensitivity is herein defined as the simulated change in average summer groundwater discharge temperature for a particular climate scenario divided by the driving change in average summer AT. Note that this is a low-frequency (decadal climate change) sensitivity definition rather than the high frequency sensitivity definition employed by surface water temperature analysts.

Table 5.3 indicates that the mean aquifer thermal sensitivity for Configuration 1 (1.00) is significantly higher than the mean sensitivity for Configuration 2 (0.55). The aquifer thermal sensitivities in Table 5.3 exceed some previously reported values for the response of stream or river temperature to climate change. For example, Wu et al. (see Table 3, 2012) simulated an average 1.37°C summer stream temperature rise in response to an average 2.81°C summer AT rise (1.37/2.81=0.49) for Pacific Northwest streams. Others have simulated surface water sensitivities that exceed the summer thermal sensitivity of Configuration 2 but are less than that of Configuration 1. For instance, Morrill et al. (2005) modeled future stream temperatures in geographically diverse locations and demonstrated that the majority of streams will exhibit increases in stream temperature of 0.6 to 0.8°C for each 1°C increase in AT. Thus, the site specific results shown in Table 5.3 suggest that the thermal regimes of certain aquifers (and thus groundwater-sourced refugia) may, in some case, actually be more responsive to decadal climate change than the thermal regimes of streams and rivers. The high aquifer thermal sensitivities reported in Table 5.3 arise in part because aquifers do not experience free water surface evaporation, which cools the thermal regimes of surface water bodies.
during high temperature events and thereby reduces their thermal sensitivity (Mohseni and Stefan 1999).

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<th>Config. 1 STS</th>
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<td></td>
<td><strong>0.55</strong></td>
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1 The change in summer AT values are with respect to the reference period (1981-2000). These deviate slightly from those reported by Kurylyk et al. (2013, Table 2) because the present study only considers the last ten years of the climate scenario simulations (2056-2065).
2 ΔGWDT = the change in summer (June-August) groundwater discharge temperature compared to the reference period (1981-2000) summer groundwater discharge temperature.
3 STS= summer thermal sensitivity for each aquifer due to low frequency climate variations: = ΔGWDT/ΔAT.
4 Average STS values were calculated without including the anomalous CSIRO Mk 3.0-B1 and CSIRO Mk 3.5-B1 data, as the very low summer AT changes for these two scenarios yielded unreasonably high sensitivity values.

Groundwater discharge temperature does not typically respond to high frequency (e.g., weekly) AT variability, and thus baseflow-dominated streams tend to be characterized by lower sensitivity to weekly or daily climate variability compared to direct-flow dominated streams (Bogan et al. 2003, Hayashi and Rosenberry 2002, Risley et al. 2010, Tague et al. 2007). However, one should not infer from this that baseflow-dominated streams and rivers will also be less responsive to decadal climate change. We have demonstrated that aquifer thermal regimes can exhibit considerable sensitivity to low frequency AT variations (Figure 5.11).
The results portrayed in Figure 5.11 and Table 5.3 indicate that there is significant variability in how climate change may influence aquifer thermal regimes. Consequently the response of aquifers, groundwater-sourced thermal refugia, and baseflow-dominated streams to climate change may be more complex than previously thought. This merits further analysis, as the temperature of future groundwater discharge is only one of the factors controlling the future thermal regimes of baseflow-dominated streams.

5.5.3 Lag between climate change and groundwater temperature rise

Previous researchers (e.g., Gunawardhana et al. 2011, Kurylyk and MacQuarrie 2014) have demonstrated that there is a lag between a GST increase due to climate change and its thermal impact on deeper subsurface environments. However, the thermal regime of the shallow subsurface (e.g., depth <10 m) exhibits a very short lag in response to surficial climate change. For example, Figure 5.12a demonstrates that the simulated future groundwater discharge temperatures for Configurations 1 and 2 respond very quickly (< 5 yr) to a step change in GST due to atmospheric climate change. Furthermore, the lag would be even shorter if the boundary condition represented gradual climate change rather than the step change simulated in the present study. Configuration 2 takes approximately 2 years longer than Configuration 1 to achieve thermal equilibrium with the new GST due to its deeper discharge point (Figure 5.12a). Similarly short lags for the observed response of groundwater temperature to surficial climate changes were obtained from the statistical analysis conducted by Menberg et al. (2014).
The thermal response of groundwater to climate change is often assumed to be a consideration limited to the distant future. For example, Chu et al. (2008) state that ‘the potential changes in groundwater temperatures and fish habitat may take decades or centuries to be realized’. This proposition was used to justify a modeling approach that
did not explicitly consider changes to groundwater conditions. The short lags shown in Figure 5.12a strongly demonstrate that assessments of future surface water temperature should consider relatively imminent increases in the temperature of groundwater discharge from shallow aquifers, at least in the case where surface water thermal regimes are strongly influenced by groundwater discharge.

5.5.4 Subsurface thermal influence of advection and pore water phase change

The reference period simulations performed with and without groundwater flow (Figure 5.7) indicate that advective influences can be significant (max difference between series = 3.1°C). In general, models ignoring advection can under-predict summer groundwater discharge temperature and thus over-predict the ability of that discharge point to provide thermal refuge. This illustrates the limitations of employing a simple conduction-only model to investigate the response of groundwater discharge to climate change or seasonal temperature variation. A more detailed investigation of the internal aquifer energy balance and the relative roles of conduction and advection is included in Appendix 4 (Figure A4.1).

Figure 5.8 indicates that the influence of pore water phase change on groundwater discharge temperature can be significant in a given year, but it tends to be minimal in this catchment when averaged over the 20 year simulation period. The thermal influence of the dynamic freeze-thaw process is more apparent in shallow subsurface temperature profiles (e.g., 1 or 2 m depth) recorded halfway up the hillslope (Figure A4.3, Appendix 4). In general, not accounting for the latent energy released during freezing yields
shallow subsurface temperatures that are too low during the onset of freezing in the late fall-early winter. Conversely, not accounting for the latent heat absorbed during thawing will typically yield shallow subsurface temperatures that are too high during the late winter-early spring thaw. At high latitudes or altitudes, the frost penetration depth increases, and latent heat effects influence thermal regimes at greater depths (Zhang et al. 2008, Woo 2012). Latent energy effects due to phase change are also more apparent in soils with higher porosity and moisture retention properties, such as peat (McKenzie et al. 2007). Thus, the influence of pore water phase change on the temperature of groundwater discharge would be even more evident in colder climates or for different soils than those considered in the present study. The simulations presented in this study help elucidate the complex thermal dynamic of shallow aquifers and thereby inform researchers of the limitations arising from employing simplified models.

5.5.5 Impact of climate change on the magnitude and timing of mean annual and maximum groundwater discharge

The hydraulic conditions of shallow aquifers may also respond rapidly to atmospheric climate change. Figure 5.12b presents the annual total discharge and recharge rates for the reference period and the climate scenario exhibiting the greatest increase in recharge and discharge for the last ten years of the climate scenarios. Clearly, both aquifer configurations adjust quickly to the surface hydraulic perturbations. The step increase in groundwater recharge due to climate change results in an approximately corresponding step increase in groundwater discharge within one year. This rapid response in annual groundwater discharge due to changes in annual groundwater recharge
is expected given that these shallow unconfined aquifers also responded quickly to intra-annual recharge variability (Figure 5.9). Figure 5.12b also indicates that total groundwater discharge will not exactly equal total groundwater recharge in a given year due to aquifer storage capabilities.

Figure 5.10 presents the simulated changes in groundwater discharge for each climate scenario throughout the year. Despite similarities noted in the results, there are also discernible differences between the simulated groundwater discharge series for each climate scenario. For example, average annual groundwater discharge rates increase significantly for the CGCM3-A2 and CGCM3-A1B scenarios compared to the reference period simulation (increase of mean annual discharge approximately 60% for both configurations). These increases in discharge are not surprising given the significant projected increases in precipitation (50%, Figure 5.3a) and recharge (58%, Figure 5.3b) for these two climate scenarios. Changes to groundwater discharge on this order can significantly alter the total river discharge of baseflow-dominated rivers, and total river discharge has been shown to influence the thermal response of rivers to warming climates (Deitchman and Loheide 2012, van Vliet et al. 2011). Thus, changes to groundwater discharge may need to be accounted for when modeling future riverine thermal regimes. Figure 5.10 also indicates that the peak discharge rates for the CGCM3-A1B simulation increase by 53% for Configuration 1 and only 35% for Configuration 2. These differences arise because Configuration 2, which has a deeper discharge point than Configuration 1, attenuates the temporal variability in recharge and consequent discharge more than Configuration 1.
As Figure 5.10 illustrates, the average timing of the maximum discharge rate is shifted earlier in the year for most of the climate scenarios due to the earlier snowmelt simulated in HELP3 (Kurylyk and MacQuarrie 2013). These shifts are similar for Configuration 1 (range = -17 to +1 days, negative implies earlier discharge event) and Configuration 2 (range = -17 to +4 days). As indicated by the MIROC3.2-HIRES-A1B simulation (Figure 5.10), a shift in the timing of the major discharge event may cause shallow, unconfined aquifers to be more fully drained by early summer and thereby reduce early summer discharge rates.

5.5.6 Impact of climate change on the magnitude of summer groundwater discharge

Figure 5.10 indicates that future summer groundwater discharge rates simulated for Configuration 1 deviate more from the reference period simulation than summer groundwater discharge rates simulated for Configuration 2. For example, simulated changes in average summer groundwater discharge rates for Configuration 1 range from -6% to +31% with a mean of +8.4%, while simulated changes in average summer discharge for Configuration 2 range from -12% to +19% with a mean of +2.1%. The maximum simulated increases in summer groundwater discharge for both aquifers are significant (+31% and +19% for Configurations 1 and 2), but they are damped with respect to the increase in annual precipitation for this climate scenario (+50%, Figure 5.3A). This damping arises because much of the increase in annual precipitation for this particular climate scenario (CGCM3-A1B) occurs during the winter months. This precipitation temporarily accumulates in the form of snow and recharges the aquifer in
the spring. This increased aquifer storage is predominantly discharged prior to the summer months. Additionally, due to the projected increases in summer AT for the CGCM3-A1B scenario, simulated evapotranspiration rates increase and reduce the soil moisture available for recharge (Kurylyk and MacQuarrie 2013). In general, the physical dimensions of Configuration 2 cause it to be less hydraulically sensitive to increases in summer precipitation than Configuration 1.

5.5.7 Implications for groundwater-sourced salmonid refugia

The relationship between climate change, rising river water temperatures, and the loss of future cold-water fish habitat has been well-studied (e.g., Chu et al. 2005, Isaak et al. 2012, Jonsson and Jonsson 2009, Wu et al. 2012, and references therein). Given the importance of groundwater-sourced thermal refugia in warming riverine ecosystems, there remains a surprising paucity of studies investigating physical processes that generate refugia and their vulnerability to climate change. The results from the present study can be utilized to perform a preliminary analysis of the ecological implications of changes to the thermal and hydraulic regimes of unconfined, alluvial aquifers.

