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The Impacts Of Climate Change On Precipitation And Hydrology In The Northeastern United States

Justin Guilbert
University of Vermont

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THE IMPACTS OF CLIMATE CHANGE ON PRECIPITATION AND HYDROLOGY IN THE NORTHEASTERN UNITED STATES

A Dissertation Presented

by

Justin Guilbert

to

The Faculty of the Graduate College

of

The University of Vermont

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Dissertation Examination Committee:

Arne Bomblies, Ph.D., Advisor
Donna Rizzo, Ph.D., Co-Advisor
Beverley Wemple, Ph.D., Chairperson
Lesley-Ann Dupigny-Giroux, Ph.D.
Alan K. Betts, Ph. D.
Cynthia J. Forehand, Ph.D., Dean of the Graduate College
Abstract

Shifting climatic regimes can increase or decrease the frequency of extreme hydrologic events (e.g., high and low streamflows) causing large societal and environmental impacts. The impacts are numerous and include human health and safety, the destruction of infrastructure, water resources, nutrient and sediment transport, and within stream ecological health. It is unclear how the hydrology of a given region will shift in response to climate change. This is especially the case in areas that are seasonally snow covered as the interplay of changing temperature, precipitation, and resulting snowpack can lead to an increased risk of flood or drought.

This research aimed to understand the ways temperature and precipitation are changing using general circulation models and observed weather station data in the northeastern United States. With the knowledge that general circulation models do not accurately represent precipitation statistics and trends from the historical period, a large network of climate stations was utilized to further investigate shifts in precipitation. A hydrology model was utilized for further study of regional hydrology. The model used was the Regional Hydro-Ecologic Simulation System, which was calibrated to snow coverage and streamflow for a historical time period. The hydrology model was used to investigate the relationship of snow and streamflow in a changing climate.

We characterized climate change and related impacts in the northeastern United States and estimated a decrease in snowfall of 50% and the number of days below freezing by 45 days by the end of the century. We also showed that precipitation is not only becoming more intense: but it is also more persistent – a finding that may have significant hydrological implications including increased flood risk throughout the year. The 95th percentile of daily precipitation has increased by 0.5 mm per day per decade, while the probability of successive days with precipitation increased by 0.6 percent per decade. We also explored the role of snowpack in a changing climate. We found that temperature plays a larger role than precipitation in shifting hydrologic regime, because the warming-induced reduction of snowpack reduced the maximum flows more than the increasing precipitation increased the maximum flows. However, because of the increasing intensity and persistence of precipitation, instantaneous peak flows occurring outside of the snowmelt season will likely continue to increase during all times of the year. We shed light on the complexity of the modes of climate change and the interactions that increases in temperature and precipitation can have on the hydrology of a region.
Citations

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Chapter 1: Introduction

1.1 Motivation

Climate change as a whole is a result of global processes, but is locally realized through non-stationarity in temperature and precipitation with pronounced trends in extreme precipitation as one of the primary concerns. In recent memory, most people remember and many were impacted by severe flooding whose likelihood has increased as a partial result of anthropogenic climate change (Milly et al., 2002). The northeastern United States has experienced the largest increasing trend in the frequency of severe floods (Groisman et al., 2001) in the nation. Thus, the Northeast is a very interesting case study of climate change impacts on regional hydrology. Well-documented events such as Tropical Storm Irene and Hurricane Sandy made landfall in the northeastern United States with devastating results. However, much smaller precipitation events have caused similar magnitude flooding events on a localized scale due to prior climatic conditions and increased rainfall persistence. For example, in central Vermont during the summer of 2013 there was a flood in early July that resulted from a precipitation event occurring on saturated soils. In some areas, the flows generated during this flood were comparable to the flows generated by Tropical Storm Irene but the precipitation event that immediately preceded the peak flow was much smaller in magnitude. The setting for this event was persistent rainfall in May and June due to a stalled atmospheric flow, which primed local watersheds by saturating soils, and was followed weeks later by a relatively intense precipitation event that triggered the flooding. This example illustrates the disconnect between trends in precipitation and trends in flooding, where increases in individual
precipitation events are not linearly correlated with increases in flooding. In general, the translation of climate signals to flow signals is highly nonlinear and the watershed response to climate change is not well understood. The knowledge gap exists in the combined effects that climate change can have on a watershed. With a rise in temperature, the dynamics of wintertime precipitation may change, snowpack will be more prone to melting (although more winter precipitation may make up for this deficit), evaporation may increase, and expected higher water vapor pressures may yield more precipitation. Due to the complex interactions of these various processes, teasing apart the various complex linkages of changing hydrologic regime and flood risk requires a hydrologic model rooted in physical processes?. The impact of warming on flood risk remains poorly understood but is of great societal relevance.

