# Climate Change Impacts on Snowmelt-Driven Streamflow in the Grand River Watershed: Implications for Water Resource Management

by

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## Author's Declaration

I hereby declare that I am the sole author of this thesis. This is a true copy of the thesis, including any required final revisions, as accepted by my examiners.

I understand that my thesis may be made electronically available to the public.

## Abstract

Climate change is one of the most significant global environmental drivers threatening the quality and quantity of future water resources. Global temperature increases will have significant effects on the hydrologic regime of northern regions due to changes in snowfall and snowmelt. Considerable research has been conducted in western Canada to rigorously quantify snowmelt-driven streamflow processes, however, less focus has been directed towards understanding these processes in eastern Canada and Ontario. In the southern Ontario Grand River Watershed (GRW), snowmelt contributions to streamflow (freshet) make up a significant portion of the annual water yield, and the period of snowmelt is also of key concern for flood mitigation. This thesis aims to quantify historical and projected changes to timing and streamflow during freshet in the Nith River, an unregulated tributary of the Grand River. Climate data (temperature, rainfall, snowfall, and snow proportion) from observations and future scenarios were analyzed to quantify the contributions of climate conditions surrounding the timing and volume of the freshet. The annual timing of snowmelt-driven streamflow was quantified using centre time (CT), and streamflow volumes were quantified by various percentiles of streamflow ($Q_n$) during four periods of the water year (October-December, January-February, March-April, and May-September). Historical trends in streamflow and climate data were examined using hydrometric data (1914-2016) of a stream gauge from the Water Survey of Canada, and climate data (1950-2016) from Environment and Climate Change Canada at two stations. Projected climate data were from an ensemble of models used in the Intergovernmental Panel on Climate Change's Fourth Assessment Report (AR4). A total of nine distinct models ran two scenarios from AR4 for the 2050s; moderate (B1) and high (A1B). These time-slice projections were then used to force the hydrologic model GAWSER to simulate future streamflow data. The results show that CT in the Nith River has advanced by 17 days, on average, from 1914 to 2016 ($P=0.036$), and the advance is projected to continue as a function of future emissions scenario (approximately 12 days for scenario B1, and 17 days for A1B). Historical CT was weakly negatively correlated with temperature (-0.51, $P < 0.001$), where colder winters were associated with a later CT. Results from a multiple regression model using climate variables to predict CT were inconclusive. Historical streamflow at the $Q_{10}$ level has increased from 1914-2016 ($P < 0.001$), but no significant change has been observed at the $Q_{50}$, and $Q_{90}$ levels. Future freshet streamflow is projected to increase for both scenarios at the $Q_{10}$ level (an average of 18.1% and 23.6% respectively) as well as the $Q_{50}$ level (an average of 20.8% and 26.6% respectively). No change was observed in $Q_{90}$ for either future scenario. The results of this thesis will inform water resource managers of climate change impacts to average hydrologic conditions in the GRW.

# Acknowledgements

I would like to thank my supervisor Chris Fletcher for his invaluable guidance throughout this degree and thesis project. Thank you also to Mike Stone for his help throughout this thesis, as well as his support throughout my time at UWaterloo. I would also like to thank my committee member Linda Mortsch for taking the time to review this document. Thanks also to Stephanie Shifflett who provided data and support for this project.

# Dedication

This thesis is dedicated to my family, in particular to my parents. Thank you for all of your support.

I would also like to acknowledge and thank the friends that I made at UWaterloo for making this experience a great one. The Winchester crew, the 2016 GEMs, my CWP cohort, and my lab mates, these last few years wouldn't have been the same without all of you.

Lastly, to Amanda and Gilmour; thanks.

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# 1. Introduction

## 1.1 Problem Statement

Climate change is one of the most significant global environmental threats to the quality and quantity of future water resources (Delpla et al. 2009; Nijssen et al. 2001; Rodell et al. 2018). The seasonal distribution and availability of surface water is expected to change regionally due to climate change pressures, which can challenge water resource managers to make informed decisions regarding water planning and use (Rodell et al. 2018). These challenges are particularly acute in rapidly growing urban areas where concerns about water quantity and quality are increasing. Regional changes in variables like surface air temperature and precipitation patterns are closely related to changes in hydrologic processes, but knowledge of how these variables change, such as variability in snowfall patterns and the timing of melt and its effect on streamflow, are poorly understood for many regions (Clark and Slater 2016; Sillmann et al. 2017). These knowledge gaps must be addressed to inform management and policy in order to adapt to the negative effects of climate change on streamflow.

Changes to the spatial and temporal distribution of snow cover and related hydrologic processes associated with climate change remain a key concern for water resources management in Canada (Jones et al. 2015). Although global surface air temperature is expected to continue to increase during the 21st century, these temperature changes are not evenly distributed (Ljungqvist et al. 2016). Greater temperature increases have been observed over cold regions and mid and high latitudes when compared to global trends (Ljungqvist et al. 2016). Accordingly, these temperature increases will have significant effects on the hydrologic regime of northern regions because of changes in snowfall and snowmelt (Allchin and Déry 2017; Brown and Mote 2009; Danco et al. 2016).

Considerable research has been conducted in western Canada to rigorously quantify snowmelt-driven streamflow processes (Déry et al. 2009; Fritze et al. 2011). Results of these studies show that earlier freshet and overall smaller water yields can be expected due to a declining snowpack (Déry et al. 2009; Fritze et al. 2011; Jones et al. 2015). However, less focus has been directed towards understanding these processes in eastern Canada and Ontario, where changes to water availability in highly populated areas present a significant challenge for water supply. Winter and spring runoff in many southern Ontario snow-dominated watersheds is strongly influenced by snowmelt dynamics (Liu et al. 2016). In the Grand River watershed a large portion of the annual water yield occurs during the late winter and early spring, much of which is stored in reservoirs and subsequently released to augment streamflow in the summer (Etienne 2014). Accordingly, the snowpack is a crucial store of water that can increase the response time of streamflow from precipitation events and alter the transfer in surface and groundwater flows to rivers (Adam et al. 2009). In many regions globally, the hydrologic regime is dominated by the amount and timing of snowmelt. One of the most important impacts of climate change on streamflow in mid-latitude regions is related to changes in the spatial and temporal patterns of precipitation and temperature (Allchin and Déry 2019a). However, there is considerable uncertainty in our ability to model and predict the impact of these changes on streamflow in watersheds at a basin scale (Clark et al. 2016; Prudhomme and Davies 2009). In order to make informed planning decisions regarding water management, an improved knowledge base is required to better understand how streamflow might change in relation to changes in snowmelt dynamics. Because the issue of water supply depends on both the availability and quality of freshwater as well as the demand, projecting how management practices should change and planning for adequate infrastructure renewal relies on understanding how both the supply and demand of water resources will change. Population increase and urbanization will increase pressure on the Grand River, its tributaries, and groundwater to provide enough water for the growing municipalities, and climate change will increase uncertainties regarding water availability (Etienne 2014). This is complicated further by changes to flooding intensity, frequency, and duration, particularity during the spring freshet, which are also of key concern for the Grand River Watershed (GRW) as hydrologic processes are impacted by climate change.

Freshet is a critical time when snowmelt contributes to streamflow, and knowledge of freshet impacts on streamflow dynamics is necessary for planning because freshet flows constitute a significant portion of the annual water yield in the GRW. Currently, freshet derived streamflow is stored in reservoirs throughout in the GRW primarily to mitigate flooding (Etienne 2014). In the Grand River Conservation Authority (GRCA) water management plans climate and hydrologic uncertainty have been identified as a key concern for water managers (Etienne 2014). Accordingly, there is a critical need to rigorously quantify streamflow response to variable snowmelt in the context of a changing climate.

## 1.2 Changes to Snowmelt-Driven Runoff

Climate change poses a specific set of challenges and uncertainties in snow-dominated watersheds where a significant proportion of the annual water yield comes from melting snow and runoff during the shoulders of the cold season (Barnett et al. 2005). The spring freshet, which is defined by the annual additional streamflow pulse occurring due to snowmelt, is a critical process to understand as climate change continues to alter it (Jones et al. 2015). The spring freshet has been an important area of research in Canada because of its importance for water resource planning and management and it is used as a hydrologic indicator of climate change (Koshida et al. 2015). Freshet-period flows account for a large portion of the annual water yield of rivers in many snow-dominated watersheds and this is the time of year when the highest streamflow volumes occur (Adam et al. 2009). Changes to the annual mean hydrologic regime have implications for water resource managers as well as for flood mitigation.

In the GRW, several challenges surrounding changes to the hydrologic regime resulting from climate change, and specifically, changes to snowmelt-driven streamflow have been identified (Jyrkama and Sykes 2007; Li et al. 2016). The GRW has a large and concentrated population that relies on groundwater and streamflow to supply its municipalities and rural communities with water. In addition to municipal water use, which comprises approximately 60% of water demand in the watershed, other stakeholder groups include agriculture, industry, rural water supply, and recreation (Etienne 2014).

The GRCA manages water resources in the GRW through dam operations, as well as conservation and land-management initiatives. Seven large dams operated by the GRCA are used to augment flows and reduce flood damage (Etienne 2014). Dam operations are the most critical during the freshet period as a fine balance must be kept between storing enough water during the freshet when increased discharge is occurring to augment flows during the dry season, while still maintaining sufficient storage in the reservoirs to mitigate flood damage in the event of an extreme precipitation and/or melt event. The GRCA uses historical weather and streamflow data to make operational decisions regarding volume of water to store, and to release in order to have storage available for the incoming spring melt. However, this is more difficult to respond to as climate change alters the timing and volumes of streamflow resulting from snowmelt. This is further complicated by the fact that climate change has the potential to alter the window of variability of extreme precipitation and flooding events in the future, as the hydroclimatic system is nonstationary (Milly et al. 2008). Despite temporal trends being commonly used to inform water resources management decision-making, streamflow is widely accepted as a nonstationary process, particularly in the context of climate change, where changes to extremes and variability are potentially magnified and accelerated (Milly et al. 2008; Vogel et al. 2011).

The effects of changes to the freshet on the hydrologic regime can be quantified and analyzed on two timescales; 1) the mean streamflow conditions over time (Allchin and Déry 2019b; Fritze et al. 2011) and 2) the variability of flows and the intensity/frequency/duration individual extreme events (Arnell and Gosling 2016; Gizaw and Gan 2016). Much of our understanding of watershed-scale hydrology and its driving processes is based on climatological means, which have long-term temporal records to create an average annual hydrologic regime. Mean annual streamflow and meteorological data can be used to classify rivers and their annual regimes by identifying the average distribution and magnitudes of streamflow (Gleick 1986; Harris et al. 2000). Such changes to average streamflow conditions are very significant for water resource managers to respond to changes in flow variability in order to make operational decisions regarding seasonal flows (Anghileri et al. 2016). It is also important to quantify and measure changes to extreme events in order to prepare for and mitigate flood damages, although these are more challenging to track over time because they are often based on individual events and flood thresholds (Chen et al. 2013; Stevens 1992). In both cases, knowledge of changes to the freshet is key in order to mitigate risk for flooding and for reservoir management.

## 1.3 Processes Controlling Streamflow Variability

Climate variability is one of the main processes controlling the temporal and spatial variability of river flows (Barnett et al. 2005; Gleick 1986). In the GRW, the largest contributor to streamflow for most of the year is precipitation, and specifically rain, where the lag time of the streamflow response to runoff is short (Jyrkama and Sykes 2007; Li et al. 2016). During the cold season, temperature and precipitation both play a significant role in determining the timing and magnitude of runoff (Fan and He 2015; Sospedra-Alfonso and Merryfield 2017). The magnitude and frequency of precipitation events, and the number of antecedent dry days strongly influence the timing of runoff volume (Berghuijs et al. 2014). Other factors affecting the quantity and timing of runoff include soil type, slope vegetation, land use, and parent geology (Ayers and Ding 1967; Li et al. 2012). Regional characteristics such as basin physiography, and proximity to large bodies of water also play an important role in determining temperature and precipitation patters on a regional scale, which can strongly affect the hydrologic regime (Thomas and Benson 1970).