The results presented in Figure 5.10 indicate that summer groundwater discharge rates will increase or be relatively constant for most of the climate scenarios. However, for the MIROC3.2-HIRES-A1B climate scenario, the average magnitude of the simulated summer discharge for Configuration 2 decreases 12% relative to the reference period simulation (Figure 5.10b). Such a reduction in groundwater flow at concentrated points of discharge could reduce the spatial extent of groundwater-sourced thermal refugia, which can already be overpopulated during high temperature events.
A simple probabilistic approach can yield a first order estimate of the ecological impacts of the aquifer warming depicted in Figure 5.11. Juvenile Atlantic salmon do not typically aggregate within thermal refugia in the Little Southwest Miramichi River (LSW) and surrounding river systems until ambient river temperatures exceed 22°C (Cunjak et al. 2005). Surface water surveys conducted in the cold-water tributary considered in the present study (Otter Brook, Figure 5.1 and Configuration 1, Figure 5.4a) indicate that once groundwater discharges into the tributary it will experience significant downstream warming due to thermal mixing with warmer surface water and exposure to solar radiation. This warming can be on the order of 4°C in regions of significant groundwater discharge to the tributary. In the case of 4°C intermediate warming, groundwater discharge temperature to the tributary should not exceed 18°C (22°C - 4°C), or the refuge at the confluence with the LSW will not provide optimal conditions for thermally stressed salmonids. This target temperature (18°C) was selected to examine the probability that the aquifer discharging to the tributary (Configuration 1) could continue to provide suitable thermal habitat under the most extreme warming scenario (MIROC3.2-HIRES-A1B, Figure 5.3).
Figure 5.13. Summer (June 1-August 31) groundwater discharge temperature from Configuration 1 for the recent reference period (1991-2000, run R.1, Table 5.2) and MIROC3.2-HIRES-A1B (2056-2065, run 6.1) simulations vs. the probability that the temperature would be exceeded on any given day in the summer. The daily exceedance probabilities for a groundwater discharge temperature of 18°C are shown.

The simulated mean daily summer groundwater discharge temperatures for the reference period (run R.1) and the warmest climate scenario (run 6.1) were ranked and assigned a probability in accordance with the Weibull ranking method (McCuen, 1993, p. 60). Figure 5.13 indicates that the reference period probability of exceeding the 18°C target temperature on any given day in the summer is 9%, but this probability increases to 66% in the warming climate. Thus, this shallow, unconfined aquifer may not continue to provide optimal thermal refuge for the majority of the summer under the most extreme warming scenario. Note that this simplistic approach assumes that the future change in the water temperature of Otter Brook, which is baseflow-dominated, will correspond to the change in the groundwater discharge temperature. The actual future water temperature increases in Otter Brook may also be influenced by other factors, including changes to
future air temperature and precipitation regimes and land cover changes. A similar analysis conducted for Configuration 2 (results not shown), which discharges directly into the LSW and thus experiences no intermediate warming, indicates that the aquifer will continue to consistently provide a small thermal refuge under each of the climate scenarios considered in this study. These analyses should not be used to make definitive predictions regarding the future state of the Otter Brook or seep thermal refugia (Figure 5.1), but they do provide insight into how varying aquifer morphologies and refugia characteristics (e.g., direct or indirect groundwater discharge) may influence their thermal response to climate change.

5.5.8 Limitations

The modeling approach used in the present study (Figures 5.2 and 5.4) employed several simplifying assumptions. Firstly, the recharge output from HELP3 and the GST output from ForHyM2 were expressed as point conditions. Thus, the surface conditions were assumed to be thermally and hydraulically uniform, which is a reasonable assumption given the small size (< 10 km²) and low topographic relief of the catchments studied. Secondly, the recharging water was assumed to be in thermal equilibrium with GST. The temperature of precipitation and infiltration is influenced by the thermal regime of the lower atmosphere and may deviate from GST. Thirdly, the aquifer thermal and hydraulic conductivities were assumed to depend only on the moisture conditions or direction of groundwater flow, but saturated subsurface environments are characterized by heterogeneity in their thermal and hydraulic properties (Domenico and Schwartz 1990). Despite these limitations, the simulation results provide valuable insight into the
hydraulic and thermal sensitivities of unconfined aquifers to climate change and the
associated impacts to groundwater-sourced fish habitat.

5.6 Conclusions

The sensitivity of groundwater discharge to climate change was investigated by
utilizing downscaled climate scenarios to drive surface and subsurface simulations in
idealized catchments/aquifers whose properties were derived from field observations of
aquifers generating thermal refugia. To our knowledge, this research is the first to utilize
downscaled climate scenarios to form the boundary conditions for multi-dimensional
simulations of groundwater flow and heat transport.

Five main conclusions can be extracted from the results and discussion.

(1) The thermal regimes of thin shallow aquifers are not resilient to decadal
climate change. In certain instances, the simulated aquifer thermal
sensitivities to climate change (average values of 1.0 for Configuration 1
and 0.55 for Configuration 2) exceed those previously reported for rivers
and streams. Thus high groundwater discharge rates should not necessarily
be presumed to provide a buffer to future climate change. Furthermore,
changes to groundwater discharge temperature will not necessarily closely
track changes to AT in seasonally snow-covered catchments due to the
complex dynamics of snowpack evolution.
(2) The lag between a rise in AT and the thermal response of shallow groundwater may be overestimated in previous studies. Groundwater discharge from the aquifers in the present study reached thermal equilibrium with changing climatic conditions in less than five years. Therefore groundwater temperature evolution due to climate change may need to be considered in deterministic models of surface water thermal regimes.

(3) In catchments experiencing seasonal freeze-thaw, daily groundwater discharge temperature can be impacted by advection, conduction, and the dynamic freeze-thaw process. Thus, future research investigating the sensitivity of shallow groundwater temperature to climate change should not employ simplified conduction models unless they have been shown to perform well for the aquifer being considered.

(4) The timing and magnitude of groundwater discharge may also respond quickly to changes in precipitation and groundwater recharge. Projected changes in the magnitude of mean annual recharge and discharge range from approximately -6% to +60% compared to the reference period recharge. The peak groundwater discharge event associated with snowmelt is generally shifted earlier in the year. Discharge from these aquifers is sensitive to changes in groundwater recharge on temporal scales ranging from weekly to decadal. Changes to groundwater discharge magnitude and timing will influence the thermal regimes of baseflow-dominated streams.
(5) The ability of groundwater discharge to continue to produce biologically significant thermal refugia in a warming climate depends strongly on the aquifer configuration, refuge characteristics (indirect or direct groundwater discharge), and the particular climate scenario employed. Furthermore, we have demonstrated that the probability of exceeding critical temperature thresholds within certain thermal refugia will likely increase in the coming decades. Although these results are site specific, they should challenge researchers to reconsider their preconceptions regarding the future states of groundwater dependent ecosystems.

Although shallow groundwater temperature will respond to a gradually warming climate, groundwater discharge will continue to buffer river temperatures throughout the summer months, particularly during extreme events. Consequently, the ecological significance of groundwater-surface water interactions should be considered in future comprehensive land-use management strategies (e.g., Nichols et al. 2013, Saltveit and Brabant 2013). Catchments overlying unconfined aquifers generating thermal refugia should be protected from anthropogenic activity such as deforestation and groundwater and aggregate extraction, as these processes have been shown to increase shallow groundwater temperature (Alexander et al. 2003, Barlow and Leake 2012, Markle and Schincariol 2007, Moore et al. 2005, Risley et al. 2010). The results from the present study further indicate that the thermal regimes of certain unconfined aquifers will be more resilient to climate change than others. Thus, our results suggest that knowledge of local hydrogeology will assist in prioritizing appropriate catchments for the preservation of
future thermal refugia. Further research is required to provide additional information on
the preservation and enhancement of groundwater-sourced thermal refugia.

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6.1 Recommendations for Future Research

There are several possibilities for advancing the research detailed in this dissertation, and these primarily relate to the atmospheric, surface and subsurface modelling previously described. Firstly, future studies should consider a larger array of climate scenarios to obtain probabilistic information regarding the future state of aquifer thermal and hydraulic regimes. In accordance with the recommendations of Chapter 2, climate modeling combinations (i.e., GCMs and downscaling methods) that have inadequately reproduced recent climate conditions (e.g., 2000-2010) should be abandoned in favor of GCMs and downscaling methods that produce more realistic results. Probabilistic information (e.g., mean and quartiles for mean annual air temperature and precipitation) and the extreme scenarios (e.g., the driest, wettest, coldest or warmest climates) could be extracted from a large suite of scenarios. It should be noted that such a study could generally only be conducted in close collaboration with climatologists as most geographic locations have few downscaled climate scenarios available.

Secondly, the analytical solution proposed in Chapter 3 could be coded into software such as R (R Core Team 2013) or Matlab (MathWorks Inc. 2012) to improve the efficiency of applying this solution in future analysis. The program could read in text files for measured temperature-depth profiles and the climate projections for a particular region and assign ‘best fits’ with the exponential initial condition and boundary condition functions proposed in Chapter 3. The code could simulate the subsurface temperature evolution in wells with available data for equation parameterization (e.g., thermal...
properties and vertical Darcy velocity) and could be used to facilitate a global study of the subsurface thermal impacts of climate change and urbanization.

Thirdly, the empirical ground surface temperature to groundwater temperature transfer function described in Chapter 4 could be tested in other forested catchments to determine how site conditions (e.g., climate, recharge rates, soil lithology, and depth to groundwater table) influence the empirically-derived fitting parameters. The function could also be tested against higher spatial resolution (e.g., 25 cm) groundwater temperature measurements to determine the variability of the parameters with depth. Furthermore, the recharge and ground surface temperature modeling could be combined in a single soil-vegetation-atmosphere transfer model rather than employing separate models (HELP3 and ForHyM2) as described in Chapters 2 and 4.

Fourthly, the numerical modeling results presented in Chapter 5 for aquifer Configuration 1 (Otter Brook) only consider changes to the groundwater discharge temperature. However, the thermal regime of the refugium at the mouth of Otter Brook is also influenced by heat fluxes in the brook (e.g., long and shortwave radiation and surface water thermal mixing). Thus, a more detailed modeling approach could consider the entire aquifer-river-land system and account for all thermal exchanges in the surface water. Such an endeavor would be ambitious and could easily provide the basis for another dissertation.

Finally, the present study focused on summer cold-water thermal refugia. Future research could also consider groundwater-sourced warm thermal refugia. It is well known that relatively warm groundwater discharge to streams and rivers during the winter months generates stable temperatures for salmonid egg incubation and provides critical
habitat for brook trout and juvenile Atlantic salmon by inhibiting ice formation in channels (e.g., Brown et al. 2011, Cunjak and Power 1986, Power et al. 1999). The decline in the temperature of winter groundwater discharge simulated for several climate scenarios (Chapter 5) may negatively impact overwintering salmonids and salmonid egg survival.