In this research I investigated how increasing temperature may impact hydrology throughout the water year. Overall, I seek to identify the impacts that climate change has on precipitation in the northeastern United States and how shifts in precipitation and temperature may increase or decrease the likelihood of severe flooding. This is significant for purposes of planning and policy. Understanding the likelihood of given severe flood magnitudes can provide valuable information to stakeholders responsible for the delineation of flood plains, the delineation of flood zones for insurance, the size and composition of engineered structures, and setting limits on nutrients and sediment transported to larger bodies of water.

To address the uncertainties of climate change impacts on hydrology in the study region, the probable changes in temperature and precipitation must be determined. Due to the
complicated nature of precipitation changes, further study must be completed to more fully understand the varying ways that precipitation change can be manifested. To do this, a historical data set must be used, specifically because the downscaled general circulation model precipitation series perform poorly in parts of the northeast (Mohammed et al., 2015). Combining the results of temperature change generated by the general circulation models and the observed historical precipitation changes will allow greater understanding of future climate. These results may be used to further force a hydrology model to analyze shifts in snow and streamflow in future climate scenarios.

1.2 Research Goals and Hypotheses

The broad goals of this research are to (1) determine the shifts in mean temperature and precipitation as projected by general circulation models, (2) convey these results as digestible climate change impact metrics, (3) identify shifts in precipitation from observed records across the United States with focus on the persistence and intensity, (4) calibrate a hydrology model to snow coverage data and streamflow data for a small mountainous watershed, and (5) investigate the interaction of a shift in climate and hydrology with a focus on the role of snow. The overarching hypotheses of this research are:

1) Under future emission scenarios as developed by the Intergovernmental Panel on Climate Change, there will be an identifiable shift in temperature and precipitation.

2) Precipitation changes will not be well modeled by general circulation models.
3) Precipitation shifts will be identifiable in the historic records of long standing climate stations.

4) A detailed physically based hydrology model will identify the hydrological impacts of a changing climate.

5) Increases in temperature and precipitation will impact future snow packs and the timing of snowmelt, which will shift the timing, total volume and peak streamflow magnitude of snowmelt driven streamflow.

1.3 Dissertation Organization

The organization of this dissertation is as follows:

Chapter 1 is the introduction, which provides the motivations, objectives and hypotheses that build the basis for the following work.

Chapter 2 is a comprehensive literature review that contains pertinent information to provide a background of global climate change, general circulation models, and the hydrological impacts of climate change.

Chapter 3 is a study of the projections of precipitation and temperature under different emissions scenarios where impact metrics were derived as a way to aid end users in evaluating and quantifying the impacts of climate change.

Chapter 4 is an intensive study of all available observed precipitation data in the northeastern United States. Historic precipitation data were analyzed by month and annually for anticipated shifts in both precipitation intensities and persistence.
Chapter 5 is an evaluation of climate change scenarios forcing a hydrology model. Snow coverage and stream gauge data were used to calibrate a hydrology model. The model was then used to analyze the impacts of climate change on snow and streamflow statistics with varying degrees of shifts in temperature and precipitation.

Chapter 6 provides a summary of results and suggests future work that builds on the findings studying this dissertation.
Chapter 2 : Comprehensive Literature Review

2.1 Global Climate Change Overview

Human activity has shifted the composition of Earth’s atmosphere to such an extent that it is part of the reason that climate scientists refer to our current geological epoch as the Anthropocene (Crutzen, 2002). Human activity during the industrial period has elevated the concentrations of greenhouse gases (GHGs) such as methane, carbon dioxide and nitrous oxide (Collins et al., 2007), with current concentrations at their highest levels in at least the last 800,000 years (IPCC, 2014). An increased greenhouse effect can significantly change the global climate with dramatic regional consequences (Crutzen, 2002). Because human population is expected to reach 10 billion by the end of the century, and per capita energy use is expected to increase, without intervention GHG concentrations will continue to increase, exacerbating the current radiative imbalance on Earth (IPCC, 2014).