In the Great Lakes Basin, meteorological and hydrological processes related to climate change are impacted by the availability of water vapor due to changes in evaporative fluxes from the Great Lakes (McBean and Motiee 2008). The geology, glacial-fluvial deposits and soils of a region, as well as land-use, can influence the distribution of streamflow within a watershed due to differences in infiltration ability and storage capacity (Thomas 1966). The geological makeup within a watershed is responsible for aspects of the monthly distribution of streamflow, where the coarseness of soil determines the distribution of streamflow percentages vs. groundwater storage seasonally (Ayers and Ding 1967; Berghuijs et al. 2014). Ayers & Ding (1967) found that the dominant surficial materials in southern Ontario watersheds influenced the seasonal distribution of runoff, where the finer the dominant material in soil was associated with a higher percentage of the total annual runoff occurring in March-April, and lower summer flows. Watersheds with coarser soils typically have a greater amount of groundwater recharge, and as a result, higher summer flows, where watersheds with finer soils had more runoff occurring during the freshet and lower summer flows, rather than being stored as groundwater. The GRW was characterized as having medium-textured soils, and a significant portion of its annual discharge (approximately 58%) occurring in March-April (Ayers and Ding 1967). Land-use change is commonly responsible for changing the hydrologic regime of a region, although hydrologic alteration caused by land-use change can be difficult to separate out from climate changes and internal variability over time (Kalnay and Cai 2003; Tomer and Schilling 2009). Typically, the development of land from forest to agricultural, or from either of these classes to paved (urban and/or residential) create a shorter lag time in runoff to rivers, and can also intensify flooding, due to reducing infiltration capacity of land surface (Kalnay and Cai 2003).

## 1.4 Climate Change Impacts on the Freshet

Climate change has already had an impact on hydrologic processes in Canada, and annual and seasonal changes to temperature and precipitation patterns are projected to continue to alter the hydrologic regime of watersheds (Jones et al. 2015; Zhang et al. 2000). For example, an increased variance of precipitation globally has been observed, with dry areas becoming dryer and wet areas becoming wetter (Dore 2005). It has also been established that there is an emerging pattern of increasing precipitation in higher latitude regions, and that temperature increases are more impactful in higher latitudes due to their effects on snow processes (Dore 2005). Because of this increased warming, snowfall is projected to decrease in the Northern Hemisphere, despite the observed and projected increases to overall precipitation (Danco et al. 2016). There is evidence that in watersheds with nival regimes, climate change could cause a decrease in flood risk due to a decrease in the accumulation of a winter snowpack and a shift of the regime from being snow-dominated to rain-dominated (Hamlet and Lettenmaier 2007). However, much of this research has been conducted in mountain snowpack-dominated watersheds (Déry et al. 2009; Fritze et al. 2011; Hamlet and Lettenmaier 2007). In the GRW, which has a dense concentrated population and where the snowpack contributions to streamflow are evenly distributed throughout the watershed, the effects of climate change on seasonal streamflow and flood risk should be studied further. In the GRW, it is likely that climate change will cause an overall increase in groundwater recharge, due to the reduction of ground frost and the shift of snowmelt regimes from spring to winter (Jyrkama and Sykes 2007).

The main variables determining the timing and the streamflow volumes during the freshet period are related to two processes; the accumulation and ablation of the snowpack. These processes are directly affected by the amount and form of precipitation during the winter months, and the temperature conditions during the winter in order to sustain the snowpack as well as determining when snowmelt occurs. There has been extensive research in western North American mountainous watersheds on changes to the timing and volume of the spring freshet (Adam et al. 2009; Barnett et al. 2005; Fritze et al. 2011; Jones et al. 2015). The timing of snowmelt has been observed to occur earlier in the water year in many watersheds, shifting from the spring to occurring in the winter. These changes are projected to continue in these regions as winters warm (Burn 2008; Clow 2009; Jones et al. 2015). The shift in timing of freshet flows from spring to winter are strongly associated with temperature changes, where streamflow volumes during the freshet are more dependent on changes in precipitation and temperature (Clow 2009; Fan and He 2015; Jones et al. 2015; Pradhanang et al. 2013). The key changes that have been observed in the freshet are increasing winter temperatures driving an earlier snowmelt, where freshet streamflow volumes are increasing in some regions and decreasing in others due to a combination of temperature and precipitation changes. Trends in much of eastern Canada and Ontario have not been as conclusive as in western North America (Jones et al. 2015). In Ontario, the most recent projections show that along with an increase in temperature to varying degrees across scenarios and seasons, precipitation will also increase (Wang et al. 2015). Southern Ontario in particular is expected to get more precipitation in the autumn, winter, and spring, with more uncertainty associated with summer precipitation increases (Wang et al. 2015).

### 1.5 Sources of Uncertainty in Climate and Hydrologic Data

There are several sources of uncertainty in both analyzing historical trends and variability in data, and projecting future changes to the hydrologic regime of a watershed with models. The main sources of uncertainty within this analysis come from the issue of stationarity in observed data, climate model structure, and radiative forcing scenario uncertainty. First of all, the concept of is based on the assumption that the way that a variable changes over time is constant, or that the processes of change and internal variability themselves do not change (Milly et al. 2008). In this sense, when we make assumptions based on trends in climate or hydrologic data, we are assuming that the internal variability itself is not changing. Milly et al. (2008) argue that it is problematic to draw conclusions from observed hydrometric data and modeled hydrologic response to General Circulation Model (GCM) forcings when making engineering decisions, because climate and hydrologic change are inherently nonstationary. In observed data, the chosen sampling frame can be arbitrary in regards to trend testing, not only in the magnitude of trends but even in the direction of change. In observed records, there are many instances where the nonstationary nature of climate and hydrologic data is evident where variability changes as well as mean conditions (Chiew et al. 2014). In this thesis, hydrologic change is analyzed in both observed trends as well as modeled climate change scenarios in order to better understand processes that cause changes to the freshet. In addition, because of the nature of the way the modeled data were calculated using a time-sliced changefield method, there are no conclusions drawn regarding the potential changes to climate or hydrologic variability. Secondly, in drawing conclusions from climate change projection data, uncertainty from the chosen GCM must be addressed. GCMs vary in their structure, and have different responses to the same radiative forcing. In the field of hydrologic modeling climate change scenarios, GCM uncertainty is typically the largest source of uncertainty when compared to uncertainty from scenario or hydrologic modelling (Prudhomme and Davies 2009; Kay et al. 2009a). These differences become even more apparent when GCM data is downscaled to a level at which a hydrologic model can be forced, where the variability between models' precipitation is larger than temperature. In this study, this will be addressed by using an ensemble of GCMs to simulate multiple scenarios in order to calculate a 95% confidence interval around the average projected changes to different variables. Finally, because the actual climate change scenario that will unfold is determined by many complex and difficult to model factors, such as global political and socioeconomic conditions, technological advancements, population growth determining greenhouse gas emissions, a range of scenarios must be considered (Solomon et al. 2007). The Intergovernmental Panel of Climate Change (IPCC) created potential scenarios that estimate emissions and the resulting change in radiative forcing for the Special Report on Emissions Scenarios (SRES) in order to model the climate response of various increases in radiative forcing (Nakicenovic and Swart 2000). In this project, two scenarios are examined, B1 with a global mean temperature increase of between 1.1°C to 2.9°C, and A1B, with a mean temperature increase of between 1.7°C and 4.4°C by 2090-2099 relative to 1980-1999. By assessing both the range of uncertainty related to differences in individual GCMs as well as different scenarios across models, we can better understand the main sources of uncertainty in modeling hydrologic response to future change.

## 1.6 Research Objectives

The goal of this research is to examine the impacts of climate change on spring freshet and snowmelt-driven runoff on streamflow in the Grand River Watershed. Long-term climate and streamflow data are used to identify the temporal changes in snowmelt-driven runoff. These data are used as input parameters in an ensemble of 9 Coupled Model Intercomparison Project phase 3 (CMIP3) models from the IPCC's Fourth Assessment Report (AR4) which are then used to force the Guelph All-Weather Sequential Events Runoff (GAWSER) model to provide estimates of potential future changes to these variables. An analysis of climate variability will be used to explain long term temporal variation in the hydrologic response of both timing and volume of streamflow during spring freshet in the Nith River. Such information can be used to consider how climate change may influence the timing and volume of streamflow during the spring freshet and develop strategies to adapt to these climate induced changes in stream flow. Specifically, the research objectives of this thesis are to;

1. Identify and quantify changes to the timing and volume of streamflow during spring freshet in the Nith River for the period 1914-2016.
2. Examine whether a climate change signal can be detected in temporal changes in the historical freshet streamflow data.
3. Quantify the projected changes to the future timing and volumes of streamflow during the freshet under two different radiative forcing scenarios using a GCM ensemble.
4. Identify main sources of uncertainty in projecting changes to hydrologic variables as a result of climate change.

# 2. Methods

## 2.1 Study Area

The Grand River drains an area of 6,800 km² and is the largest southern Ontario tributary to Lake Erie (Figure 2.1). The population of the watershed is > 1 million people, and contains several urban centres such as Kitchener-Waterloo, Guelph, and Cambridge. The Grand River and its tributaries supply a significant amount of the water to municipalities and communities including Guelph, Waterloo Region, Brantford, and Six Nations Territory within the watershed, along with groundwater (Etienne 2014). Several large dams and reservoirs were constructed between 1942 and 1976 with the goal of maintaining flows for municipal water supply and water quality in the dry season, as well as mitigating flooding damages (Etienne 2014).

The present study was conducted in the Nith River sub-watershed, which is located in the western region of the Grand River watershed. This basin drains an area of 1,030 km². Land land-use in the Nirth River watershed is mainly agricultural with some small municipalities, and its surficial geology is mainly clay-till in the western regions and silty/clayey in the eastern regions (Shifflett 2014; Presant and Wicklund 1971). The river flows from northwest to southeast in the subbasin, mainly along the clay till regions. The Nith River was selected for study in this thesis because it is the largest unregulated tributary of the Grand River and it has a long historical record of observed daily streamflow at one of its gauging stations. Although the Grand River has some stream gauges with long observational records, due to damming and flow augmentation throughout the watershed, streamflow records do not necessarily reflect the true hydrologic response to precipitation and temperature changes. Land-use in the upper and middle reaches of the sub-basin is predominantly agricultural and has remained fairly steady since the beginning of the historical streamflow record.

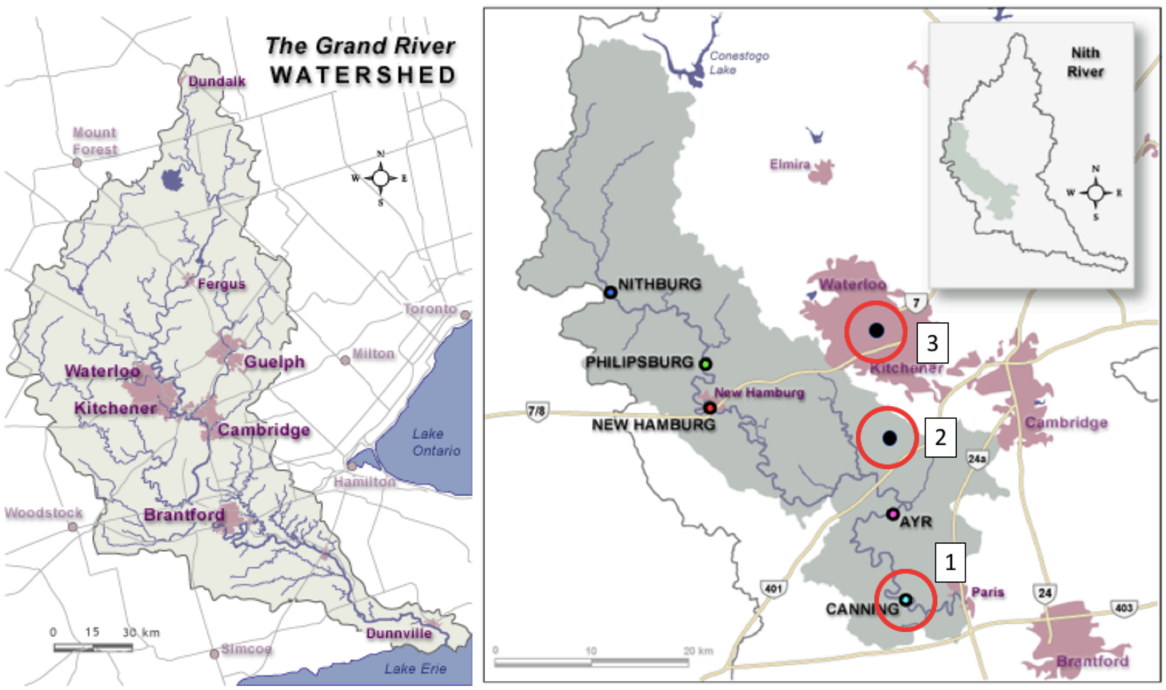


*Figure 2.1: Location of the Grand River Watershed, and the Nith River Subbasin within. Locations of the stream gauge used ("CANNING", 1), as well as "ROSEVILLE" weather station (2), and "KITCHENER" weather station (3). Source: Grand River Conservation Authority, 2006.*