6.2 Recommendations for Land-Use and River Management Strategies in New Brunswick, Canada

6.2.1 Public and scientific support for the preservation of salmonid thermal refugia

The importance of thermal refugia for providing critical salmonid habitat in Canadian rivers has been well-documented in scientific literature (Breau et al. 2011, Cunjak et al. 2005, Gibson 1966, Monk et al. 2013), and the general public is now becoming increasingly aware of the vital function of cold-water refugia (e.g., Hume 2011). In particular, Atlantic Canadians are cognizant of the many threats to Atlantic salmon survival and are willing to invest in publically-funded habitat restoration or enhancement programs. For example, Gardner Pinfold Consultants Inc. (2011) found that there was ‘over 80% public support in Eastern Canada for a sustained 20-year program... [to] restore Atlantic salmon abundance to 40%-80% of historic highs.’

Many researchers have recently noted that groundwater and/or cold-water refugia should specifically be considered in effective land-use or river management strategies for preserving salmonid habitat (e.g., Goniea et al. 2006, Nichols et al. 2013, Saltveit and Braband 2013). For example, Brewer (2013) suggested that groundwater-sourced thermal
refugia should be ‘recognized for their biological potential to provide suitable habitat as climate change and other land-use alterations increase temperature regimes and alter flow patterns’.

Despite the increased recognition of the need to include groundwater-sourced thermal refugia in comprehensive land-use or river management strategies, very few guidelines have been proposed that describe how this can be practically accomplished. A notable exception is Torgersen et al. (2012) who suggested that incorporated policy for the protection and restoration of refugia should consider ‘(1) physical factors, such as riparian condition and hydrologic connectivity in the floodplain, (2) biological interactions with native and non-native species that may influence the effectiveness of cold-water refuges for cold-water fish, (3) effects of other aquatic species and life history stages, and (4) unanticipated human effects, such as climate change’. The research contained in this dissertation can provide guidance on the last point noted above (i.e., 4). The following paragraphs describe how the findings of the present study could be incorporated into a land-use or river management strategies for the protection of Atlantic salmon and brook trout in New Brunswick.

6.2.2 Preserving existing cold-water refugia

It is extremely unlikely that every existing thermal refugium will be protected via stricter land-use management policies. Such an endeavor would consume excessive time and financial resources. Consequently, a hierarchy must be established to prioritize the preservation of certain refugium morphologies. In accordance with the recommendations by Torgersen et al. (2012), such a hierarchy must consider not only refuge-generating
capabilities for the present climate, but also their resilience to future climate change. The former information can be easily obtained by conducting in-stream thermals surveys and monitoring during high temperature events, but the latter should be determined through site-specific physically-based modeling or estimated from modeling results obtained for representative aquifers.

In lieu of extensive modeling, temperature loggers located at points of groundwater discharge can indicate the groundwater discharge thermal damping factor (ratio of the amplitude of the annual groundwater discharge temperature cycle to the amplitude of the annual AT cycle), which can then be utilized to estimate the thermal sensitivity of a refugium to climate change. In seasonally snow-covered catchments, refugia generated from deeper groundwater systems (lower thermal damping factor) will typically exhibit decreased sensitivity to changes in both summer AT and precipitation in comparison to refugia generated from shallower systems (Chapter 5). The land cover overlying the catchments generating resilient cold-water thermal refugia should be prioritized for preservation provided that these refugia have been shown to be functional in the present climate. Preservation strategies could include limiting deforestation, urbanization, and aggregate extraction within these catchments as these activities has been shown to increase groundwater temperature (Bourque and Pomeroy 2001, Ferguson and Woodbury 2007, Gunawardhana et al. 2011, Markle and Schincariol 2007, Taniguchi and Uemura 2005, Taniguchi et al. 2008).
### 6.2.3 Augmentation or creation of cold-water refugia

In many river reaches, there is currently an inadequate distribution of suitable thermal refugia (Martins et al. 2011). In these cases, effective refugia management options could include (1) augmenting existing riverine thermal anomalies that are not functioning as refugia and/or (2) creating new refugia in river reaches that lack thermal anomalies.

Thermal anomalies may not function as thermal refugia due to a variety of physical conditions that are not conducive to salmonid health. Thermally-stressed salmonids often encounter a trade-off between experiencing thermal refuge and increasing their exposure to predators or relatively anoxic water (Ebersole et al. 2001, 2003, Keefer et al. 2009, Matthews and Berg 1997), and these trade-offs can delay migration into refugia until river temperatures reach lethal thresholds. Existing cold-water plumes that never (or rarely) function as thermal refugia can be augmented so that their physicochemical conditions comply with salmonid habitat requirements (e.g., Armstrong et al. 2003). For example, channel deflectors installed in the mainstem near the mouth of cold-water tributaries can limit thermal mixing in the confluence and thereby increase the spatial extent of the cold-water plume (Bilby 1984). Other long-term options to improve the suitability of cold-water plumes for salmonid habitat include: (1) reforesting catchments that overly aquifers generating cold-water plumes, (2) increasing riparian zone shading, (3), increasing refugia depth to limit salmonid visibility and (4) installing appropriately-sized substrate for salmonid concealment.

This dissertation has highlighted the importance of groundwater discharge for inducing riverine thermal diversity, and the hydrogeological principles addressed in this
research can be applied to create new thermal refugia in river reaches that are thermally uniform. For example, ‘engineered’ thermal refugia can be created by pumping groundwater from adjacent alluvial aquifers to predetermined locations along the river. Pumping water from an adjacent aquifer will not necessarily increase or decrease the total groundwater input to the river; rather it will transform groundwater discharge from a diffusive input that slightly cools the entire river to a concentrated input that significantly cools a smaller plume. Engineered thermal refugia should be protected via strict legislation to prevent exploitation via the sports fishery. A general flow chart for the identification, augmentation, creation, prioritization and preservation of suitable cold-water thermal refugia is presented in Figure 6.1.
(a) Identify thermal anomalies

A1. A priori predictions via a landscape variable model

A2a. In-stream thermal survey

A2b. Aerial infrared survey

A3. Do potential refugia exist?

B1. Augment existing thermal heterogeneities

B1a. Limit advective thermal mixing by installing channel deflectors

B1b. Increase riparian zone vegetation to limit solar radiation and salmonid visibility

B1c. Install appropriate substrate to induce surface turbulence and provide salmonid cover

B2. Create refugia by inducing concentrated groundwater discharge

B3. Monitor to ascertain if augmented or created refugia are functional

C1. Identify refugium-generating mechanisms (e.g., direct or indirect groundwater discharge?)

C2. Prioritize which refugia to protect via hydrogeological analysis, thermal monitoring, or ecologically based spatial distribution analysis.

C3. Inhibit anthropogenic activities in prioritized catchments

C4. Monitor to ascertain if protection and enhancement measures are effective

(b) Augment or create refugia

(c) Protect refugia

No

Yes

No

Yes

Figure 6.1. A flow chart for (a) identifying/exploring for, (b) augmenting/creating, and (c) preserving cold-water thermal refugia.
References


CHAPTER 7: Conclusions

7.1 Contributions to Science

The influence of atmospheric climate change on groundwater resources is a matter of international concern. Recent studies (e.g., Crosbie et al. 2011, 2013, Döll 2009) have employed an array of climate scenarios to drive hydrological models and produce a range of future groundwater recharge projections. Uncertainty arises when projecting future groundwater recharge because (1) precipitation is poorly resolved in GCMs and (2) the change in recharge may be significantly amplified compared to the change in precipitation (Crosbie et al. 2013).

While a general understanding of the relationship between future climate change and groundwater recharge has been established from these previous contributions, they did not investigate the influence of climate change on groundwater recharge caused by varying all three main climate modeling components (emission scenarios, GCM/RCM’s, and post-processing methods). Varying only one or two of these climate modeling options obscures the true uncertainty in the future climatic regime and the resultant changes to groundwater recharge. Furthermore, no synthesis of these recent groundwater recharge modeling studies had been produced prior to the research presented in this dissertation. Chapter 2 addresses the limitations of previous analyses noted above by first synthesizing the current body of knowledge related to modeling future groundwater recharge and then adding to this body by considering the effects of varying the emission scenario, GCM, and post-processing method and by performing simulations in a climate
where snow accumulation and melt are important hydrological processes. The simulation results demonstrate that the projected changes in mean annual recharge for the Otter Brook catchment may vary between -6% to +58% for 2045-2065 in comparison to the reference period (1961-2000) simulation. Thus, like many other regions of the world, it is difficult to predict the magnitude or even direction of future groundwater recharge changes in New Brunswick.

The influence of atmospheric climate change on groundwater temperature is also a matter of international concern. This concern arises because there is a link between groundwater temperature and groundwater quality due to the temperature-dependence of many subsurface biogeochemical reactions (Andrushchyshyn et al. 2009, Rike et al. 2008, Sharma et al. 2012). Furthermore, groundwater temperature exhibits a relative lack of seasonal variability in comparison to surface water temperature. Thus, groundwater-surface water interactions can induce riverine thermal diversity, which can provide thermal refuge for cold-water fishes in otherwise thermally uninhabitable river reaches. Groundwater-sourced refugia are already important in many regions of the globe, including eastern North America (Breau et al. 2011, Gibson 1966), western North America (Sutton et al. 2007, Tanaka 2007), and New Zealand (Olsen and Young 2009), and salmonid reliance on thermal refugia is projected to increase in a warming climate.

Chapter 3 details the development of an analytical solution that is well-suited for investigating climate change-induced groundwater temperature perturbations in deeper aquifer systems characterized by predominantly one-dimensional groundwater flow and heat transport. This solution was applied to characterize the thermal evolution of temperature-depth profiles beneath the Sendai Plain and Tokyo, Japan. Results obtained
with the new analytical solution deviate significantly from those obtained with the classic solution (Taniguchi et al. 1999), thereby demonstrating the importance of the flexible boundary and initial conditions employed.

Chapter 4 describes the influence that atmospheric climate change will have on the ground surface thermal regime. The modeling results suggest that rising mean annual air temperature (AT) and associated snowpack evolution may paradoxically lead to increased summer ground surface temperature (GST) and decreased winter GST. This decoupling has profound implications for borehole temperature inversion studies that infer paleoclimates by assuming that decadal GST changes follow AT trends (Chapman et al. 2004, Stieglitz and Smerdon 2007), and thus these findings are significant in the broader study of climate change impacts.

Chapter 4 also details the development of an innovative empirical GST to groundwater temperature transfer function and the application of this function to estimate future groundwater temperature. The form of this empirical function was derived from a physical understanding of subsurface heat transport processes. The results of this study strongly suggest that the thermal sensitivity of groundwater to projected (2046-2065) climate change may, in certain circumstances, exceed that of nearby surface water.