The Mauna Loa Observatory has been recording measurements of atmospheric carbon dioxide since 1958, at which time the annual average carbon dioxide concentration was approximately 316 ppm (Keeling et al., 1976). In May 2013, the station recorded a measurement above 400ppm for the first time; and as of May 2016, the weekly average value exceeded 408 ppm (esrl.noaa.gov). Paleoclimate records from ice cores show significant GHG variability throughout the past, but also indicate an unprecedented rate beginning with the industrial revolution (eg. Craig & Chou, 1982; Fischer et al., 1999; Otto-Bliesner et al., 2006; Pearman et al., 1986; Rasmussen et al., 1984).
One study found that the increase in carbon dioxide concentration lagged behind the increase in temperature from analysis of an Antarctic ice core (Fischer et al., 1999). This was then explained in a study that compared temperature records from around the world during the last deglaciation and found that carbon dioxide concentration increases do, in fact, precede increases in temperatures (Shakun et al., 2012). Both findings were accurate, as temperature increases did precede carbon dioxide in the Antarctic. However, this is believed to be due to a weakening of the Atlantic meridional overturning circulation (AMOC) that allowed the southern hemisphere to warm faster than the northern hemisphere (Clark et al., 2004). This see-saw effect helped deglaciate the southern oceans, increasing their temperatures, which led to the release of carbon dioxide from deep within the far southern oceans (Skinner et al., 2010). This furthered the greenhouse effect of the earth’s atmosphere, which led to further warming of the global atmosphere during the last interglacial period (Skinner et al., 2010). Currently the earth is experiencing warming that is often compared to the last interglacial period (eg. Kubatzki et al., 2000; Otto-Bliesner et al., 2006). However, human activity, particularly the combustion of fossil fuels, is the primary cause of current warming (Collins et al., 2007) unlike in the last interglacial period where orbital forcing was the primary cause (Otto-Bliesner et al., 2006).

2.1.1 Impacts of Global Climate Change

The impacts of global climate change have been realized globally; but they vary regionally in magnitude and direction. For example, rapidly warming arctic temperatures have reduced sea ice coverage, resulting in a strong positive feedback (Stroeve et al.,
2008; Wang & Overland, 2009). Also, as a result of mass loss from terrestrial ice sheets, sea level rise has been observed (Gardner et al., 2013). Furthermore, there has been a decrease in snow cover, extent and duration in many areas (Fassnacht et al., 2016), an increase in heat waves (Meehl & Tebaldi, 2004), an increased intensity of tropical storms (Knutson et al., 2013), and a poleward shift of the peak intensity of tropical storms (Kossin et al., 2014). Also, when combined with sea level rise, rare storm surges may become more frequent causing devastation in low lying economic centers such as New York City as illustrated by Hurricane Sandy in 2012 (Lin et al., 2012). Overall, a primary impact of global climate change is that the strength and frequency of extreme weather has increased and is likely to continue to increase (Coumou & Rahmstorf, 2012; Francis & Vavrus, 2012).

An observed increase in extreme events includes increases in drought; and as such, it is expected that more people globally will be water stressed in the coming decades (Arnell, 1999). This may be especially true for the tropics (Dore, 2005) and those in snowmelt dominated regions such as the pacific northwestern United States, the Colorado River Basin and the large cities of arid coastal Peru (e.g. Lima) (e.g., Barnett et al., 2005). The return periods of prolonged dryness have been decreasing across the eastern and southwestern United States (Groisman & Knight, 2008), which is potentially hazardous for ecosystem health and agriculture.

Ecologically, there has been a movement of species in some locations and an expansion of their natural ranges toward the poles (Hickling et al. , 2006; Parmesan & Yohe, 2003; Thomas & Lennon, 1999). Certain species are expected to become extinct by the middle
of the century due to a lack of mobility and inability to adapt with shifting climates and loss of habitat (Root et al., 2003; Thomas et al., 2004). Certain cold, fresh water fish populations have been challenged as streamflows decrease and water temperatures increase (Dudley et al., 2008).