## 2.2 Historical Hydrometric and Weather Data

### 2.2.1 Streamflow Gauge & Weather Stations

Historical streamflow data from the Water Survey of Canada Nith River near Canning gauging station (02GA010) (43°11'23" N, 80°27'18" W) for the period 1914 to 2016 were used in this study in order to detect trends in hydrometric data and analyze the climate conditions associated with the timing of and streamflow during freshet. For the period 1914 to 2016, no daily streamflow data were available for the period 1923 to 1948. The data from before the gap in observations were kept for data completeness, as they span almost 10 years and the gauge location was not changed during the record. In addition, historical data from two Environment and Climate Change Canada weather stations were because of their proximity to the streamflow gauge and period of record. The Roseville weather station is located, approximately 21km from the Canning stream gauge (43°21'13.026" N and 80°28'25.056" W) and the data spans a period from 1972-2016 (Figure 2.1, [2]). The Kitchener weather station is located, approximately 30km from the Canning at (43°26'00.000" N and 80°30'00.000" W) and has data for the period 1950 to 1978 (Figure 2.1, [3]). Data at both stations include daily maximum temperature, rainfall, and snowfall data for the period 1972 to 1978. These two stations were selected in order to create a record of adequate length for trend analysis. Data from the stations were overlapping from 1972-1978 and were examined for any bias in one station against the other (plots of these bias checks can be found in the Appendix, Figure 6.1). Scatterplots and correlations showed that maximum daily temperature between stations was very closely correlated (0.99), with no bias in the record of either station. Precipitation was slightly less closely correlated between stations due to the more spatially variable nature of precipitation, where rain had a correlation coefficient of 0.87, and snowfall was 0.92.

### 2.2.2. Analysis of Hydrologic Data

Average streamflow conditions in the Nith River basin vary seasonally due to differences in precipitation and temperature and most notably due to snowpack accumulation and ablation. Because of the influence of snow accumulation and snowmelt during the cold season on streamflow, the hydrologic regime of the Nith river watershed has been divided into 4 distinct time periods for analysis in this study. Table 2.1 outlines these phases of the average annual flow regime from 1961-1990, presented in Figure 2.2.

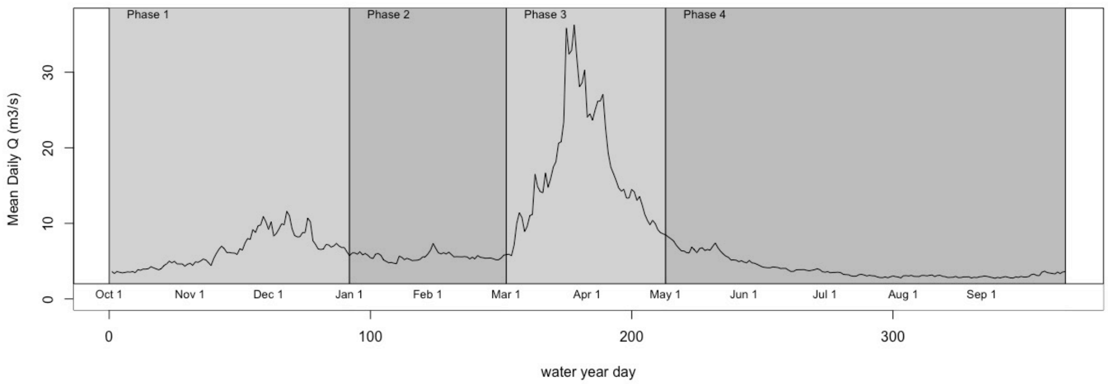


*Figure 2.2: Average hydrologic regime in the Nith River and 4 defined phases of streamflow for the period of record 1961-1990.*

**Table 2.1: Phases of the water year**

| Time period | Water year day | Range of Dates |
|---|---|---|
| P1 – Fall | 1-92 | OCT 1 – DEC 31 |
| P2 – Accumulation | 93-152 | JAN 1 – FEB 28 |
| P3 – Freshet | 153-213 | MAR 1 – APR 30 |
| P4 – Dry season | 214-365 | MAY 1 – SEP 30 |

Period 1 (P1, fall) ranges from October 1, to the end of December. This period is characterized by flows fed by rain events as well as the beginning of winter snowfall and some melt events. Period 2 (P2, “the accumulation phase”) spans through January and February, and typically has decreased streamflow from P1 due to colder temperatures and a shift in precipitation from rain-dominated to snow-dominated. Period 3 (P3, “the freshet period”) goes

from March 1-April 30, and is the period of the annual regime which typically has the largest daily streamflow volumes of the year, mainly due to snowmelt. Low temperatures and a high volume of snow during P2 result in a lower volume of streamflow in P2, and a larger freshet pulse in P3. This pulse lasts several weeks, before the river shifts into Period 4 (P4, "the dry season") which goes from May 1 to September 30 and is characterized by low summer flows, where the river no longer receives contributions from snowmelt. Some larger flows and floods that are caused by rain events occur during this time, but this is typically a low-flow time of year. Water is typically stored in various reservoirs around the Grand River watershed during high P3 flows, then used to augment streamflow in P4.

### 2.3 CMIP3 ensemble and radiative forcing scenarios

Climate change scenarios used in this project were originally created for the Grand River Conservation Authority's (GRCA) 2014 Water Management Report, which aimed to help in the planning and future management of water resources in the basin (Shifflett 2014). In total, there were 9 distinct GCM simulation outputs (Table 2.3.1 along with the emissions scenario they were forced by) that were downscaled using the change-field method where changes to temperature and precipitation are calculated by taking monthly relative changes from GCMs from 2041-2070 (2050s) and applied to observed data from the baseline time period (1961-1990) (Shifflett 2014). GCMs were forced using the B1, A1B, and A2 emission scenarios to assess the sensitivity of the climate system to changing concentrations of greenhouse gases. B1 is the scenario used with the lowest amount of total surface warming, where average lobal surface temperatures increase approximately 1°C warming by the year 2050 (Figure 2.3). A1B is the more aggressive scenario, with approximately 2°C of warming occurring by the year

2050. A2 is the scenario with the highest amount of warming. However, in this scenario global surface warming does not differ from scenario A1B by the 2050s, as more of this warming occurs by the 2100s. Only 1 GCM was forced with scenario A2, so it was excluded from scenario uncertainty analysis using multi-model means, but the individual run was used when analyzing model uncertainty.

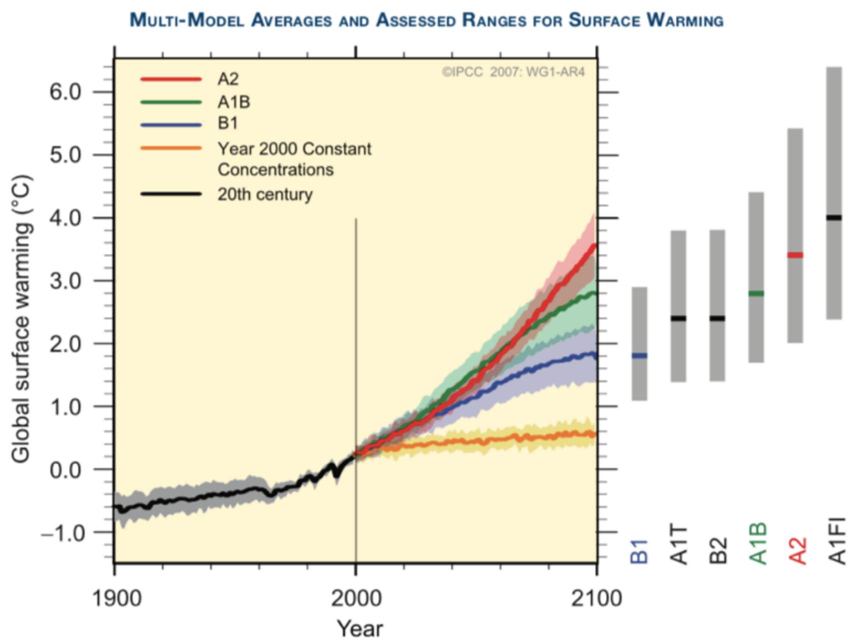


*Figure 2.3: Projected global surface warming by radiative forcing scenario from the IPCC's Fourth Assessment Report, 2007. (Solomon, Dahe, Manning, Avyt, & Marquis, 2007)*

**Table 2.2: Models within the ensemble & the scenario**

| Model | B1 | A1B | A2 |
|---|---|---|---|
| CGCM 3T47 | X | X | |
| CSIROMk 3.5 | X | | |
| HADCM 3 | X | | X |
| MIROC 2.3 | X | X | |
| MRICGCM 2.3.2 | X | | |
| ECHAM 5OM | | X | |
| ECHO-G | | X | |
| INMCM 3.0 | | X | |
| NCARCCSM | | X | |

## 2.4 Hydrologic model (GAWSER)

The Grand River Tier 3 surface water model, created from the Guelph All-Weather Sequential Events Runoff (GAWSER) model was used to produce future daily streamflow and runoff (City of Guelph 2011). The model uses inputs from precipitation and temperature data, to simulate streamflow, runoff, groundwater recharge, and evapotranspiration. The Roseville climate station data was used to force hydrologic response in the Nith River sub-basin where the climate is generalized over the entire Zone of Uniform Meteorology (ZUM) (Figure 2.4). Weather data are averaged over the entire ZUM in order to simulate response in overland runoff, streamflow, and groundwater recharge for each zone (City of Guelph 2011; Shifflett 2014).


*Figure 2.4: A map of the Grand River Watershed outlining the 26 different climate zones. The Nith River subbasin is #18, which is represented by the Roseville station (Shifflett 2014).*

In order to test the suitability of GAWSER to simulate streamflow during the spring freshet, the mean hydrographs of the 30-year baseline simulation (1961-1990) as well as the observed daily streamflow from the same period were compared (Figure 2.5 a). Several randomly selected individual years from each decade were also compared for this purpose, three of which are shown in Figure 2.5 b-d. The baseline scenario was created by forcing GAWSER with observed temperature and precipitation data for the same time period. In general, the streamflow response from GAWSER replicated the timing of flow but underestimated the volume of freshet streamflow. All major features of the hydrograph are represented, including the mean discharge increase in the rainy fall season, as well as the freshet (Figure 2.5 a). On average, GAWSER has lower freshet-period flows than what was observed, and also has higher flows later in the spring than the baseline. Changes resulting from future climate change scenarios used to force GAWSER were always calculated against the baseline scenario rather than observed in order to compare equivalent results.