Chapter 5 discusses the linking of process-oriented atmosphere, surface, and subsurface models to investigate the influence of climate change on the timing, magnitude, and temperature of groundwater discharge. The groundwater flow and heat transport numerical modeling results generally confirm the findings of Chapter 4 by indicating that groundwater temperatures in shallow aquifers can be very sensitive to increases in AT and that baseflow-dominated streams and rivers may actually warm more
than direct flow-dominated streams and rivers. The numerical model simulations further demonstrate how an aquifer’s configuration will influence its response to atmospheric climate change. The thermal regimes of thicker aquifers, which are driven by mean annual GST, will exhibit less sensitivity to increases in summer AT or GST than thinner aquifers, which are strongly influenced by seasonal GST variability. Chapter 5 also considers the influence of the seasonal soil water freeze-thaw cycle on subsurface thermal and hydraulic regimes. The results indicate that pore water phase change strongly controls the upper 2 m of the subsurface thermal regime within the Otter Brook catchment during the winter months and that cryogenic processes should be included when simulations are performed for very shallow groundwater. Chapter 6 suggests several opportunities for preserving, augmenting, and creating thermal refugia based on the primary findings of the other chapters. The objectives and primary research findings of this dissertation are summarized in Table 7.1.
<table>
<thead>
<tr>
<th>Objective</th>
<th>Primary findings</th>
</tr>
</thead>
<tbody>
<tr>
<td>1. Estimate future groundwater recharge via a daily soil water balance model and quantify the associated uncertainty (Chapter 2).</td>
<td>Change in future recharge = -6 to +58%; resulting in an uncertainty of 64%; Downscaling/post-processing yielded the greatest uncertainty; It is difficult to predict the magnitude or direction of future groundwater recharge.</td>
</tr>
<tr>
<td>2. Develop and apply an improved analytical solution to the conduction-advection equation (Chapter 3).</td>
<td>Initial and boundary conditions are better represented with exponential functions; Including the exponential functions contributes more complexity to the resultant solution but improves model fidelity; Simulations performed with the improved solution deviate from those obtained with the classic solution for the scenario considered.</td>
</tr>
<tr>
<td>3. Examine the role of changing AT and snow cover in controlling future GST via a process-oriented, surficial heat transfer model (Chapter 4).</td>
<td>The duration and thickness of snowpack will decrease in a warming climate (maximum decrease of 49 days and 77%, respectively); Changes in summer GST mimic changes in summer AT, but changes in winter GST do not follow changes in winter AT; Trends in mean annual GST and AT are decoupled. The average simulated increases in mean annual AT and GST were 2.11°C and 1.0°C, respectively.</td>
</tr>
<tr>
<td>4. Develop an empirical GST to groundwater temperature transfer function and investigate trends in future groundwater temperature (Chapter 4).</td>
<td>Measured groundwater temperature and GST can be employed to produce a reasonably accurate empirical GST to groundwater temperature function; The results indicate that changes to monthly groundwater temperature can be significant (up to ~ +3.5°C).</td>
</tr>
<tr>
<td>5. Link surface models to a finite element model of subsurface water flow and energy transport to investigate changes to the timing, magnitude and temperature of groundwater discharge (Chapter 5).</td>
<td>Rising AT will cause an earlier snowmelt which will in turn shift the timing of the major groundwater discharge event; Uncertainty in changes to the magnitude of recharge (Chapter 2) results in uncertainty in changes to the magnitude of mean annual or seasonal groundwater discharge; Groundwater temperature will respond to rising air temperature, but the sensitivity is dependent on the aquifer configuration; The simulated sensitivities suggest that the thermal regimes of baseflow-dominated streams may be less resilient to climate change than previously imagined.</td>
</tr>
</tbody>
</table>
References


Appendix 1: Derivation of an Analytical Solution to the Transient, One-Dimensional Conduction-Advection Equation
Derivation of the analytical solution (Equation 3.8)

The governing equation for one dimensional groundwater flow and heat transport in a homogeneous medium with constant groundwater velocity can be written in the following form:

\[
\lambda \frac{\partial^2 T}{\partial z^2} - q c_w \rho_w \frac{\partial T}{\partial z} = c \rho \frac{\partial T}{\partial t}
\]  

(A1.1)

where \( \lambda \) is thermal conductivity (M L t\(^{-3}\) T\(^{-1}\)), \( T \) is temperature, \( z \) is depth below ground surface (L), \( t \) is time, \( q \) is the Darcy velocity (L t\(^{-1}\)), \( c_w \rho_w \) is the volumetric heat capacity of the water, and \( c \rho \) is the volumetric heat capacity of the medium (M L\(^{-1}\) t\(^2\) T\(^{-1}\)). This equation can be expressed in the standard form employed by Carslaw and Jaeger (1959) and those who have compiled the analytical solutions to advection-dispersion equations:

\[
D \frac{\partial^2 T}{\partial z^2} - \nu \frac{\partial T}{\partial z} = \frac{\partial T}{\partial t}
\]  

(A1.2)

where \( D \) is the thermal diffusivity \( \lambda/(c \rho) \) (L\(^2\) t\(^{-1}\)) and \( \nu = q c_w \rho_w/(c \rho) \) (L t\(^{-1}\)). The initial and boundary conditions can be expressed as follows:

\[
T(z, t = 0) = T_i + a z + \delta \exp(dz)
\]  

(A1.3)

\[
T(z = 0, t) = T_i + b \exp(ct)
\]  

(A1.4)

To reduce the partial differential equation to an ordinary differential equation, the Laplace transform of each term in Equation A1.2 can be taken with respect to \( t \) with \( z \) treated as a constant (Farlow 1982):

\[
L\left[ D \frac{\partial^2 T}{\partial z^2} \right] = D \frac{\partial^2 \tilde{T}}{\partial z^2}
\]  

(A1.5)
Thus, the subsidiary equation becomes:

\[
D \frac{\partial^2 \tilde{T}}{\partial z^2} - v \frac{\partial \tilde{T}}{\partial z} - \tilde{p} \tilde{T} = -T_i - a z - \delta \exp(dz) \tag{A1.8}
\]

Equation A1.8 can be rewritten in the following form:

\[
\frac{\partial^2 \tilde{T}}{\partial z^2} - \frac{v}{D} \frac{\partial \tilde{T}}{\partial z} - \frac{p}{D} \tilde{T} = -T_i - a z - \delta \exp(dz) \tag{A1.9}
\]

This non-homogeneous linear ordinary differential equation can be solved by the method of undetermined coefficients (Zill 2005). The homogeneous auxiliary equation is:

\[
\frac{\partial^2 \tilde{T}}{\partial z^2} - \frac{v}{D} \frac{\partial \tilde{T}}{\partial z} - \frac{p}{D} \tilde{T} = 0 \tag{A1.10}
\]

The solution to Equation A1.10 is known as the complementary function. It can be shown to be (Zill 2005):

\[
\tilde{T}_c = A \exp(m_1 z) + B \exp(m_2 z) \tag{A1.11}
\]

where \(m_1\) and \(m_2\) are the roots of Equation A1.10 when the differential operator is conceptualized as a parameter:

\[
m_1 = \frac{1}{2} \left\{ \frac{v}{D} + \sqrt{\frac{v^2}{D^2} + \frac{4p}{D}} \right\} \quad \text{and} \quad m_2 = \frac{1}{2} \left\{ \frac{v}{D} - \sqrt{\frac{v^2}{D^2} + \frac{4p}{D}} \right\} \tag{A1.12}
\]

The domain in question is spatially semi-infinite, thus \(A\) must equal zero to allow for a bounded solution as \(z\) approaches infinity. Therefore, the complementary function is:
The particular solution can now be found if the parameters from the initial condition are inserted back into the governing subsidiary equation. Because the form of the right hand side of Equation A1.9 is a first order polynomial with an exponential function, the particular solution is assumed to be of the following form (Zill 2005):

\[ \bar{T}_p = \kappa z + \sigma + \eta \exp(dz) \quad (A1.14) \]

\[ \frac{\partial \bar{T}_p}{\partial z} = \kappa + nd \exp(dz) \quad \text{and} \quad \frac{\partial^2 \bar{T}_p}{\partial z^2} = \eta d^2 \exp(dz) \quad (A1.15) \]

These terms can be inserted into Equation A1.9:

\[ \eta d^2 \exp(dz) - \frac{v}{D} (\kappa + \eta d \exp(dz)) - \frac{p}{D} (\kappa z + \sigma + \eta \exp(dz)) = \frac{-T_i - az - \delta \exp(dz)}{D} \quad (A1.16) \]

Equation A1.16 can be further simplified:

\[ (-D\eta d^2 + v\eta d + p\eta) \exp(dz) + \nu \kappa + p\kappa z + p\sigma = T_i + a z + \delta \exp(dz) \quad (A1.17) \]

Equation A1.17 can be used to solve for the unknown parameters:

\[ p\kappa z = az \rightarrow \kappa = \frac{a}{p} \quad (A1.18) \]

\[ \frac{\nu}{p} a + p\sigma = T_i \rightarrow \sigma = \frac{T_i}{p} - \frac{va}{p^2} \quad (A1.19) \]

\[ \left(-D\eta d^2 + v\eta d + p\eta\right) = \delta \rightarrow \eta = \frac{\delta}{-Dd^2 + vd + p} \quad (A1.20) \]

Therefore, the particular solution can be shown to be of the form shown below:

\[ \bar{T}_p = \frac{az}{p} + \frac{T_i}{p} - \frac{va}{p^2} + \frac{\delta}{-Dd^2 + vd + p} \exp(dz) \quad (A1.21) \]
The solution to the subsidiary equation is the superposition of the complementary and particular solutions:

\[
\bar{T} = \bar{T}_c + \bar{T}_p = B \exp \left[ \left( \frac{\nu}{D} - \sqrt{\frac{v^2}{D^2} + \frac{4p}{D}} \right) \frac{z}{2} \right] + a \frac{z}{p} + \frac{T_i}{p} - \frac{va}{p^2} + \frac{\delta}{-Dd^2 + vd + p} \exp(dz) \quad (A.122)
\]

The unknown \( B \) parameter can be determined by taking the Laplace transform of the boundary condition:

\[
L[T(z = 0,t) = T_i + b \exp(\nu t)] = \frac{T_i}{p} + \frac{b}{p-c} \quad (A.123)
\]

\[
\bar{T}(z = 0, p) = \left\{ B + \frac{T_i}{p} - \frac{va}{p^2} + \frac{\delta}{(-Dd^2 + vd + p)} \right\} = \left\{ \frac{T_i}{p} + \frac{b}{p-c} \right\} \quad (A.124)
\]

\[
\therefore B = \frac{T_i - T_i}{p} + \frac{va}{p^2} + \frac{b}{p-c} - \frac{\delta}{(-Dd^2 + vd + p)} \quad (A.125)
\]

The \( B \) term can now be entered into Equation A1.22 to give the solution to the subsidiary equation:

\[
\bar{T} = \left( \frac{T_i - T_i}{p} + \frac{va}{p^2} + \frac{b}{p-c} - \frac{\delta}{(-Dd^2 + vd + p)} \right) \exp \left[ \left( \frac{\nu}{D} - \sqrt{\frac{v^2}{D^2} + \frac{4p}{D}} \right) \frac{z}{2} \right] + \frac{az}{p} + \frac{T_i}{p} - \frac{va}{p^2} + \frac{\delta}{-Dd^2 + vd + p} \exp(dz) \quad (A.126)
\]