Regionally, in the northeastern United States, the impacts of climate change have also been observed; the duration, depth and extent of snow cover have been decreasing (Campbell et al., 2010; Huntington et al., 2004). As a result, the ski industry in the region is likely to experience economic hardships in the coming century as energy requirements increase and participation decreases (Dawson & Scott, 2007; Guilbert et al., 2014). The duration of ice cover on lakes has been observed to be decreasing (Betts, 2011; Hodgkins et al., 2002). In the future he maple syrup season will likely occur earlier and be shorter (Skinner et al., 2010). Hydrologically, there has been an increase in the frequency of flooding (Armstrong et al., 2012; Collins, 2009) and extreme precipitation events (Douglas & Fairbank, 2011), which are discussed in greater detail in sections 2.1.3, 2.1.4, and 2.3.

2.1.2 Positive Feedbacks of Global Warming

GHGs alone cannot fully explain temperature increases, as there are a number of positive feedbacks that result from the warming. A rise in temperatures increases evaporation and increases the ability of the earth’s atmosphere to hold water vapor, which is a potent greenhouse gas (Held & Soden, 2000). A study in 1861 first showed that water vapor has the ability to absorb radiation (Tyndall, 1861). Since then, water vapor has been
identified as the most important greenhouse gas, which accounts for approximately 60% of the greenhouse effect in clear sky conditions (Kiehl & Trenberth, 1997). Increases in the concentration of water vapor in the upper troposphere are particularly important because the amount of outgoing terrestrial radiation that is trapped is proportional to the logarithm of water-vapor concentration (Allan, 2011; Held & Soden, 2000). It is expected that the upper troposphere will experience amplified moistening as a feedback resulting from a warming climate (Chung et al., 2014). Increased water vapor in the atmosphere then increases the sensitivity of the earth’s surface temperature to increases in carbon dioxide and other greenhouse gases (Held & Soden, 2000).

With increases in temperature, arctic sea ice has been shown to be decreasing in area by 12 percent per decade in the summer, and 2.7% per decade in the winter (Cavalieri & Parkinson, 2012; Stroeve et al., 2008; Stroeve et al., 2014). The trend in September sea ice is also thought to be decreasing at an increasing rate as sea ice is more frequently comprised of seasonal ice (Serreze & Stroeve, 2015). With increasing temperatures, it is thought that in non-polar regions that snowfall will decrease and snowmelt will occur earlier (Stewart, 2009). Interestingly, snow cover extent is thought to be increasing in some areas (Cohen et al., 2012). The western United States has experienced a decrease in snow water equivalent from 1925 to 2000 (Mote et al., 2005). This decrease may decrease the persistence of snowpack, which could decrease the average annual surface albedo of many seasonally snow covered regions. Many studies have also shown that ice out dates on lakes around the globe are occurring earlier with time (Anderson et al., 1996; Betts, 2011; Palecki & Barry, 1986; Robertson et al., 1992; Schindler et al., 1990). Both
decreases in snow cover duration and the area covered by ice decrease albedo allowing more radiation to be absorbed by land and ocean masses, which in turn causes further increases in temperatures (Griggs & Noguer, 2002; Ingram et al., 1989).

The loss of ice in high latitude regions has amplified warming in those regions to a rate of twice the global average (IPCC, 2014). This amplified warming is causing permafrost to melt (Brown & Romanovsky, 2008; Romanovsky et al., 2010). Permafrost is thought to contain twice the amount of carbon currently in the atmosphere (Tarnocai et al., 2009; Zimov et al., 2006). As permafrost melts, the carbon stored within the permafrost is then made available for decomposition by microorganisms within the soil. This carbon, once broken down, is released as carbon dioxide or methane. Therefore, with decreases in permafrost, more methane will be released from previously frozen areas (Walter et al., 2006). This results in a positive feedback that is unknown with respect to the effects on further amplifying warming.

2.1.3 Precipitation Shifts

In general, it has been suggested that wet regions have become wetter, and dry regions have become drier as a result of a changing climate (Dore, 2005); although a recent paper challenges that commonly-cited global impact (Ljungqvist et al., 2016). The intensification of the hydrologic cycle due to warming is rooted in physics: the Clausius-Clapeyron relation states that both the saturation vapor pressure of water increases with temperature, as does the rate of increase. This temperature sensitivity of atmospheric water vapor gives rise to the water vapor feedback, because water vapor itself is a
greenhouse gas. Many have noted an acceleration or intensification of the water cycle associated with this relationship (Berg et al., 2013; Durack et al., 2012; Loaiciga et al., 1996; Trenberth, 1999). It has been suggested that increases in precipitation could increase by 3.4% per degree Kelvin due to energy limitations rather than the 6.5% per degree Kelvin as suggested by the Clausius-Clapeyron relation (Allen & Ingram, 2002).