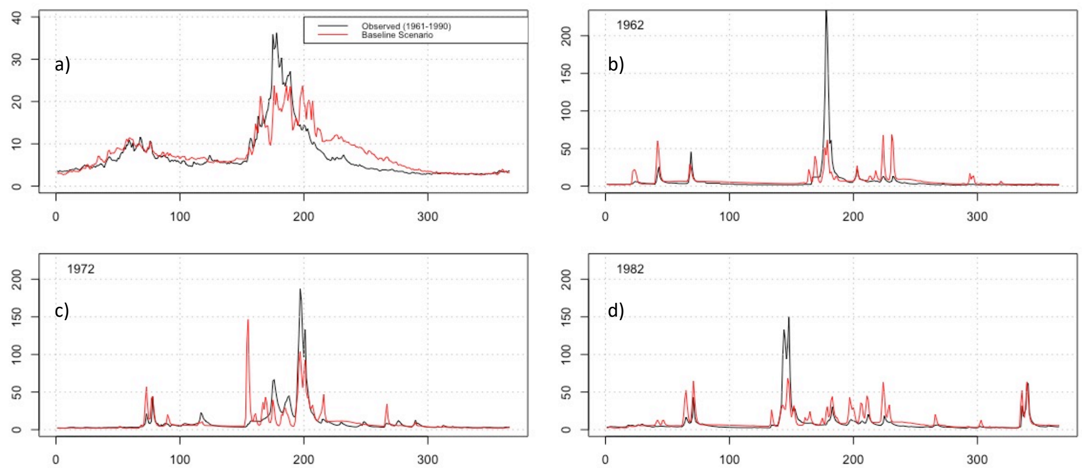


*Figure 2.5: Hydrographs of observed flows (black) vs. the baseline simulated GAWSER flows (red) for the 1961-1990 mean (a), and for 3 individual years (1962 (b), 1972 (c), and 1982 (d)).*

## 2.5 Metrics of hydrologic alteration & statistical analysis

Two aspects of the spring freshet were analyzed for historical trends and projected changes; its approximate timing, and the volumes of freshet-period flows at different river stages. First, the timing of the freshet was quantified using the metric Centre Time (CT), or centre of mass flow. CT is used to analyze the timing of streamflow in studies focusing on temporal shifts and changes to the snowpack and snowmelt is widely used and is more sensitive to change than other metrics such as the date of the annual maximum flow or the percentage of flow occurring during fixed months (Court 1962; Hodgkins and Dudley 2006). CT is calculated by summing all daily flows from a streamflow gauge in one year, and finding the day on which approximately half of the water yield has occurred. This method has been widely used to measure changes to the timing of snowmelt-related streamflow (Clow 2009; Déry et al. 2009; Dudley et al. 2017). It is a fairly coarse estimate of a highly complex and variable process, and it can shift earlier or later due to one or several storm events outside of the snowmelt season (therefore not properly representing the true freshet period). Accordingly, spot-checking several random year's hydrographs from each decade was calculated to compare the CT value, which was then calculated for each year. In the spot-checking procedure, the CT occurred during the freshet period for all years except for the year 2000. During this year, several extreme precipitation events that occurred in the summer changed the CT to day 225 (May 12th). Accordingly, this year was removed from analysis, along with any years when the data were incomplete (9 years total). The resulting correlation coefficient of CT vs. the dates of annual maxima is 0.73, which confirms the suitability of CT to represent the time period when the freshet is occurring.

In order to quantify changes to streamflow volumes at various streamflow stages, flow percentiles over the cold season were calculated at the 10$^{th}$ ($Q_{10}$), 50$^{th}$ ($Q_{50}$), and 90$^{th}$ ($Q_{90}$) percentiles for the period of December to April. The use of low, median, and high flow percentiles allows for analysis of trends in different streamflow conditions throughout the cold season for a more robust understanding of changes than simply using the mean daily streamflow (Campnell et al. 2010; Lins and Slack 2005). Changes to temperature and precipitation can impact the way in which the snowpack forms and melts, which is why it is important to consider trends in streamflow at the lowest flows (typically associated with lower temperatures which keep snow in the snowpack rather than becoming runoff, and/or lower amounts of precipitation) and higher flows (which are associated with more precipitation in the form of rain and higher temperatures).

The Mann-Kendall test was used to analyze trends in hydrologic and climate variables over the observed period at the 95% confidence level. The observed data used in trend analysis (for both CT and at each percentile) were normally distributed, and the assumption of stationarity was satisfied. To project future changes to climate change and hydrologic response, the mean and 95% confidence interval were used to create multi-model means of CT and flow percentiles in future projections. All metrics of change in scenarios B1 and A1B were calculated against the baseline scenario.

# 3. Results

## 3.1 Historical Changes in Streamflow

### 3.1.1 Historical Freshet Timing

The freshet in the Nith River has advanced considerably over the historical period from 1914-2016, where the centre time (CT) has advanced by approximately 17 days (1.7 days per decade, on average) (Figure 3.1). Excluding extreme out-of-season outliers, the earliest observed CT occurred on day 107 in 2012, and the latest occurred on day 198 in 1992. The average CT for the 1914-2016 period occurred on day 166 with a standard deviation of 19 days. Despite this considerable inter-annual variability, the change was statistically significant (Mann-Kendall, P = 0.036).

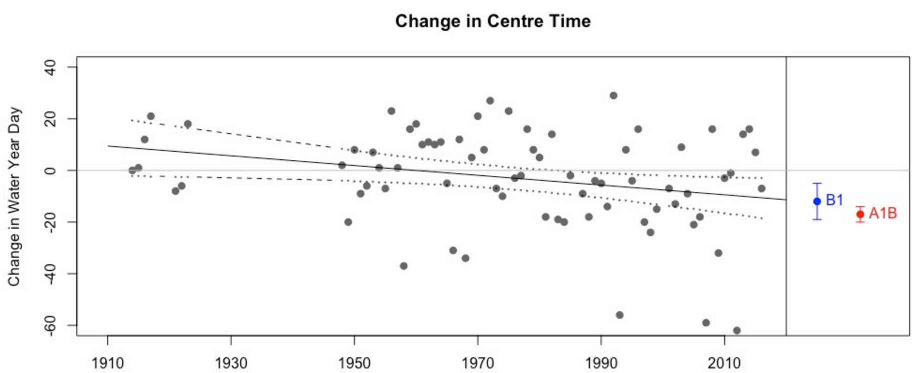


*Figure 3.1: Historical (left, grey) and multi-model average projected (right, blue & red) change in the CT. Historical change is calculated against the 1961-1990 observed annual mean, and projected change is calculated from the baseline model scenario annual mean of the same period.*

For the period of record, the temporal change in CT shows a negative trend (Figure 3.1). The 3 years that exhibited the earliest CT occurred in the last 2 decades (the CT in 1993 occurred on day 113, in 2007 it occurred on day 110, and in 2012, the earliest observed CT occurred on day 107), and the range of CT dates appears to have increased in recent decades as well. Several processes affect the timing of the annual CT. The actual volume of streamflow occurring during the freshet, CT in theory is advanced by increasing streamflow in the first half of the water year (October-March), and/or decreasing streamflow in the second half (April-September). This could be related to several which include decreases in summer precipitation, increases in winter precipitation, and an increase in the frequency of smaller melt events throughout the winter (which release the water that would have otherwise been stored in the snowpack for longer) (Déry et al. 2009). The three earliest CT years were characterized by their significant melt events that occurred earlier in the winter than average, as well as relatively low summer and fall streamflow (Figure 3.2). This not only suggests that CT is an indication of the shifts in the timing of the freshet, but also suggests that the advancing CT from 1914-2016 is due to a greater volume of melt water being discharged earlier in the winter. All three of these earliest CT years do exhibit a pulse of additional streamflow around the time when the average CT occurs in March. However, these years also had a large pulse of streamflow or several smaller pulses occur from November and December onward through the early winter. These years also had average summer streamflow but lacked higher-flow periods or major flood events that would cause the CT to occur later (with the exception of 1993, which had one large flood event occur in June).

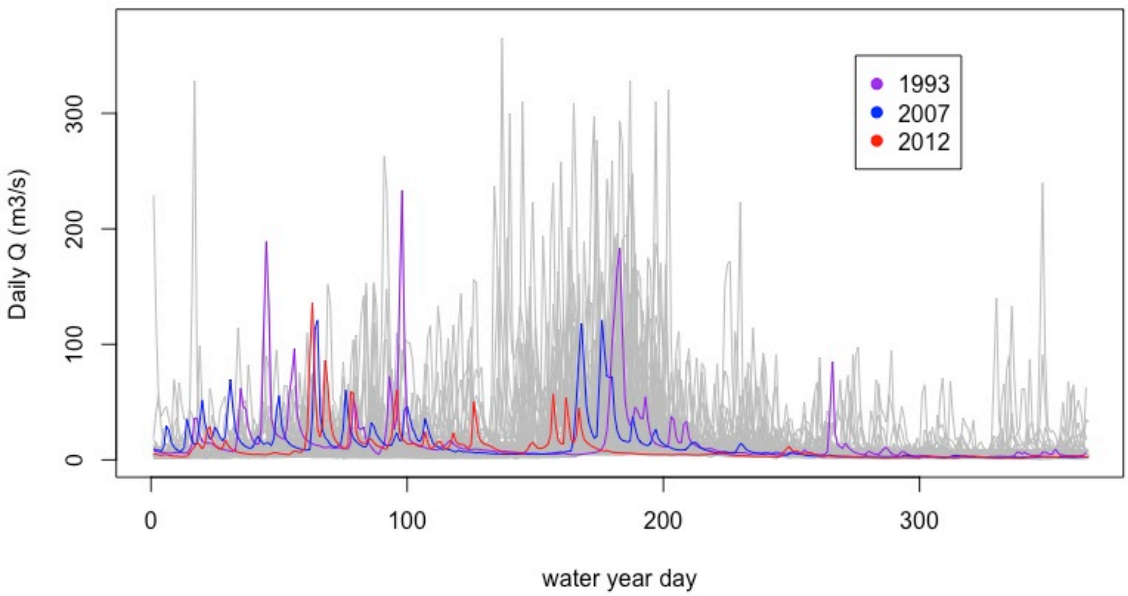


*Figure 3.2: Hydrographs comparing daily streamflow amongst all observed years (grey), and the 3 years with the earliest centre times. All 3 years have a significant freshet pulse during March, but also experienced one or several large pulses as early as November and December.*

The date of the annual daily streamflow maxima was determined and compared to the CT in order to evaluate the utility of CT to characterize the onset of freshet. The CT and date of the annual maximum flow were correlated (correlation coefficient = 0.71, $P < 0.001$, Figure 3.3a) the date of the annual maximum flow advanced over the period of record by 5.6 days per decade (Figure 3.3b).

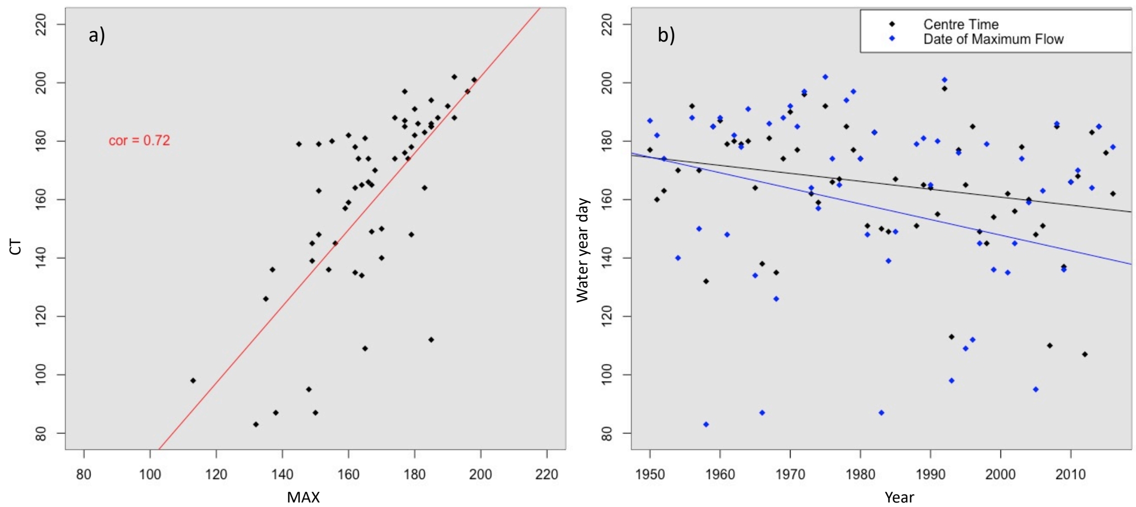


*Figure 3.3: Scatterplot of the annual date of CT vs. date of the maximum daily streamflow (a), and trends in CT and date of maximum flow over the observational period (b).*

### 3.1.2 Historical freshet volume

Knowledge of the volume of streamflow generated during the freshet is of critical importance to water managers because this is typically a period of increased flood risk and active reservoir management. Historically, snowmelt generated streamflow and the runoff during the freshet contribute a large portion of the total annual water yield in the GRW (Shifflett, 2014). Since streamflow during freshet is critical for flood management, and for water supply and irrigation requirements, changes to freshet streamflow will change the way in which water managers can plan to rely on these flows for augmentation in the dry season, and to adapt to changes to flooding in the spring.