This can be rewritten in the following form:

\[
\bar{T} = \left( \frac{az}{p} + \frac{T_i}{p} - \frac{va}{p^2} + \frac{\delta}{-Dd^2 + vd + p} \right) \exp(dz) + \exp \left( \frac{vz}{2D} \right) \times \left( \frac{T_i - T_i}{p} + \frac{va}{p^2} + \frac{b}{p-c} - \frac{\delta}{(-Dd^2 + vd + p)} \right) \exp \left[ -z \sqrt{\frac{v^2}{4D^2} + \frac{p}{D}} \right] \quad (A.127)
\]

The shift theorem can be applied to convert Equation A1.27 into a form for which inverse Laplace transforms are tabulated (Trim 1990):
Let \( P = \frac{v^2}{4D} + p \) and \( p = P - \frac{v^2}{4D} \) (A1.28)

\[
T = \exp\left(\frac{-v^2}{4D}\right) \cdot L^{-1}\{E\} + \exp\left(\frac{-v^2}{4D}\right) \cdot \exp\left(\frac{vz}{2D}\right) \cdot \left[L^{-1}\{F\} + L^{-1}\{G\} + L^{-1}\{H\} + L^{-1}\{I\}\right] \quad (A1.29)
\]

The \( E, F, G, H \) and \( I \) terms are defined below:

\[
E = \left(\frac{az}{P - \frac{v^2}{4D}} + \frac{T_i}{P - \frac{v^2}{4D}} - \frac{va}{(P - \frac{v^2}{4D})^2} + \frac{\delta}{(-Dd^2 + vd + P - \frac{v^2}{4D})} \exp(dz)\right) \quad (A1.30)
\]

\[
F = \frac{T_i - T_i}{(P - \frac{v^2}{4D})} \cdot \exp\left\{-z \sqrt{\frac{P}{D}}\right\} \quad (A1.31)
\]

\[
G = \left(\frac{va}{(P - \frac{v^2}{4D})^2}\right) \cdot \exp\left\{-z \sqrt{\frac{P}{D}}\right\} \quad (A1.32)
\]

\[
H = \left(\frac{b}{P - \frac{v^2}{4D} - c}\right) \cdot \exp\left\{-z \sqrt{\frac{P}{D}}\right\} \quad (A1.33)
\]

\[
I = \left(-\frac{\delta}{(-Dd^2 + vd + P - \frac{v^2}{4D})}\right) \cdot \exp\left\{-z \sqrt{\frac{P}{D}}\right\} \quad (A1.34)
\]

The inverse Laplace functions of \( E, F, G, H \) and \( I \) with respect to \( P \) are shown below (Carslaw and Jaeger 1959, Zill 2005):

\[
216
\]
\[ L^1\{E\} = \alpha \exp\left(\frac{\nu^2}{4D}\right) + T_r \exp\left(\frac{\varphi^2}{4D}\right) - \nu \exp\left(\frac{\nu^2}{4D}\right) + \delta \exp(dz) \exp\left(\frac{\nu^2}{4D} + Dd^2 - \nu d\right) \]  
(A1.35)

\[ L^1\{F\} = \frac{(T_i - T_r)}{2} \exp\left(\frac{\nu^2}{4D}\right) \left\{ \exp\left(-z\sqrt{\frac{\nu^2}{4D^2}}\right) \text{erfc}\left(\frac{z}{2\sqrt{Dt}} - \sqrt{\frac{\nu^2}{4D}}\right) + \right. \right. \]  
\[ \left. \left. \exp\left(z\sqrt{\frac{\nu^2}{4D^2}}\right) \text{erfc}\left(\frac{z}{2\sqrt{Dt}} + \sqrt{\frac{\nu^2}{4D}}\right) \right\} \]  
(A1.36)

\[ L^1\{G\} = \frac{v_0}{2} \exp\left(\frac{\nu^2}{4D}\right) \left\{ \left(\frac{t - \frac{z}{\sqrt{\nu^2/4}}}{\sqrt{\nu^2/4}}\right) \exp\left(-z\sqrt{\frac{\nu^2}{4D^2}}\right) \text{erfc}\left(\frac{z}{2\sqrt{Dt}} - \sqrt{\frac{\nu^2}{4D}}\right) + \right. \right. \]  
\[ \left. \left. \left(\frac{t + \frac{z}{\sqrt{\nu^2/4}}}{\sqrt{\nu^2/4}}\right) \exp\left(z\sqrt{\frac{\nu^2}{4D^2}}\right) \text{erfc}\left(\frac{z}{2\sqrt{Dt}} + \sqrt{\frac{\nu^2}{4D}}\right) \right\} \]  
(A1.37)

\[ L^1\{H\} = \frac{b}{2} \exp\left(\frac{\nu^2}{4D^2} + c\right) \left\{ \exp\left(-z\sqrt{\frac{\nu^2}{4D^2} + c/D}\right) \text{erfc}\left(\frac{z}{2\sqrt{Dt}} - \sqrt{\frac{\nu^2}{4D}}\right) + \right. \right. \]  
\[ \left. \left. \exp\left(z\sqrt{\frac{\nu^2}{4D^2} + c/D}\right) \text{erfc}\left(\frac{z}{2\sqrt{Dt}} + \sqrt{\frac{\nu^2}{4D}}\right) \right\} \]  
(A1.38)

\[ L^1\{I\} = -\frac{\delta}{2} \exp\left(\frac{\nu^2}{4D} + Dd^2 - \nu d\right) \left\{ \exp\left(-z\sqrt{\frac{\nu^2}{4D^2} + d^2 - \nu d/D}\right) \text{erfc}\left(\frac{z}{2\sqrt{Dt}} - \sqrt{\frac{\nu^2}{4D}}\right) + \right. \right. \]  
\[ \left. \left. \exp\left(z\sqrt{\frac{\nu^2}{4D^2} + d^2 - \nu d/D}\right) \text{erfc}\left(\frac{z}{2\sqrt{Dt}} + \sqrt{\frac{\nu^2}{4D}}\right) \right\} \]  
(A1.39)

Therefore, the solution to Equation A1.2 under the given initial and boundary equations is the following:
\[ T = \exp\left(\frac{-v^2 t}{4D}\right) \left[ a z \exp\left(\frac{v^2 t}{4D}\right) + T_i \exp\left(\frac{v^2 t}{4D}\right) - v a t \exp\left(\frac{v^2 t}{4D}\right) + \delta \exp(z) \exp\left(\frac{v^2 t + Dd^2 t - v d t}{2D}\right) \right] + \exp\left(\frac{-v^2 r}{4D}\right) \exp\left(\frac{v z}{2D}\right) \times \]
\[
\left[ \frac{(T_i - T)}{2} \exp\left(\frac{v^2 t}{4D}\right) \exp\left(-\frac{v z}{2D}\right) \operatorname{erfc}\left(\frac{z}{2\sqrt{D}t} - \frac{v t}{2 \sqrt{D}}\right) + \exp\left(\frac{v z}{2D}\right) \operatorname{erfc}\left(\frac{z}{2\sqrt{D}t} + \frac{v t}{2 \sqrt{D}}\right) \right] + \]
\[
\frac{v a}{2} \exp\left(\frac{v^2 t}{4D}\right) \left[ \left( t - \frac{z}{2\sqrt{D} t} \right) \exp\left(-\frac{z}{\sqrt{4D}t}\right) \operatorname{erfc}\left(\frac{z}{2\sqrt{D}t} - \sqrt{\frac{v^2 t}{4D}}\right) + \right] \]
\[
\frac{v a}{2} \exp\left(\frac{v^2 t}{4D}\right) \left[ \left( t + \frac{z}{2\sqrt{D} t} \right) \exp\left(\frac{z}{\sqrt{4D}t}\right) \operatorname{erfc}\left(\frac{z}{2\sqrt{D}t} - \sqrt{\frac{v^2 t}{4D}}\right) + \right] \]
\[
\frac{b}{2} \exp\left(\frac{v^2 t}{4D} + c r\right) \left[ \left( t - \frac{z}{2\sqrt{D} t} \right) \exp\left(-\frac{z}{\sqrt{4D}t}\right) \operatorname{erfc}\left(\frac{z}{2\sqrt{D}t} - \sqrt{\frac{v^2 t}{4D}}\right) + \right] \]
\[
\frac{b}{2} \exp\left(\frac{v^2 t}{4D} + c r\right) \left[ \left( t + \frac{z}{2\sqrt{D} t} \right) \exp\left(\frac{z}{\sqrt{4D}t}\right) \operatorname{erfc}\left(\frac{z}{2\sqrt{D}t} - \sqrt{\frac{v^2 t}{4D}}\right) + \right] \]
\[
\frac{\delta}{2} \exp\left(\frac{v^2 t}{4D} + Dd^2 - v d t\right) \left[ \left( t - \frac{z}{2\sqrt{D} t} \right) \exp\left(-\frac{z}{\sqrt{4D}t}\right) \operatorname{erfc}\left(\frac{z}{2\sqrt{D}t} - \sqrt{\frac{v^2 t}{4D}}\right) + \right] \]
\[
\frac{\delta}{2} \exp\left(\frac{v^2 t}{4D} + Dd^2 - v d t\right) \left[ \left( t + \frac{z}{2\sqrt{D} t} \right) \exp\left(\frac{z}{\sqrt{4D}t}\right) \operatorname{erfc}\left(\frac{z}{2\sqrt{D}t} - \sqrt{\frac{v^2 t}{4D}}\right) + \right] \]
\]

This can be extensively simplified:
References


Appendix 2: ForHyM2 Model
Mechanics and Parameterization
Abstract

The first section in Appendix 2 describes the model mechanics for simulating both thermal and hydrologic processes in ForHyM2. The second section lists the ForHyM2 input parameters utilized in the present study.

ForHyM2 Mechanics

The following description of the model mechanics is based primarily on the papers by Arp and Yin (1992) and Yin and Arp (1993). It should be noted that the model has been updated from a monthly model to a daily model. The thermal processes simulated in ForHyM2 are indicated in Figure A2.1. In ForHyM2, the surface boundary temperature is obtained by rearranging a simplified energy balance equation (Yin and Arp 1993):

\[
T_s = \frac{AT}{1+\beta_e} + \frac{\beta_e}{1+\beta_e} \left[ T_1 + \frac{L+S}{\lambda_l/(z_1/2)} \right] \tag{A2.1}
\]

where \( T_s \) is the surface boundary temperature (T), \( \beta_e \) is the dimensionless effective ground to air conductance ratio, \( \lambda_l \) is the thermal conductivity of the top layer (snowpack, forest floor, or soil, M L t\(^{-3}\) T\(^{-1}\)), \( z_1 \) is the thickness of the top layer (L), \( T_1 \) is the temperature midway through the upper layer (T), \( S \) is the net shortwave radiation (E t\(^{-1}\) L\(^{-2}\)), \( L \) is the net longwave radiation (E t\(^{-1}\) L\(^{-2}\)). Net shortwave radiation is found from the site characteristics (e.g., slope and aspect) and the declination angle (Yin and Arp 1993). The net longwave radiation is computed using a modified Stefan-Boltzmann
equation (Lee 1980). The $\beta_e$ parameter is a measure of the relative rate of heat transfer from the soil to the surface and the heat transfer from the air to the surface:

$$\beta_e = \frac{\lambda_i / (z_i / 2)}{h_e} \quad (A2.2)$$

The effective surface heat transfer coefficient $h_e$ (M t$^3$ T$^{-1}$) is a lumping parameter that includes the effect of the forest canopy (e.g., reduced radiation, altered evapotranspiration, decreased convective exchange) on the air-surface heat transfer rate. Yin and Arp (1993) developed an empirical relationship to estimate $\beta_e$ for a particular site:

$$\beta_e = 8.1L[1 - \exp \{\min(V_v, V_c) - 6.8\}] \quad (A2.3)$$

where $V_v$ and $V_c$ are the vegetation surface area index (m$^2$ m$^{-2}$) of the forest in question and the effective maximum vegetation surface area index (6.5 m$^2$ m$^{-2}$), respectively.