The 10th, 50th, and 90th percentiles were calculated from daily streamflow during the defined snow season (December-April). Streamflow percentiles were analyzed at the low-flow

($Q_{10}$), median flow ($Q_{50}$), and high-flow ($Q_{90}$) levels, in order to broadly determine how these different seasonal streamflow conditions have changed over the historical period (Figure 3.4).

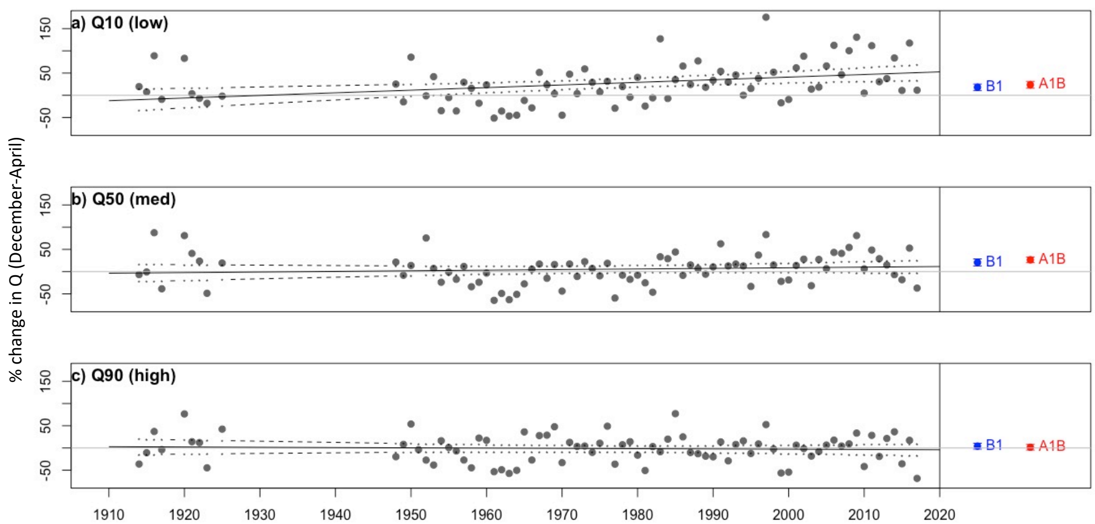


*Figure 3.4: Historical and multi-model projected percentage changes to cold season (December-April) streamflow at low (a), median (b), and high (c) percentiles.*

Observed hydrometric data show an overall increase in winter base flow in the Nith River. In this river system spring runoff is occurring earlier (December-February) than during the historical March/April freshet period which results in a shifting CT and increasing low and median streamflow percentiles. December, January, & February show increase in 90th percentile flows, which appears to be largely related to changes in the form and amount of precipitation. This is discussed further in section 3.1.3 (Figure 3.6). These changes will likely increase as winter temperatures continue to increase, and changes to precipitation occur at the basin scale.

The lowest percentile of flows (Figure 3.4a) has an increasing trend, with a significant Mann-Kendall test ($P = 0.0002$). When average $Q_{10}$ flows were fit using the multiple linear regression (MLR) model with various combinations of climate predictors, low flows were most strongly correlated with increasing temperature. Various combinations of climate variables during the cold season of December-April (mean maximum temperature, total precipitation, total rainfall, total snowfall, and snow proportion) were used, and mean maximum temperature was consistently the only significant predictor of change in streamflow. In a MLR model using maximum temperature, total precipitation, and total snow, 12% of the variation in mean $10^{th}$ percentile streamflow was explained (global $P = 0.02$), and the P-value for temperature was 0.005. Climate change in cold regions has been associated with an intensification of the hydrologic cycle in the winter, because increasing temperatures has decreased snowfall and increased rainfall and runoff (Lins and Slack 2005; Pradhanang et al. 2013; Quilbe et al. 2007).

No significant trends in median percentile flows ($Q_{50}$) were observed in the historical climate data (Figure 3.4b). During the decade 1914-1925, the seasonal 50th percentile and mean streamflow were highly variable. There was no clear trend in historical high-flow levels ($Q_{90}$) (Figure 3.4c). Trends in flood frequency were also evaluated, with flooding being defined by using 2 thresholds; Flood events were first defined as a day where mean streamflow exceeded $120 m^3/s$, which is the threshold for bankfull flow at the Canning stream gauge. The number of flow events $>100 m^3/s$ were determined to capture a broader window of potential high flow events or flooding, where peaks might have exceeded 120m3/s but were not reflected by daily averaging. No clear increase in the annual wintertime flood frequency using daily streamflow was observed. In both cases, the earliest decade on record (1914-1924) had a relatively low number of flood events, and some of the largest and more frequent flood events occurred in more recent decades.

### 3.1.3 Climate trends & correlations

It is widely reported in the literature that the timing of the freshet is most strongly associated with air temperature changes and specifically, in cases where warming winter temperatures have been measured (Burn 2008; Clow 2009; Jones et al. 2015). This supports the idea that the metric of CT adequately captures changes to snowmelt, as despite large differences in precipitation from one year to another, a warming trend can explain the most in streamflow variability (Campnell et al. 2010; Fan and He 2015; Hodgkins et al. 2003; Pradhanang et al. 2013). It has also been widely established in the literature that changes to flow volumes during the snowmelt period are associated with both temperature changes, as well as changes to the amount and phase of precipitation (Campnell et al. 2010; Pradhanang et al. 2013). In order to evaluate the climate conditions responsible for driving changes to the spring freshet, trends in local climate variables from nearby weather stations were analyzed. Precipitation is highly variable interannually in both its quantity and in its solid vs. liquid phase, and it is common for both rain and snow to occur throughout the winter in the Nith River basin and the Grand River. Figure 3.5a-d outline changes during December-April to Mean Maximum Temperature, Total Precipitation, Total Snowfall, and Snow Proportion; historical changes are calculated for each individual year against the average of observed data from 1961-1990, and modeled change are calculated as change from the modeled baseline scenario also spanning from 1961-1990. Despite a relatively short historical record of climate data, mean maximum daily air temperature has increased at an average rate of 0.2°C per decade from 1950-2016. However, this trend is not statistically significant (Figure 3.5a). Over the last three decades, inter-annual variability in mean maximum daily temperature has increased (Figure 3.5a). Significant temperature trends have been previously identified in areas of Southern Canada and Ontario, but results have been mixed (Zhang et al. 2000). The largest amount of warming across much of Canada has occurred during the winter months, and in Ontario, specifically during the spring (Zhang et al. 2000). In the study area no clear trend in total precipitation was observed during the freshet period for the period of record (Figure 3.5b). However, a slight decreasing trend in snowfall or snow proportion has occurred (Figure 3.5c & d). There was no significant change to the annual number of days with rain or snow events in this data record, which was investigated in order to rule out the possibility of changes to precipitation intensity vs. frequency over time.

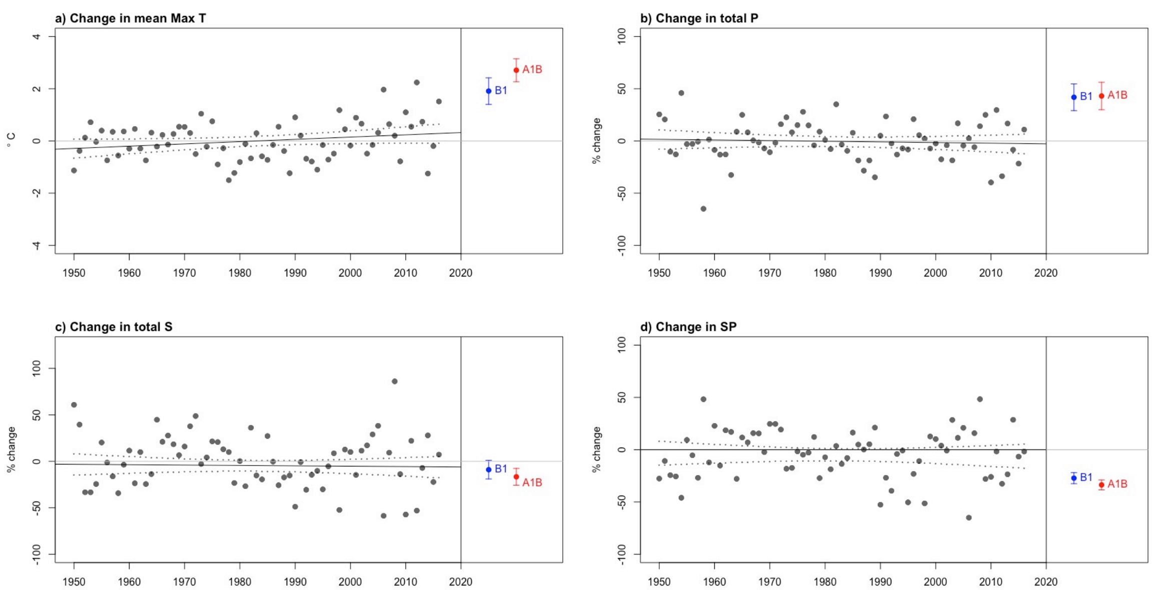


*Figure 3.5: Historical (grey) and projected climate model ensemble mean (blue & red) changes to climate variables during the freshet period (December-April): a) Change in mean maximum daily temperature, b) % Change in total precipitation, c) % change in total snowfall, and d) % change in snow proportion*

In order to understand the relationship of climate variability with the variability in the timing and streamflow during freshet, correlations between hydrologic and climate variables were calculated. Figure 3.6 is a scatterplot matrix between CT, the date of the annual maximum daily streamflow (MAX), streamflow conditions during the 4 previously established phases of the water year, as well as the temperature and precipitation conditions during these periods. Table 3.1 contains abbreviations used in Figure 3.6, and a table containing all correlation coefficients in the scatter matrix can be found in the Appendix (Table 6.1). Unsurprisingly, CT (and by extension, the date of maximum streamflow) are correlated negatively with streamflow in the autumn and early winter (P1) and the snowpack accumulation period (P2), and positively correlated with the freshet period (P3) and the summer (P4). CT is also negatively correlated with temperature in P3 (-0.51), where later CT is associated with cooler temperatures. A weak negative relationship between CT and temperatures is apparent during P2 (-0.31). This observation suggests that colder winters are typically associated with a later CT, because of longer periods of snow cover and its subsequent impact on the timing of streamflow. Similarly, the increased snowfall during P3 will result in a later CT and MAX. During the snowpack accumulation period (P2), streamflow was positively correlated with both CT, rainfall and temperature. Winters that experience warmer temperatures and more rainfall during this time reduces the snowpack through increasing melting. Phase 3 streamflow (F3) had the strongest relationship with snowfall in P2, which indicates the importance of snow accumulation on freshet streamflow. They are also weakly positively correlated with snowfall in P3, but less so than in P2. Snowfall in P1 to P3 is unsurprisingly most strongly correlated with temperature and rainfall during the accumulation phase had the strongest relationship with temperature.

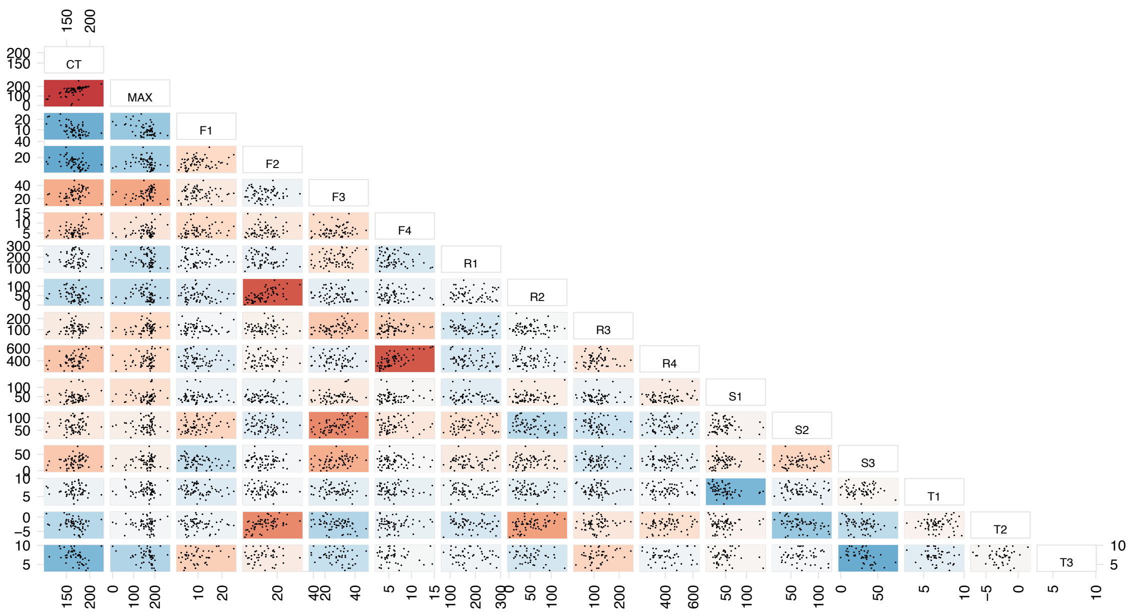


*Figure 3.6: Scatterplot matrix of hydrologic and climate variables. Darker colours represent stronger correlations. Abbreviations of the hydrologic and climate variables represented in the matrix are outlined in Table 3.1, and correlation coefficients for each pair are outlined in Table 3.2.*

**Table 3.1: Abbreviations of hydrologic and climate variables.**

| | |
|---|---|
| CT | Date of Centre Time |
| MAX | Date of annual maximum daily streamflow |
| F1-F4 | Average streamflow in seasonal phases of the water year |
| R1-R4 | Total rainfall (mm) in seasonal phases of the water year |
| S1-S3 | Total snowfall (cm) in seasonal phases of the water year |
| T1-T3 | Average temperature (°C) in seasonal phases of the water year |

A multiple regression approach was used to account for variables that explain variance in CT, using different combinations of climate predictors (R1-R4, S1-S3, & T1-T3) for the cold season of December-April. A combination of variables including temperature, total precipitation, total snow and snow proportion resulted in an adjusted R-squared value of 0.26. Averages and totals from the 4 phases of the water year were also examined but these regressions provided similar results. These relatively weak correlations highlight the need for further research on defining better numerical models for predicting streamflow timing in relation to snow accumulation and ablation dynamics. The common underlying themes in this analysis are that warmer winters are associated with water yields of the Nith River shifting earlier in the year (and higher baseflow occurring in the winter months). These results are likely due to both a precipitation phase shift from snow to rain-dominated precipitation in these years, as well as higher temperatures causing more immediate runoff from snowmelt. Links to changes in flood frequency and flood intensity are complex, as there are many sources of uncertainty associated with this including the natural variability of river flows, the spatial and temporal variability of past precipitation, the difficulty of projecting climate change impacts to precipitation on a regional watershed scale, as well as being physically complex (depending on factors such as temperature, precipitation, the state of the ground in freeze/thaw cycles, and the spatial and temporal occurrence of melt events). There is evidence in the results that this shift has already occurred in the observed record, as well as being projected to continue to shift in future climate change scenarios, despite model and scenario uncertainty.

## 3.2 Future projected changes

### 3.2.1 Projected freshet timing

Results of the hydro-climatic data analysis show that the timing of the freshet is projected to continue to advance as a function of emissions scenario, with CT occurring an average of 12 days earlier than the baseline in scenario B1, and an average of 17 days earlier than baseline in scenario A1B by the 2050s (Figure 3.1). Scenario B1 had greater variability in the CT between models, where A1B and A2 all show an advancement, and all fell within a smaller range from the baseline. All models except one predicted an advancement from the baseline scenario, where CSIROMk3.5 forced with scenario B1 projected no change to CT by the 2050s. Although on average the multi-model mean CT by scenario occurred earlier given the forcing scenario, the spread between all models, including within the same scenario, was varied.

Occurrence of an earlier CT occurring in higher warming scenarios is expected, given that the timing of the freshet and that changes to the distribution of streamflow to earlier in the water year are more related to winter temperature increases. All models show an increase in winter temperatures, and the increase is larger in the higher emissions scenario (Figure 3.5a). There is an overall increase in total precipitation during the snow season (Figure 3.5b). However, warming winter temperatures cause a decline in snowfall (Figure 3.5c), and result in a higher proportion of total precipitation in the form of rain (Figure 3.5d). In scenario B1, the overall mean annual temperature increase from baseline is 1.91°C, and 2.71°C in A1B. The one model outcome forced with scenario A2 showed a mean annual increase of 2.36°C. Many of the largest monthly increases in individual models took place in the winter, although some projected greater summer increases. Overall, the annual and the monthly temperature increases were generally greater in the higher emissions scenarios, with scenario B1 showing smaller annual and monthly increases then in A1B and A2.

Despite this clear signal of temperature response between emissions scenarios, this effect is variable between different models within each scenario, as well as by month. Figure 3.7 outlines the temperature response of different model mean maximum temperature increase by month, including the entire ensemble (Figure 3.7a) as well as the 3 models that were run with different scenarios (Figure 3.7 b-d). Based on these monthly temperature changes alone, it becomes apparent that the model uncertainty (or variance between models within the ensemble) is greater than scenario uncertainty, as the differences in temperature increase between two scenarios within the same model are smaller than the overall differences between models.

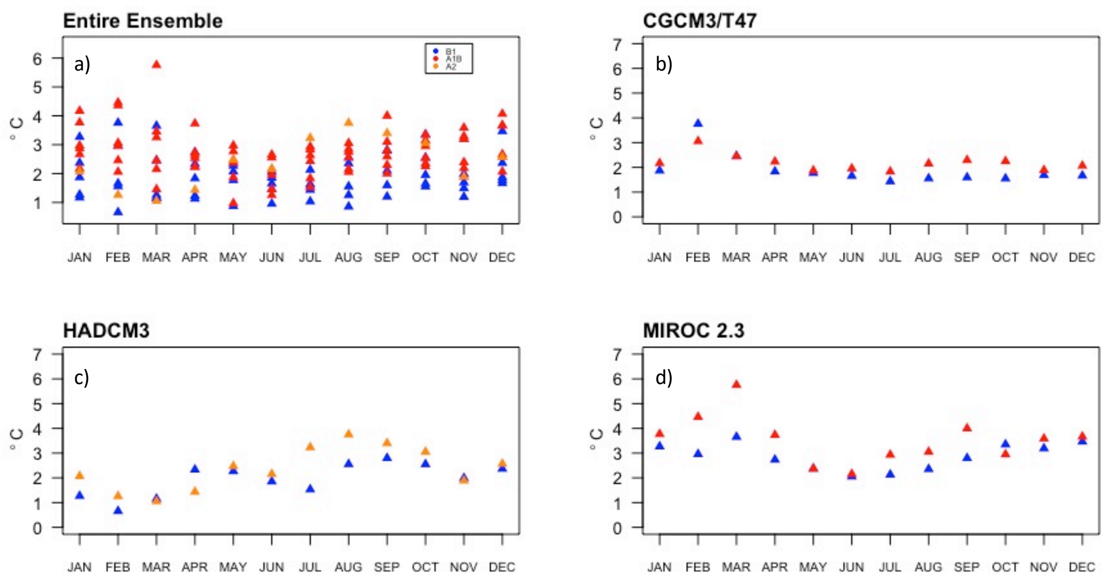


*Figure 3.7: All individual models temperature increase by month (a), and 3 models which were run for multiple scenarios (b-d). Blue indicates scenario B1, red is A1B, and orange is A2. The horizontal dotted lines indicate the multi-model mean annual temperature increase of different scenarios.*

Model uncertainty in projected precipitation also appears larger than scenario uncertainty in future projections. Figure 3.8 shows the projected monthly precipitation change from baseline in the entire ensemble, and separately for the three models that provided two emissions scenarios (CGCM3/T47, HADCM3, & MIROC 2.3). Almost all models show a projected annual increase on average, as well as an increase in precipitation during the winter months. Other times of the year show a more varied signal, May-August having the greatest spread of increases and decreases in precipitation. In the 3 models that provided two scenarios, the similarity of projected monthly precipitation changes between different scenarios within the same model make it apparent that there is greater model uncertainty than scenario uncertainty in these projections. In fact, the average annual projected precipitation increases between scenario B1 and A1B are extremely close (+252.15mm and +254.30mm, respectively).

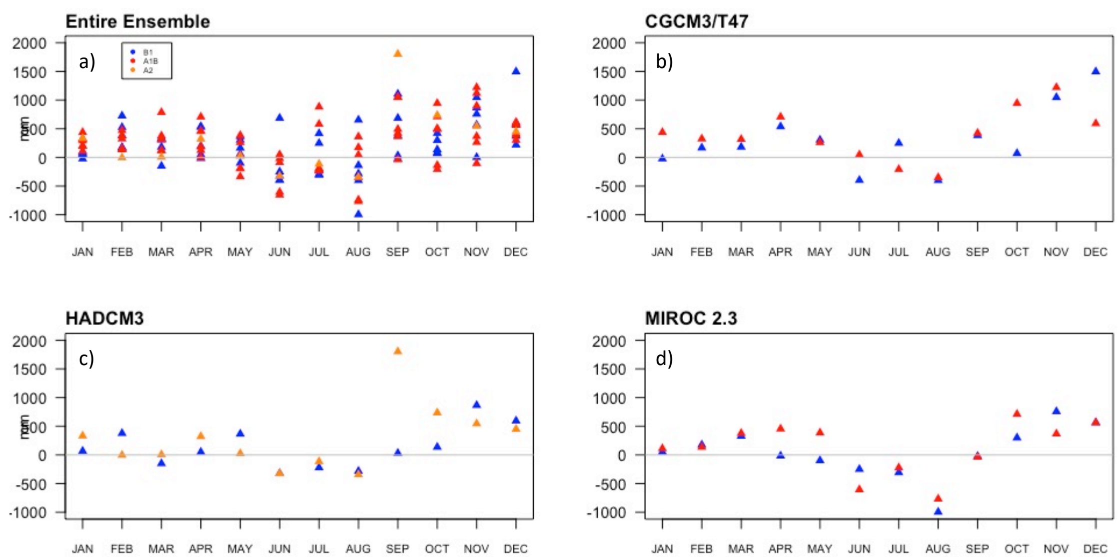


*Figure 3.8: Projected precipitation change by month across all models (a), and in models run for multiple scenarios (b-d). The dotted line shows the mean annual precipitation change, averaged by scenario. This annual increase is extremely close for both B1 (252.15mm) and A1B (254.30mm)*

In spite of this relatively large degree of model uncertainty and variability, given the historical and projected response of CT to winter warming the signal and strength of response of the CT to warming, it is likely that any temperature increase will result in an advancement of the spring freshet, and will increase the portion of annual discharge occurring earlier on in the water year. The model projections show a clear hydrologic response of warmer winters causing an earlier freshet, with less effect of precipitation.

### 3.2.2 Projected streamflow during freshet

All models used in this thesis show an increase in winter temperatures, and the increase is greater in the higher emissions scenario (Figure 3.4a). There is an overall increase in total precipitation during the snow season (Figure 3.4b). However, the warming temperatures cause a decline in snowfall (Figure 3.4c), and a higher proportion of total precipitation in the form of rain (Figure 3.4d). For each climate scenario used to force the hydrologic model, the imposed temperature change derived from the GCM output was not constrained to be uniform throughout the year. Freshet-period streamflow volume is heavily dependent on cold-season temperatures, which dictate the accumulation and melt of the snowpack.

Changes to snowmelt-driven streamflow volumes based on future projections are greater based on the scenario level. Low and median flows were both projected to increase from the baseline scenario during the freshet period, with the multi-model mean of scenario A1B increasing more than B1 (Figures 3.4a & 3.4b). The increases projected at the $Q_{10}$ level do not appear to be of the same magnitude as the average historical trend, but still showed an increase from the modeled baseline. Scenario B1 had an average increase of 18.1% at the $Q_{10}$ level, and scenario A1B increased an average of 23.6%. At the median $Q_{50}$ percentile, streamflow increased 20.8% in scenario B1 from the baseline, and 26.6% in scenario A1B.

There was no change projected at the $Q_{90}$ level for either scenario, similar to the historical period with no detectable change.