Heat fluxes through each model layer below the forest canopy (i.e., snowpack, forest floor, soil, and subsoil) are assumed to occur via conduction according to a modified version of Fourier’s one-dimensional transient heat conduction law that includes the effect of thermal sources and sinks due to phase change (Yin and Arp 1993):

$$\frac{\partial T}{\partial t} = \frac{1}{C} \left( \frac{\lambda}{C} \frac{\partial T}{\partial z} \right) + \frac{s}{C} \quad (A2.4)$$

where $C$ is the volumetric heat capacity of the medium (M L$^{-1}$ t$^2$ T$^{-1}$), and $s$ is the internal heat contribution due to phase change (M L$^{-1}$ t$^3$) An implicit finite difference scheme is
employed to obtain an approximate solution to the governing conduction equation. Further descriptions of the thermal model mechanics can be found in Yin and Arp (1993).

The hydrologic processes simulated by ForHyM2 are also indicated in Figure A2.1. Precipitation input data are first partitioned into rain and snow at a predetermined temperature threshold; for the present study, this was assumed to be 0°C. Precipitation entering the forest canopy is separated into throughfall (rain or snow), stemflow, and interception. Snow and rain interception are functions of the leaf area index which is taken as the weighted average of the coniferous (8 m$^2$ m$^{-2}$) and deciduous (5.5 m$^2$ m$^{-2}$ at full-leaf season) fractions. Under a partial or full deciduous canopy, the LAI varies seasonally; leaf shedding is a function of cumulative-degree days. During the winter, the snowpack accumulates through snow throughfall until it is reduced via melting or evaporation. Snow throughfall ($TF$, M L$^{-2}$ t$^{-1}$) and snow stemflow ($SF$, M L$^{-2}$ t$^{-1}$) are related to the canopy water ($W_c$) by proportionality constants $A_{TF}$ and $A_{SF}$ (Arp and Yin 1992):

$$TF = A_{TF} \times W_c$$  \hspace{1cm} (A2.5)

$$SF = A_{SF} \times W_c$$  \hspace{1cm} (A2.6)
The snowmelt process in ForHyM2 is described by Meng et al. (1995). Snowmelt $SM \ (\text{M} \ \text{L}^{-2} \ \text{t}^{-1})$ is computed by first summing the shortwave radiation at the snow surface $S_s$, longwave radiation at the snow surface $L_s$, convective sensible heat exchange between the snow and the atmosphere $H_c \ (\text{M} \ \text{t}^{-3})$, rain-induced energy input to the snowpack $H_r \ (\text{M} \ \text{t}^{-3})$, and the negative of the heat transfer from the snow surface to the snowpack $H_s \ (\text{M} \ \text{t}^{-3})$ and then dividing through by the heat of fusion $h_f \ (\text{L}^2 \ \text{t}^{-2})$:

$$SM = \left( S_s + L_s + H_c + H_r - H_s \right) / h_f \quad (A2.7)$$
Formulae are given by Meng et al. (1993) for computing the net shortwave and longwave radiation at the snow surface. The convective heat exchange is calculated as follows:

$$H_c = h[1 - \exp\{\min(V_c, V_v) - 6.8\}] \times [AT - ST]$$  \hspace{1cm} (A2.8)

where \( h \) is the heat transfer coefficient (M t^{-3} T^{-1}), \( ST \) is the snow surface temperature, and all other terms have been previously defined. Thus, SM is calculated based on the daily AT input to ForHyM2. The heat conduction to the snowpack, \( H_s \) is computed as the thermal conductivity of the snowpack times the difference between \( ST \) and the temperature of the snow/forest floor interface. The rain-induced energy input is assumed to be proportional to the difference between the mean monthly AT and mean monthly snowpack temperature. Further details for the snowmelt equations and related applications are given by Meng et al. (1993), Balland (2002), and Jutras et al. (2011).

Snowmelt, stemflow, and rain throughfall enter the forest floor layer until the saturation reaches field capacity, at which point percolation occurs to the soil layer. The flow of water between layers is always taken as proportional to the available moisture in that layer. This process continues to the subsoil. Water is also lost from the snowpack, forest floor, and soil layers through evapotranspiration (Arp and Yin 1992). Potential evapotranspiration (PET) is calculated according to the modified Hamon equation (Federer and Nash 1978). Actual evapotranspiration is equal to the lesser of PET or the soil water available (soil water content minus the permanent wilting point times a proportionality constant). Further descriptions, equations and applications of the hydrological component of ForHyM2 can be found in Arp and Yin (1992).
ForHyM2 Inputs

ForHyM2 requires the input of several parameters to describe the climate and site conditions. Table A2.1 details the values and justification for each parameter. ForHyM2 also has default values for soil and subsoil mineral composition and thermal properties. These subsurface values were not altered as the intent of the present study was to utilize ForHyM2 to simulate the impact of atmospheric processes on surface processes. Although surface and subsurface processes are interrelated, surface conditions (e.g., GST) are relatively robust in ForHyM2 compared to subsurface conditions (soil temperature and moisture content). Because of the relative robustness of GST, no calibrations of input parameters were required.

Table A2.1. Input parameters for the ForHyM2 simulations

<table>
<thead>
<tr>
<th>Input Parameters</th>
<th>Value</th>
<th>Justification and/or Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Latitude</td>
<td>46.8670°</td>
<td>-</td>
</tr>
<tr>
<td>Altitude</td>
<td>50 m</td>
<td>New Brunswick DEM (NBADW 2013)</td>
</tr>
<tr>
<td>Aspect</td>
<td>0</td>
<td>The watershed is quite flat (NBADW 2013)</td>
</tr>
<tr>
<td>Slope</td>
<td>1°</td>
<td>See above</td>
</tr>
<tr>
<td>Coniferous fraction</td>
<td>0.65</td>
<td>(Alexander 2006)</td>
</tr>
<tr>
<td>Deciduous fraction</td>
<td>0.35</td>
<td>(Alexander 2006)</td>
</tr>
<tr>
<td>Root depth index</td>
<td>2</td>
<td>This is the model value for medium shallow species.</td>
</tr>
<tr>
<td>Mean snow depth</td>
<td>13 cm</td>
<td>(EC 2013)</td>
</tr>
<tr>
<td>Climate factor</td>
<td>0.5</td>
<td>1 = continental, 0 = maritime (Alexander 2006)</td>
</tr>
<tr>
<td>Distance to coast</td>
<td>200 km</td>
<td>Number of kilometres to the reference coast (Bay of Fundy)</td>
</tr>
</tbody>
</table>
References


Appendix 3: SUTRA
Mechanics and Parameterization
SUTRA Governing Equations and Parameterization

The previous version of SUTRA (Voss and Provost 2002) was enhanced to include the dynamic freeze-thaw process by McKenzie et al. (2007). Unsaturated-zone freeze-thaw capability in SUTRA was thereafter developed by C.I. Voss, J.M. McKenzie and A. Provost as part of a U.S. Geological Survey research project focusing on cold-regions hydrology.

The governing groundwater flow equation in this version of SUTRA is:

\[
S_L \rho_L S_{op} + \varepsilon \left( \rho_L \frac{\partial S_L}{\partial p} \right) \frac{\partial p}{\partial t} + \varepsilon \left( S_L \frac{\partial \rho_L}{\partial T} \right) \frac{\partial T}{\partial t} - \nabla \cdot \left( \frac{k_{eff} \rho_L}{\mu} \right) (\nabla p - \rho_L g) = Q_p \quad (A3.1)
\]

where \( S_{op} \) is specific pressure storativity [L \(^2\) M\(^{-1}\)], \( \varepsilon \) is the soil porosity, \( p \) is the pore water pressure [M L\(^{-1}\) t\(^{-2}\)], \( t \) is time, \( T \) is temperature, \( k \) is the solid matrix permeability tensor [L\(^2\)], \( k_r \) is the relative permeability, \( \mu \) is the fluid viscosity [M L\(^{-1}\) S\(^{-1}\)], \( g \) is the gravitational acceleration vector [L t\(^{-2}\)], and \( Q_p \) is a fluid mass source [M L\(^{-3}\) t\(^{-1}\)]. The \( \rho_L \) and \( S_L \) terms represent the liquid water density [M L\(^{-3}\)] and the liquid water saturation (volume liquid water/pore volume), respectively.

The governing energy transport equation in this version of SUTRA is:

\[
C_{eff} \frac{\partial T}{\partial t} = -\varepsilon S_L \rho_L c_L \mathbf{v} \cdot \nabla T + \nabla \cdot \left( \frac{\lambda_{eff} \nabla T}{\rho_L} \right) + Q_p c_L (T^* - T) + \varepsilon S_L \rho_L \gamma_L + (1 - \varepsilon) \rho_s \gamma_S \quad (A3.2)
\]

where \( C_{eff} \) is the effective volumetric heat capacity of the ice-water-soil matrix [E L\(^{-3}\) °C\(^{-1}\)], \( c_L \) is the specific heat of liquid water [E M\(^{-1}\) °C\(^{-1}\)], \( \mathbf{v} \) is the average groundwater velocity vector [L t\(^{-1}\)], \( \lambda_{eff} \) is the effective thermal conductivity tensor of the
ice-water-soil matrix \([E\, t^{-1} \, L^{-1} \, °C^{-1}]\), \(T^*\) is the temperature of a fluid source, and \(\gamma_L\) and \(\gamma_S\) are the energy sources in the water and solid grains, respectively \([EM^{-1} \, t^{-1}]\).

The effective heat capacity is equal to the weighted arithmetic average of the heat capacities of the matrix constituents (i.e., liquid water, ice and solid grains) and includes latent heat effects during phase change:

\[
C_{\text{eff}} = \varepsilon (S_L \rho_L c_L + S_I \rho_I c_I) + (1 - \varepsilon) \rho_S c_S - \Delta H_f \varepsilon \rho_l \frac{\partial S_I}{\partial T}
\]  

(A3.3)

where \(\Delta H_f\) is the latent heat of fusion \([EM^{-1}]\), \(S_I\) is the ice saturation, \(\rho_I\) is the ice density \([ML^{-3}]\), and \(c_L, c_I, c_S\) are the specific heats of liquid water, ice, and the soil grains, respectively \([EM^{-1} \, °C^{-1}]\).

The effective thermal conductivity of the matrix is computed as the weighted average thermal conductivity of the matrix constituents plus the thermal effects from mechanical dispersion:

\[
\lambda_{\text{eff}} = \frac{\varepsilon (S_L \lambda_L + S_I \lambda_I) + (1 - \varepsilon) \lambda_S}{\varepsilon S_L \rho_L c_L + (1 - \varepsilon) \rho_S c_S} \mathbb{I} + \varepsilon S_L \rho_L c_L D
\]  

(A3.4)

where \(\lambda_L, \lambda_I, \lambda_S\) are the thermal conductivities of liquid water, ice, and the soils grains, respectively \([E \, t^{-1} \, L^{-1} \, °C^{-1}]\), \(\mathbb{I}\) is the identity matrix, and \(D\) is the mechanical dispersion tensor \([L^2 \, t^{-1}]\).

Simple piecewise linear functions were specified for the soil freezing curves, soil water characteristic curve (drying), and relative permeability function to maximize computational efficiency. These relationships have been proposed by McKenzie et al. (2007) and others. The total volumetric water saturation (liquid + ice) \(S_w\) is assumed to be
a function of pressure only, and this relationship can be expressed using a piecewise linear function:

\[
S_w = f(p) = \begin{cases} 
1 & \text{if } p_c < p_{ent} \\
1 - (1 - S_{wres}) \left( \frac{p_c - p_{ent}}{p_{swres} - p_{ent}} \right) & \text{if } p_{ent} \leq p_c \leq p_{swres} \\
S_{wres} & \text{if } p_c > p_{swres}
\end{cases} 
\] (A3.5)

where \( p_c \) is the pore water suction \([\text{ML}^{-1}\text{t}^2]\), \( p_{ent} \) is the air-entry suction\([\text{ML}^{-1}\text{t}^2]\), \( S_{wres} \) is the residual saturation due to drying, and \( p_{swres} \) is the suction at residual saturation \([\text{ML}^{-1}\text{t}^2]\).

In SUTRA, the total water saturation is partitioned into the solid and liquid phases based on temperature. If the soil is fully saturated \( (S_w=1) \), the liquid water saturation during saturated freezing is obtained from the soil freezing curve, which is also approximated using a piecewise linear function (McKenzie et al. 2007):

\[
S_w = f(T) = \begin{cases} 
1 & \text{if } T > T_f \\
- mT + 1 & \text{if } T_f > T > T_{wresi} \\
S_{wresi} & \text{if } T < T_{wresi}
\end{cases} 
\] (A3.6)

where \( S_{L(sat)} \) is the liquid water saturation during saturated conditions, \( m \) is the slope of the liquid water-temperature relationships, \( T_f \) is the freezing temperature of bulk water \((0^\circ\text{C})\), \( S_{wresi} \) is the residual liquid water saturation during freezing \( (< S_{wres}) \), and \( T_{wresi} \) is the temperature at when liquid water saturations first decrease to \( S_{wresi} \) \((^\circ\text{C})\).

If the soil is initially unsaturated \( (S_w<1) \) due to pre-freezing capillary suction, the process for partitioning between the liquid and ice saturations requires intermediate calculations. In the present study, \( S_{L(sat)} \), was first obtained via Equation (A3.6) for a given sub-zero temperature. This represents the maximum potential liquid water
remaining during unsaturated freezing or thawing. The maximum potential ice saturation was computed as $S_{I(sat)} = 1 - S_{L(sat)}$. The actual ice saturation $S_I$ was then taken as the lesser of the soil water available for freezing ($S_w - S_{wresi}$) or $S_{I(sat)}$. Finally, the actual liquid water saturation $S_L$ was computed as $S_L = S_w - S_I$. This approach assumes that as much ice forms as possible for a given temperature. The converse assumption (the maximum possible liquid water saturation remains for a given temperature regardless of the total saturation) was also investigated. In this process, the actual liquid water saturation was computed as the lesser of $S_{L(sat)}$ or $S_w$. The ice saturation was then calculated to be the difference between the total and liquid water saturations: $S_I = S_w - S_L$. Simulations performed in SUTRA indicated that the choice of liquid-ice partitioning formulation had little impact on the thermal or hydraulic conditions at the model output. The results presented in the current formulation employed the former algorithm, which assumes a dominance of ice formation. More complex processes have been proposed which employ the Clapeyron equation for simulating cryogenic processes (e.g., Dall’Amico et al. 2011, Painter 2011), but these are not included in this version of the SUTRA code.

The decrease in permeability as a result of soil drying or freezing was also represented by a piece-wise linear function. The relative permeability was assumed to be a function only of $S_L$ and to decrease linearly with $S_L$ until the minimum relative permeability was achieved at $S_{wresi}$. The parameters for the soil freezing curve (Equation A3.6) were estimated from laboratory tests on freezing unsaturated sand conducted by Watanabe et al. (2011).
Details concerning the values chosen for the thermal and hydraulic properties of the aquifers, the soil drying curve, the soil freezing curve, and the relative permeability function are given in Table A3.1 (next page). This table also includes information pertaining to the finite element mesh density (variable), the time step size (constant), and the SUTRA solver controls.

References


Table A3.1. Values/descriptions for the modeling parameters in SUTRA

<table>
<thead>
<tr>
<th>Model parameter</th>
<th>Value/description</th>
<th>Notes/reference</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>A) Soil properties</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Horizontal $k_{sat}$ in aquifer</td>
<td>$3\times10^{-10}$ m$^2$</td>
<td>Sieve tests (Allard 2008) + Hazen’s method</td>
</tr>
<tr>
<td>Vertical $k_{sat}$ in aquifer</td>
<td>$3\times10^{-11}$ m$^2$</td>
<td>Anisotropy ratio of 10</td>
</tr>
<tr>
<td>$k_{sat}$ in aquitard/basal till</td>
<td>$3\times10^{-14}$ m$^2$</td>
<td>4 orders of magnitude less than the aquifer $k$</td>
</tr>
<tr>
<td>Longitudinal hydro. dispersivity</td>
<td>2 m</td>
<td>See footnote$^1$</td>
</tr>
<tr>
<td>Porosity</td>
<td>0.35</td>
<td>(Domenico and Schwartz 1990)</td>
</tr>
<tr>
<td>Grain thermal conductivity, $\lambda_s$</td>
<td>3.18 W m$^{-1}$°C$^{-1}$</td>
<td>Inferred from Bonan (2008) with Eq. (A3.4)</td>
</tr>
<tr>
<td>Grain specific heat, $c_s$</td>
<td>854 J kg$^{-1}$°C$^{-1}$</td>
<td>Inferred from Bonan (2008) with Eq. (A3.3)</td>
</tr>
<tr>
<td>Grain density, $\rho_s$</td>
<td>2600 kg m$^{-3}$</td>
<td>Default SUTRA value for sand</td>
</tr>
<tr>
<td><strong>B) Water and ice properties</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Water thermal conductivity, $\lambda_L$</td>
<td>0.6 W m$^{-1}$°C$^{-1}$</td>
<td>(Bonan 2008)</td>
</tr>
<tr>
<td>Water specific heat, $c_L$</td>
<td>4182 J kg$^{-1}$°C$^{-1}$</td>
<td>(Bonan 2008)</td>
</tr>
<tr>
<td>Water density (liquid), $\rho_L$</td>
<td>1000 kg m$^{-3}$</td>
<td>Linearly varied with temperature</td>
</tr>
<tr>
<td>Ice thermal conductivity, $\lambda_I$</td>
<td>2.14 W m$^{-1}$°C$^{-1}$</td>
<td>Default SUTRA value for ice</td>
</tr>
<tr>
<td>Ice specific heat, $c_I$</td>
<td>2108 J kg$^{-1}$°C$^{-1}$</td>
<td>Default SUTRA value for ice</td>
</tr>
<tr>
<td>Ice density, $\rho_I$</td>
<td>920 kg m$^{-3}$</td>
<td>Default SUTRA value for ice</td>
</tr>
<tr>
<td><strong>C) Drying, freezing, and relative permeability functions</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Soil water curve</td>
<td>Linear</td>
<td>Linear function employed for simplicity</td>
</tr>
<tr>
<td>Air entry suction, $P_{ent}$</td>
<td>10,000 Pa</td>
<td>Estimated from Watanabe et al. (2011)</td>
</tr>
<tr>
<td>Residual saturation, $S_{wres}$</td>
<td>0.01</td>
<td>Estimated from Watanabe et al. (2011)</td>
</tr>
<tr>
<td>Suction at $S_{wres}$, $P_{wres}$</td>
<td>100,000 Pa</td>
<td>Residual sat. at approximately 10 m of suction</td>
</tr>
<tr>
<td>Soil freezing curve</td>
<td>Linear</td>
<td>Linear function employed for simplicity</td>
</tr>
<tr>
<td>Residual liquid sat ($S_{wresi}$)</td>
<td>0.005</td>
<td>Estimated from Watanabe et al. (2011)</td>
</tr>
<tr>
<td>Temperature at $S_{wresi}$, $T_{wresi}$</td>
<td>$-3^\circ$C</td>
<td>Estimated from Watanabe et al. (2011)</td>
</tr>
<tr>
<td>Minimum relative permeability</td>
<td>0.005</td>
<td>$k_{rel}$ decreases linearly with $S_L$ until $S_L = S_{wresi}$</td>
</tr>
<tr>
<td><strong>D) SUTRA discretization and solver settings</strong></td>
<td></td>
<td></td>
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<tr>
<td>Finite element mesh type</td>
<td>Fishnet</td>
<td>User-controlled mesh in Argus ONE</td>
</tr>
<tr>
<td>Min finite element height</td>
<td>0.03 m</td>
<td>Dense mesh in unsaturated freezing zone</td>
</tr>
<tr>
<td>Max finite element height</td>
<td>1 m</td>
<td>Coarse mesh in ice-free, saturated zone</td>
</tr>
<tr>
<td>Constant element width</td>
<td>2 m</td>
<td>Unsaturated flow is predominantly vertical</td>
</tr>
<tr>
<td>Number of nodes</td>
<td>85,170</td>
<td>In Config. 1 mesh (62,300 nodes in Config. 2)</td>
</tr>
<tr>
<td>Constant time step size</td>
<td>0.2 day</td>
<td>Fine resolution due to nonlinear processes</td>
</tr>
<tr>
<td>Number of time steps</td>
<td>36,500</td>
<td>20 years × 365 days per year × 5 steps per day</td>
</tr>
<tr>
<td>Solver</td>
<td>Direct</td>
<td>Banded Gaussian Elimination</td>
</tr>
<tr>
<td>Nonlinearity iterations</td>
<td>50</td>
<td>Required due to nonlinearities in eq.’s</td>
</tr>
</tbody>
</table>

$^1$In accordance with the recommendations of Constantz et al. (2003) and Vandenbohede et al. (2009), the longitudinal dispersivity was chosen to be much smaller than the solute dispersivity value for the same scale (Fetter 1993). The transverse dispersivity was an order of magnitude smaller than the longitudinal.
Appendix 4: Further Analysis of SUTRA Output
Seasonal Aquifer Energy Balance

The SUTRA output were analysed to investigate the drivers of seasonal energy change in thin aquifers. Figure A4.1a presents the rate of energy change across the aquifer boundaries (i.e., boundaries of modeling domain) in Configuration 2. The results for Configuration 1 are similar but amplified due to the greater aquifer configuration length (i.e., horizontal dimension). The left vertical aquifer boundary for either aquifer configuration was perfectly thermally insulating, therefore no energy was exchanged across this boundary condition. The upper surface boundary condition was a combination of specified temperature and specified recharge. Figure A4.1a demonstrates that the energy flux at the surface boundary through the specified temperature nodes (conduction) dominates the energy flux at the surface via the specified recharge boundary condition (advection). The energy flux through the specified recharge boundary condition is greatest in the spring when recharge is greatest.