While the timing of the freshet is likely more related to temperature changes, streamflow volume during the freshet period is associated with a combination of temperature and precipitation. $Q_{10}$ flows conditions are more closely linked to temperature changes and seem to be characterized by snowmelt rather than precipitation, likely because winters that have higher temperatures are fed more continuously by runoff from snowmelt and rain events, whereas winters that are colder cause much of this water to be stored in the snowpack until later on in the season. This response is seen in the historical record as well as in projections, and the effect is stronger with higher temperatures. $Q_{50}$ and $Q_{90}$ flows are both related to precipitation changes, and where $Q_{50}$ flows are projected to increase slightly more by a combination of both temperature and precipitation increases. Although $Q_{90}$ flows are not projected to change in either future emissions scenario, these are more strongly associated with precipitation increases than temperature increases. The outcome of $Q_{90}$ flows not changing on average in either scenario is likely due to a combination of the fact that precipitation in the future scenarios is highly variable depending on the model used, and also that the change-field method used to calculate future climate does not capture potential changes to the intensity or frequency of precipitation events.

The main impact of climate change on snowmelt-driven runoff both historically and in future projections is the warming of winters, and by extension, the shift of precipitation from snow to rain. Although there is uncertainty when quantifying the freshet's timing and volume in both historical observations and future projections, warmer winters will see a shift in the hydrologic regime where larger portions of annual water yields occur earlier, winters experiencing enhanced and more immediate streamflow in the winter months, and possibly an earlier decrease in the year to the warm-season baseflows.

# 4. Discussion

## 4.1 Summary

The work presented in this thesis suggests strongly that the spring freshet in the GRW is affected by climate change, including the timing and volumes of snowmelt-driven streamflow. This is apparent in the historical record, as well as in the future projection experiments examining the hydrologic outcomes of various climate change scenarios. Climate change will continue to affect changes in hydrologic regimes globally, including during the critical spring freshet in snowmelt dominated watersheds. Detectable shifts in the timing of the freshet in the GRW have been found in the historical streamflow record, with the centre of mass flow (CT) occurring on average 17 days earlier than in 1914 compared to 2016. A significant trend toward increasing low winter flows has been detected over the same period, suggesting that the seasonal cycle of water release is occurring earlier in the water year. Over the same period, no significant trends were detected in median or high flow volumes during the cold season, which could imply several things; that median and higher streamflow is not changing in a straightforward monotonic way, or that changes fall within the envelope of natural variability. Although there are instances of trends and changes in higher flows in historical streamflow records linked to past climate change (Dudley et al. 2017; Jones et al. 2015; Zhang et al. 2000), there are also many cases where historical hydrologic indicators don't show significant change, but in modeled future projections these same parameters show a clear response (Tang and Oki 2016; Thorne et al. 2015).

Two main scenarios were used to test hydrologic response at weak (B1) and moderate (A1B) global warming. In scenario B1, global surface air temperature increases by approximately 1.5°C by the 2050s, and in these experiments in the GRW average annual surface air temperature increases by approximately 2°C in the cold-season (during the period which would affect snow accumulation and melt) (Figure 3.5). Precipitation is expected to increase during the freshet period, but due to temperature increases during this time, this will likely be an increase in rain as the proportion of snow and overall snowfall will likely decrease (Figure 3.5). In scenario A1B, global surface air temperature increase by the 2050s is closer to 2°C warming on average by the 2050s, with an average of approximately 3°C warming during the freshet period in the GRW (Figure 3.5). Precipitation response in scenario A1B during this time in the GRW shows greater spread than in B1, but is not significantly different on average than the response of precipitation in B1. There have been documented instances where model uncertainty is greater than scenario uncertainty, and the spread of precipitation response indicates that this is the case in these experiments as well (Figures 3.7 & 3.8) (Kay et al. 2009a; Vetter et al. 2017). This is discussed further in section 4.3. Under a series of climate change scenarios ranging from weak to severe global warming, the multi-model mean projected change in CT is toward an earlier freshet, with a shift of 12 days by the 2050s in the weak warming scenario (SRES B1), and 17 days in the strong warming scenario (SRES A1B). Low and median flows are expected to increase as the winters warm and more melting events occur. There remains large uncertainty in projected changes to high flow events, and thus flood frequency and intensity, under warming: the multi-model mean shows close to zero change, with individual models projecting changes of both signs. This is likely due to GCM uncertainty, which is often the largest source of uncertainty in modeling future hydrologic response to climate change (when compared to uncertainty from hydrologic models and scenario uncertainty) (Vetter et al. 2017).

### 4.2 Climate change impacts on snowmelt in nival regimes

Although climate change is closely tied to the hydrologic cycle and changes to the hydrologic regime, there are specific ways in which it affects the snowmelt period specifically in cold regions. This has been studied in historical records, as well as observed in experiments using models (Adam et al. 2009; Dudley et al. 2017; Prudhomme and Davies 2009; Tang and Oki 2016). Changes to the snowpack due to increasing winter temperatures have been observed in various mountain environments around the world (Stewart 2009) and this process driving the resulting hydrologic shift can be seen on a smaller scale in the Grand River watershed. A similar study on the modeled impacts of future climate change on river discharge in the Grand River Watershed found that water availability is likely to increase in the winter and decrease in the summer (Li et al. 2016). Notably, this trend is reflected in the actual observed trends in the Nith River where the period of low flow and frequency of increased flood have been observed. This observed increase in streamflow and future projected intensification of flooding means adaptation to changing seasonal streamflow will be necessary in order to continue to mitigate flooding damages as well as ensuring enough water supply for the dry season, in which streamflow is projected to decrease (Li et al. 2016).

The effects of climate change of snowmelt-driven streamflow in nival regimes are largely explained by the changes in the seasonal snowpack, and the shift in these regimes from snow-dominated to rain-dominated (and a decrease in the proportion of snowfall and/or and increase in the amount of rain), as well as warmer winter temperatures creating less-ideal conditions for the accumulation of snow. Snowpack accumulation measured in snow-water equivalent (SWE) decreases significantly with increased winter temperatures (López-Moreno et al. 2013). This smaller snowpack means that less discharge occurs during the typical snowmelt season, as less water is stored over the land surface in what would previously have acted as a reservoir.

Globally, climate change has been found to affect the freshet in several ways in nival regimes. The most notable hydrologic changes caused by climate change are related to trends and projections of warmer winters which result in more snowmelt throughout the winter, and less accumulation of the snowpack. There has been a measurable decline in snow cover extent (SCE) over North America and Europe in recent decades related to increasing winter temperatures (Brown and Mote 2009). In nival regimes in a Canadian context, warmer winters and snowpack decline have also been linked to an advancement of the timing of snowmelt runoff (Burn 2008; Déry et al. 2009; Jones et al. 2015). Both the spatial and temporal extent of snow cover is being reduced, with cold-season snow cover being reduced in northern regions, as well as snowmelt occurring earlier in the season. The results of this study strongly suggest that an earlier freshet is being caused by warming winters, although the predictors of streamflow volume changes are more uncertain. The results suggest that winter baseflow has increased as more water is released as runoff throughout the winter due to precipitation falling as rain than snow, as well as more melt events occurring during the cold season with average temperatures moving closer to the melting threshold. In southern Ontario, significant increases to precipitation variability and extreme events have been observed in fall (October) and spring

(May) (Rudra et al. 2015). If these trends continue, paired with increasing winter temperatures, it is likely that earlier freshet, as well as higher baseflow in the cold season will occur. Although there are inconclusive results regarding high flows in the historical record and median and high flows in the future projections, climate change has the capacity to alter flooding intensity and frequency (Hamlet and Lettenmaier 2007; Kay et al. 2009b). Flooding during the freshet is a key concern for water managers as climate change alters the variability of extreme precipitation and melt events, which are intensified by the introduction of higher winter temperatures and more runoff from rain (Arnell and Gosling 2016; Blöschl et al. 2017). Flood risk management will continue to be of key concern in watershed management, and knowledge of how flooding will impact the GRW is crucial for GRCA reservoir management and flood prevention (see Section 4.3).

Trends that exist in the timing and volume of snowmelt-driven runoff are likely to continue as climate change progresses, because of the fact that these changes exist in the same direction to varying degrees when using climate models. It is likely that in the Northern Hemisphere the decline of the snowpack, and the resulting freshet advance and overall decline of snowmelt-driven streamflow during the spring in the future (Liu et al. 2017). Snow cover extent will likely continue to decline, advancing further north as winter temperatures rise globally. Globally and across North America, a warming climate will mean that less precipitation will fall as snow, and annual snowmelt will occur earlier (Barnett et al. 2005; Cohen et al. 2015). Even without changes to the amount of precipitation falling regionally and globally, these two changes will make for a more immediate response in the hydrologic regime, as the seasonal storage of water in the snowpack will decline. These factors are associated with earlier melt, while changes to the volume of the water yield during the winter and resulting from snowmelt depend on regional precipitation changes.

In the GRW, the results of this thesis show an expected advancement of the freshet, likely caused by temperatures reaching melting point earlier on in the winter (as well as an overall increase in winter temperatures). Results from the MLR models used in an attempt to predict CT with average temperatures for the entire cold season (December-April) as well as the freshet period (March-April) were inconclusive. However, these average temperature increases broken down by season were clearly a function of the greenhouse gas scenario, and the average CT responses were also a function of the scenario used (with greater temperature and precipitation changes occurring in A1B than in B1). GCM projections show that alongside this regime shift of having less snowfall and snow cover and more rain events in the winter due to increasing temperatures, southern Ontario is expected to experience an increase in the intensity and frequency of rainfall storms, which will have hydrologic regime effects (Wang et al. 2015). Collectively, this body of work indicates that it is likely that melting will continue to occur earlier, and more melt events will occur during the cold season as winter temperatures rise in this region. Because water yields and changes to variability and flooding are much more dependent on precipitation changes, and these changes are more difficult to model, more research must be done on downscaling techniques to quantify expected changes to streamflow during the freshet.

### 4.3 Implications for water resource management

The results of this thesis will provide user knowledge regarding the physical science regulating potential future changes to snowmelt-driven streamflow as climate change continues to alter water resources in the GRW. Managing water resources requires knowledge of the timing and availability of water, and how these parameters change because of climate change. The GRCA has identified some key areas for their climate adaptation strategy. These include 1) ensuring adequate water supply for a changing demand base made up of municipal and non-municipal water needs, and 2) adapting and mitigating flood damages as climate variability changes the annual regime (Etienne 2014). This is a complex issue that is associated with a considerable amount of uncertainty. Maintaining adequate streamflow during the dry season is dependent on optimally managing reservoirs and storing water during times of increased water availability, but this also needs to be done with extreme events in mind so that there is adequate reservoir storage to store additional water in the event of flooding to prevent too much streamflow downstream. It is likely that baseflow will increase during the winter and spring due to an increase in rainfall and snowmelt, and also that summers will be affected by increased extreme heat days, periods of drought, and increased intensity of precipitation events (Cohen et al. 2015; Jones et al. 2015; Rudra et al. 2015; Wang et al. 2015). These changes will impact the way that dams are operated and reduce uncertainty in reservoir management.

Water demand in the GRW is dependent on increasing usage needs from both increasing municipal needs as well as non-municipal uses (Etienne 2014). Water supply and demand management are closely tied to continuously evaluating and improving regulations, supply management, and involving municipalities and communities as both supply and demand change. However, these are dependent on future water supply, which is subject to several sources of uncertainty. Some areas of uncertainty that have been addressed in this thesis that are related to modeling future hydrologic conditions are model uncertainty, scenario uncertainty, and uncertainty in the stationarity of historical records. In modelling the hydrologic impacts of climate change, model uncertainty is often the largest when compared to other sources (Prudhomme and Davies 2009; Vetter et al. 2017). Although a certain degree of uncertainty will always exist, it can be reduced by identifying and characterizing uncertainties throughout the process of modelling experiments (rather than using an "ensemble of opportunity") and by excluding poor methods and poor-performing models from experiments (Clark et al. 2016). Identifying and reducing uncertainty in hydrologic modeling and in historical records are addressed further in section 4.4.

There is the potential need for increasing infrastructure in the GRW to manage the available water and changing regime, although major infrastructure upgrades are not without negative ecologic and hydrologic impacts. It is crucial that research is conducted on how water demands will increase as well as how climate change will threaten water supply with increasing summer drought and increasing frequency and intensity of storm events resulting in extreme flow events. The knowledge that the timing of runoff will likely advance and that baseflow will increase in the winter, is a first step in ensuring the key management concerns are addressed.