The energy flux through the specified GST boundary condition is positive during the warm period (DOY 85-250) and negative during the colder period (DOY 251-85). Energy is lost through the specified pressure boundary condition (advection through the aquifer discharge point). The discharge of energy to the stream (specified pressure boundary) comprises only a small amount of energy exchange in comparison with conductive energy transport across the ground surface. The rate of energy input via the specified basal energy flux (0.060 W m\(^{-2}\) = 36 W per unit aquifer width for Configuration 2) is insignificant in comparison to the rate of energy input via the other boundary conditions.
Figure A4.1. Simulated rate of energy change per unit aquifer width (a) across the boundary conditions (BC) (left legend) and (b) within the aquifer (right legend) averaged for each day of the year for Configuration 2 for the reference period (1981-2000, SUTRA run R.2, Table 5.2). The results presented are further smoothed by averaging over every 10 days. The ‘total through BC’s’ data were calculated by summing the energy fluxes through each boundary condition.

The aquifer energy changes can be described by energy fluxes across the aquifer boundaries (Figure A4.1.a) or by internal aquifer energy changes due to changes in soil temperature, soil water temperature, water mass, and latent heat. Figure A.4.1b indicates that the predominance of this energy is stored in and released from the soil grains of the aquifer. The majority of this energy change in the mineral grains occurs in the shallow subsurface where seasonal changes in temperature are highest. A significant secondary amount of energy is stored and released through the latent heat associated with pore water phase change. Figure A.4.1b demonstrates that latent heat is released from the aquifer from November to December (DOY 315-350) during the onset of soil freezing and absorbed later in the year as a result of soil thawing. Due to the presence of the snowpack, latent heat can be released during the winter months even when AT is
considerably lower than 0°C, as thawing can occur if the subsurface temperature is warming and exceeds the minimum freezing temperature ($T_{\text{wrest}}$, Equation A3.6, Appendix 3). For the present simulations, soil freezing and thawing occurred over a temperature range from 0°C to -3°C. Figure A4.1b also indicates that the changes in internal aquifer energy due to variations in water temperature or water mass are relatively insignificant. Energy changes due to variability in water temperature are low because the water content is lowest in the shallow zone that experiences greatest temperature change.

**Thermal Peclet Number Analysis**

Figure A4.1 depicts the rate of energy change across the aquifer boundaries and within the aquifer domain; however, the heat transfer mechanisms for transporting the surface thermal signals to the discharge location are also of interest and may differ from the heat transfer processes across the boundaries. The relative roles of conduction and advection in controlling the seasonal thermal regimes of the two aquifers can be investigated with the thermal Peclet number ($P_e$), which is the ratio of advection to conduction. A thermal Peclet number can be obtained for vertical heat transport in aquifers experiencing seasonal harmonic surface temperature variations and with uniform flow and thermal properties (e.g., Equation 13, Goto et al. 2005). In this approach, the arbitrary characteristic length usually employed in the Peclet number is replaced by the specific penetration depth, which is the depth at which the amplitude of the annual groundwater temperature cycle has decayed to $e^{-1}$ of the amplitude of the temperature
cycle at the ground surface (Equation 7, Goto et al. 2005). Because heat transport above the groundwater table is predominantly vertical and because the GST function could be approximated by a phase-shifted sinusoidal function (see Figure 5.5), the Peclet form of Goto et al. (2005) can be utilised to obtain a first-order approximation of the relative contributions of vadose zone advection and conduction in the aquifers on a seasonal basis.

Table A.4.1 lists the reference period (1980-2000) vadose zone Peclet numbers computed for aquifer Configuration 1 at a horizontal distance halfway up the hillslope (SUTRA run R.1, Table 5.2). According to Goto et al. (2005), conduction and advection contributions are equivalent when this form of the Peclet number equals 1. To obtain seasonal Peclet values, we obtained the average liquid and ice saturations for the upper 2 m for each season. These saturations were then utilised to obtain the average seasonal heat capacity and thermal conductivity values. Although the subsurface thermal properties exhibited relatively little seasonal variability, the Peclet numbers are still characterized by strong seasonal variability due to the seasonality of recharge applied at the ground surface. The average vertical velocities presented in the table below were computed as the average recharge rate for each season divided by the soil porosity (0.35). Clearly, the effects of advection are negligible in the winter for this catchment. In contrast, the spring advective heat pulse induced by recharge from snowmelt begins to approach the conductive heat transport contributions ($P_e = 0.71$). On average, the advective heat transport for the vadose zone estimated from this simple approach is one third of the conductive heat transport and must therefore be accounted for. This is
particularly true when seasonal groundwater temperatures are simulated as the influence of advective heat transport exhibits temporal variability.

Table A4.1. Seasonal values for the shallow Peclet number for the reference period Configuration 1 simulation

<table>
<thead>
<tr>
<th>Season</th>
<th>Average vertical groundwater velocity ($\times 10^8$ m s$^{-1}$)</th>
<th>Thermal Peclet number$^1$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Winter</td>
<td>0.80</td>
<td>0.07</td>
</tr>
<tr>
<td>Spring</td>
<td>11.7</td>
<td>0.71</td>
</tr>
<tr>
<td>Summer</td>
<td>2.90</td>
<td>0.23</td>
</tr>
<tr>
<td>Fall</td>
<td>3.95</td>
<td>0.30</td>
</tr>
<tr>
<td>Average</td>
<td>4.84</td>
<td><strong>0.33</strong></td>
</tr>
</tbody>
</table>

$^1$The specific penetration depth was found to be 4.5 m from the mean annual velocity and thermal diffusivity. The seasonal $Pe$ values were then calculated as the product of the specific penetration depth and the seasonal thermal plume velocity divided by the seasonal thermal diffusivity (Equation 13, Goto et al. 2005).

The next two figures are referred to briefly in Chapter 5. Figure A4.2 shows the simulated changes in average daily groundwater discharge temperature during the summer months for each future climate scenario (SUTRA runs 1.1-7.2, Table 5.2) compared to the reference period simulations (SUTRA runs R.1 and R.2, Table 5.2). Figure A4.3 depicts the subsurface thermal impact of pore water phase change in the shallow subsurface halfway up the hillslope in aquifer Configuration 1 by comparing simulations with and without the freezing subroutine turned on in SUTRA (runs R.1 and 9.1, Table 5.2).
Figure A4.2. SUTRA-simulated changes in summer (June 1-August 31) groundwater discharge temperature (daily) for each climate scenario (SUTRA runs 1.1-7.2, Table 5.2) compared to the reference period simulations (SUTRA runs, R.1 and R.2, Table 5.2) for (a) Configuration 1 and (b) Configuration 2 averaged for each day for the last ten years of simulation (2056-2065).
Figure A4.3. SUTRA-simulated minimum daily groundwater temperature (1981-2000) for Configuration 1 at depths of 1 m, 2 m, and 4 m, below the ground surface for each day of the year. Results are presented for when the thermal effects of pore ice were considered (SUTRA run R.1, Table 5.2) and when they were ignored (SUTRA run 9.1, Table 5.2). The soil temperature profiles were taken at a horizontal location halfway (i.e., 500 m) between the two vertical boundaries shown for Configuration 1 in Figure 5.4a.

References

Curriculum Vitae

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**Publications in Refereed Conference Proceedings**

Oral Conference Presentations (abstracts refereed)


**Poster Conference Presentations**

**Kurylyk, B.L.** (primary author), K.T.B. MacQuarrie (presenter), and C.I. Voss. 2013. Simulated impacts of future climate scenarios on the magnitude and temperature of groundwater discharge from shallow aquifers. Presented 17 September 2013 at the 40th Congress of the International Association of Hydrogeologists in Perth, Australia.

**Kurylyk, B.L.** (primary author and presenter) and K.T.B. MacQuarrie. 2012. Modeling the impact of climate change on groundwater-sourced thermal refugia. Presented 15 November 2012 at the Atlantic Climate Adaptations Solutions Conference ‘Preparing for Climate 2100- Tools and Strategies for NB Communities’ in Fredericton, New Brunswick.

**Kurylyk, B.L.** (primary author and presenter) and K.T.B. MacQuarrie. 2012. Modeling the impact of climate change on groundwater-sourced thermal refugia. Presented 27 April 2012 at the 2012 University of New Brunswick Graduate Research Conference in Fredericton, New Brunswick.

**Kurylyk, B.L.** (primary author and presenter) and K.T.B. MacQuarrie. 2012. Modeling the impact of climate change on groundwater-sourced thermal refugia. Presented 3 February 2012 at the 2012 Canadian Water Resources Association Mid-Term Meeting in Moncton, New Brunswick.

**Kurylyk, B.L.** (primary author and presenter) and K.T.B. MacQuarrie. 2011. Modeling the impact of climate change on groundwater-sourced thermal refugia. Presented 8 July 2011 at the 2011 Canadian Rivers Institute Day at the University of New Brunswick in Fredericton, New Brunswick.

**Invited Presentations**

**Kurylyk, B.L.** 2013. The influence of climate change on habitat for cold-water fishes in New Brunswick, Presented 19 September 2013 to the McGill University Civil Engineering Student Colloquium in Fredericton, New Brunswick.

Non-Refereed Professional Presentations


Kurylyk, B.L. (primary author) and K.T.B. MacQuarrie. 2011. Project 3b year 3 update: The influence of climate change on groundwater-sourced thermal refugia. Presented 21 November 2011 to the NSERC Collaborative Research and Development Grant Team Meeting at the Institut National de la Recherche Scientifique - Centre Eau Terre Environnement in Québec City, QC.


Kurylyk, B.L. (primary author and presenter) and K.T.B. MacQuarrie. 2010. Project 3b year 2 update: The influence of climate change on groundwater-sourced thermal refugia. Presented 7 December 2010 to the NSERC Collaborative Research and Development Grant Team Meeting at the INRS - Centre Eau Terre Environnement in Québec City, QC.

Kurylyk, B.L. (primary author and presenter) and K.T.B. MacQuarrie. 2010. Project 3b year 1 update: Modelling the effect of climate on thermal refugia generated by groundwater-surface water interactions. Presented 12 February 2010 to the NSERC Collaborative Research and Development Grant Team Meeting at INRS - Centre Eau Terre Environnement in Québec City, QC.

Non-Refereed Research, Technical Reports and Bulletins

Kurylyk, B.L. 2013. Soil quality guidelines for the protection of aquatic life in accordance with the Domencio (1987) contaminant transport model and the CCME. Spreadsheet-based computer program developed for GEMTEC Ltd.