## 4.4 Conclusions

In conclusion, the results from this thesis suggest that streamflow in the GRW occurs earlier during warmer winters, and that low and median flows increase during warmer winters. There is evidence that some of these shifts have occurred, as there was an increasing trend in both the timing of annual streamflow as well as increasing 10$^{th}$ percentile flows. Experiments using GCMs and a hydrologic model to simulate streamflow conditions under 2 different emissions scenarios also support these results, where the timing of streamflow advanced more when there was a greater increase in temperature, as well as low and median flows increasing more with higher temperatures in the emissions scenarios. These hydrologic changes are a result of changes to climate variables. Temperature can be expected to increase during the winter in the GRW, and a precipitation increase is also expected in southern Ontario across all models, along with a shift in the proportion of snow to rain. Although there was no trend found in 90$^{th}$ percentile flows in the observed record, and no average change in the modeled results, this does not mean that higher flows and extreme flows will not change as climate change progresses. There were differences in the projected change at the 90$^{th}$ flow percentile between models within both the B1 and A1B scenarios, where some models showed an increase in these flows from the baseline, and others showed a decrease. Further research is needed to quantify the expected changes to flood frequency, intensity, and duration. These results fit into similar trends and projections for elsewhere in Canada (Jones et al. 2015; Najafi et al. 2017; Li et al. 2016), as well as across North America and globally (Dudley et al. 2017; Blöschl et al. 2017; Hodgkins and Dudley 2006; Bell et al. 2016). Future warming will result in shorter winters, changes to the timing of high flows in the annual hydrologic regime, which will impact water availability, have potential impacts for flooding and ice jamming, and also potentially impact river ecology  depending on seasonal cycles, all of which are of concern for water resource managers like the GRCA (Jones et al. 2015).

## 4.5 Limitations & future work

### 4.5.1 Downscaling & bias correction

An area of research that still needs to be developed and better understood is in the challenges of and uncertainties associated with downscaling methods from GCM results to regional climate change projections, which are then used to force hydrologic models. Because precipitation is so spatially and temporally variable, and the GCM outputs are on a much larger grid-scale, modeling hydrologic response and change in order to inform future water resources management remains a challenge. For example, GCM outputs regarding temperature and precipitation changes exist on a global scale, whereas information that is useful for watershed management practices occur on a much finer basin scale. This becomes a challenge when trying to plan for future water supply and demand, as water supply is directly linked to changes to precipitation (Fatichi et al. 2016). This can be addressed by identifying the current and the range of potential future hydrologic and climate stressors, by preparing adaptation plans, and by continuously monitoring and reassessing changes and risks (Fatichi et al. 2016).

There are several methods used to downscale GCM projection data to a watershed scale, and the method used in this project was the change field method, or delta change. In this method, the relative changes of GCM outputs from a time slice are used to adjust baseline climate data (Shifflett 2014). This method takes projected changes from a period of time simulated by the GCMs (in this case, 2050s) and adjusts the baseline climate period (1961-1990) at the weather stations within the basin according to these changes. This method makes two major assumptions about the future climate and its resulting hydrologic response; 1) that the changes occur evenly over a large scale (e.g. the grid box of a GCM), and 2) that the relationships between variables and the variability within the baseline climate will not change in the future.

While this method is useful for understanding how averages might change through different scenarios, it does not necessarily capture any changes to extremes or to the variability of climate (e.g. extreme temperature, or the intensity/frequency of precipitation events). This method is used widely and works well for regions with relatively uniform climate, but biases will be higher using this method in cases of either higher elevation or with a low density of weather stations (Bürger et al. 2013). The GRW is located in the Great Lakes Basin, draining into Lake Erie in Port Maitland. Changes to the hydrologic cycle resulting from climate change will be impacted by the presence of the lakes, and therefore, modeling streamflow response to climate change for the purpose of informing water resource managers and aiding in decision-making would benefit from the use of physically-based dynamical models which include these regional features (Erler et al. 2019). The change field method also assumes that the temperature and precipitation variability existing in the present climate, or the boundary conditions used to simulate future hydrologic changes, remains stable and will not change over time. This assumption of stationarity remains an issue in the case of climate change exacerbating things like the intensity and frequency of storm events, changes to extreme cold and warm temperatures, and drought (Räty et al. 2014). It is expected that the GRW will continue to experience overall higher amounts of precipitation, as well as an increase in the intensity and frequency of precipitation events which may not have been captured in this study based on the downscaling method used (Jyrkama and Sykes 2007; Wang et al. 2015). In order to better understand future changes to watershed-scale hydrology, the use of dynamical models which include regional features like topography, fine-scale land surface cover, and in the case of Southern Ontario, the Great Lakes, is necessary. In addition, more research on variability changes and extreme precipitation/drought is necessary.

#### 4.5.2 Selected GCM ensemble

Future work on this topic should examine output from climate change projections using more modern GCM ensembles. At the time that the hydrologic modelling simulations were produced for this thesis, the IPCC's 4$^{th}$ Assessment Report (AR4) experiments were used (Pachauri and Reisinger 2007). This report featured the Climate Model Intercomparison Project Phase 3 (CMIP3) ensemble. Since this initial set of experiments were done, the IPCC released the 5$^{th}$ Assessment Report (AR5), using the CMIP5 ensemble, which featured an expanded suite of experimental protocols, and a larger collection of more sophisticated models (Stocker 2014). There are differences between CMIP3 and CMIP5, however many broader aspects of these experiments such as temperature projections were similar. For radiative forcing scenarios, CMIP3 uses a different family of scenarios from the Special Report on Emissions Scenarios (SRES), including scenarios B1 (moderate) and A1B (high) which were used in this thesis (Pachauri and Reisinger 2007). CMIP5 experiments use a newer set of benchmark emissions scnearios, or Representative Concentration Pathways (RCPs), which include "near-term" (to around 2035) and "long-term" (covering the period to 2100 and beyond) scenarios, including incorporating attempts to stabilize CO2 concentrations into scenarios (Moss and Intergovernmental Panel on Climate Change 2008). Advancements in GCMs can improve things like precipitation distribution, through advancement of finer resolution and added parameters such as evapotranspiration processes and clouds, however even the earliest coarse

GCMs predicted temperature response to greenhouse gas emissions with great accuracy. The distributions of temperature increase between SRES scenarios and CMIP5's RCPs are comparable, including in southern Ontario (Knutson et al. 2013; Sun et al. 2015). The RCP4.5 scenario is comparable to SRES scenario B1 in terms of temperature change by the 2050s, and the RCP6.0 scenario is between B1 and A1B (Mann and Gaudet 2018). From this study, the scenarios used to quantify scenario uncertainty can be used to demonstrate that CT advances increase with more aggressive radiative forcing scenarios, and this would likely respond in a similar manner based on temperature increase, regardless of the ensemble or scenarios used. This study also demonstrates that winter baseflow will likely increase earlier in the cold season (through January and February) as mid-winter melt events increase due to increasing temperatures and a shift in the frequency of snow to rain, as well as a general increase in precipitation. Future research should also be done using the same models for each scenario, in order to truly quantify and separate out model uncertainty from scenario uncertainty. Ideally, this study would be repeated using the entire ensemble of models for each scenario, rather than an ensemble of opportunity using different combinations of models for each scenario. IPCC's 6$^{th}$ Assessment Report (AR6) Physical Science Basis using the 6$^{th}$ Climate Model Intercomparison Project (CMIP6) is currently undergoing experiments is set to be released in 2021. The continued advancement of GCMs with a focus on things like smaller spatial resolution of temperature and precipitation changes and other processes will help improve the use of hydrologic modeling specifically for watershed management and water budgeting on a watershed scale. Information regarding general changes to processes like seasonal streamflow distribution and snowfall/snowmelt, seen in this thesis, will help watershed managers and city planners adapt.

### 4.5.3 Statistical analysis

The assumption of stationarity remains an issue in water resource management, because climate change is likely going to change the boundary conditions of variability (Milly et al. 2008). In southern Ontario, it is projected that the intensity and frequency of extreme rainfall events will increase, potentially creating more unpredictable flooding (Arnell and Gosling 2016; Gizaw and Gan 2016). These potential future changes were not captured within the scope of this study due to the nature of the way the GCM projections were downscaled, but understanding flood frequency should remain an area of further study. In this project, the attempts made at quantifying the response of the timing and volumes of the freshet using a numerical model with seasonal climate averages and totals were unsuccessful. One aspect of this challenge is that calculating CT is not only dependent on the distribution of streamflow during the fall and snowpack accumulation/melt phases, but also on the amount of summertime flow after the freshet is over for the year. Developing a model that can use climate conditions to predict the resulting conditions of streamflow surrounding snowfall and snowmelt to predict timing of the spring peaks is important in deciding how much water to store or release throughout the season. Although the assumption of stationarity is potentially problematic in drawing conclusions regarding the trajectory of hydrologic change, putting weight on the nonstationary nature of some areas of record is also problematic and might actually be "disinformative" to hydrologic models (Koutsoyiannis and Weijs 2015). Hydrologic models must be calibrated using existing data, and the uncertainty in historical data comes from the constraints of timescale (we can only use data that has been recorded in the historical record, which might not fully capture the nature of the hydrologic regime). Regardless of the ability to draw conclusions or not from historical trends, it is telling that the trend in an earlier freshet is reflected in the projected freshet response to warming. It would be beneficial for the GRCA, and in a broader sense, for other water resource managers, to develop a model that is able to better predict the timing of streamflow based on seasonal climate variables in order to make decisions for dam and reservoir use.

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# Appendix

**Table 6.1: Correlation coefficients from Figure 3.6. Shaded boxes are statistically significant (P < 0.05).**

| | | | | | | | | | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| CT | | | | | | | | | | | | | | | |
| <0.001 | MAX | | | | | | | | | | | | | | |
| <0.001 | <0.001 | F1 | | | | | | | | | | | | | |
| 0.002 | 0.123 | 0.074 | F2 | | | | | | | | | | | | |
| 0.003 | <0.001 | 0.855 | 0.962 | F3 | | | | | | | | | | | |
| 0.0170 | 0.629 | 0.062 | 0.321 | 0.034 | F4 | | | | | | | | | | |
| 0.896 | 0.639 | 0.281 | 0.660 | 0.141 | 0.540 | R1 | | | | | | | | | |
| 0.329 | 0.409 | 0.327 | <0.001 | 0.847 | 0.935 | 0.762 | R2 | | | | | | | | |
| 0.332 | 0.048 | 0.625 | 0.511 | 0.007 | 0.198 | 0.733 | 0.924 | R3 | | | | | | | |
| 0.0137 | 0.194 | 0.450 | 0.786 | 0.655 | <0.001 | 0.759 | 0.996 | 0.571 | R4 | | | | | | |
| 0.309 | 0.392 | 0.775 | 0.529 | 0.168 | 0.801 | 0.194 | 0.460 | 0.852 | 0.168 | S1 | | | | | |
| 0.268 | 0.251 | 0.218 | 0.568 | <0.001 | 0.348 | 0.144 | 0.061 | 0.480 | 0.743 | 0.606 | S2 | | | | |
| 0.009 | 0.096 | 0.030 | 0.914 | 0.009 | 0.517 | 0.551 | 0.580 | 0.317 | 0.676 | 0.334 | 0.090 | S3 | | | |
| 0.948 | 0.712 | 0.505 | 0.975 | 0.456 | 0.186 | 0.441 | 0.806 | 0.312 | 0.705 | 0.011 | 0.988 | 0.981 | T1 | | |
| 0.015 | 0.546 | 0.920 | <0.001 | 0.025 | 0.704 | 0.915 | <0.001 | 0.485 | 0.642 | 0.709 | 0.010 | 0.205 | 0.986 | T2 | |
| <0.001 | 0.026 | 0.216 | 0.257 | 0.006 | 0.186 | 0.913 | 0.460 | 0.252 | 0.995 | 0.255 | 0.030 | <0.001 | 0.891 | 0.134 | T3 |

**Figure 6.1: Scatterplots of Kitchener & Roseville weather stations (Daily Maximum Temperature (a), Rainfall (b), and Snowfall (c)):**