As discussed earlier, arctic sea ice is retreating as a part of a positive feedback to a warming climate. Sea ice cover has been shown to be much shorter in duration over large spatial areas (Stammerjohn et al., 2012); and the Arctic Ocean is predicted to be ice free seasonally within 30 years (Wang & Overland, 2009). This has increased the flux of moisture and temperature from the ocean into the atmosphere in the arctic. This phenomena has been linked with shifting summer precipitation patterns occurring across Europe and Asia (Guo et al., 2014; Screen, 2013; Wu et al., 2013; Wu et al., 2009). Francis et al. (2009) found that when the spatial extents of summer sea ice are below average, there is a significant tendency toward increased precipitation in the following winter over the region north of 40°N, which includes our study region.

Anthropogenic GHGs and associated increases in temperature are often solely attributed to increasing the intensity of precipitation, Groisman et al. (2012) suggest other anthropogenic factors, specifically land-use changes that contribute to high intensity precipitation events. DeAngelis et al. (2010) studied the impacts of increased irrigation rates across the Great Plains as a partial explanation for increases in summer precipitation. Zangvil et al. (2004) found greater precipitation recycling with increasing crop yields. Precipitation recycling refers to the contribution of evapotranspiration to
local precipitation (Eltahir & Bras, 1996). Dams were built across U.S. rivers at approximately the same time as widespread implementation of large-scale irrigation; and currently, it is believed that up to 1 year of mean runoff in the United States is stored within reservoirs (Graf, 1999). Increased numbers of reservoirs have allowed for a greater amount of evaporation from open water areas in the middle of the United States, which allows for greater levels of precipitation recycling (Eltahir & Bras, 1996). Many studies have investigated the impacts of land-use change on the regional water cycle and precipitation intensity (e.g. Avissar & Liu, 1996; Eddy et al., 1975; Feddema et al., 2005; Mahmood et al., 2010; Sacks et al., 2008). When coupling land-use change with increasing temperatures, this greatly magnifies the evaporation occurring across the United States.

When observing what is happening in our study region specifically, Keim et al. (2005) found positive trends in precipitation across the northeastern United States (NE US) but noted that trends differ across subsets of this region. Groisman et al., (2005) found spatially divergent precipitation trends within the Lake Champlain basin, a sub-region of the NE US, with linear trends between 1900-2002 of approximately +10% in the northern section but -10% in the southern part of the watershed. Ahmed et al. (2012) similarly found that maximum 5-day precipitation increased in some areas of the Lake Champlain Basin, while decreasing in others. The magnitude of historical precipitation change in the Lake Champlain basin also varied from 8% to 38% at low and high elevations, respectively, over the 40 year period from 1963 to 2003 (Beckage et al., 2008).
The NE US has experienced an annual increase in precipitation of approximately 0.4 inches per decade and has experienced the greatest increases in extreme precipitation in the United States (Horton et al., 2014). For example, the return period of daily rainfall intensity greater than 101.6mm (4 inches) has decreased in the last century from 26 to 11 years in the NE US, while the frequency of the upper 10% of rainy days has increased (Groisman et al., 2005; Groisman et al., 2001). Groisman et al. (2012) suggested a theory associating increased precipitation intensities in the central United States with increased irrigation and increases in the areas covered by reservoirs. Similar to the association suggested by Groisman et al. (2012), I believe that the decreasing duration of snow cover across southern Canada and ice cover on Hudson Bay may work as a similar mechanism that allows more evaporation to occur. This mechanism cannot be ruled out as a contributor to the observed increases in precipitation intensities in the northeastern United States.

Under a recently-proposed mechanism yielding slower-moving planetary waves (Francis & Vavrus, 2012), storms are expected to propagate more slowly, resulting in more persistent weather patterns (Guilbert et al., 2015). Precipitation magnitudes in the NE US have been shown to have little dependence on large-scale climate variability (Brown et al., 2010; Dai, 2013) and demonstrate pronounced multi-decadal variability in paleoclimate records that exceeds recent variability in magnitude (Brown et al., 2002; Brown et al., 2000; Noren et al., 2002; Stager et al., 2016). Teleconnections as described by known indicators are somewhat tenuous, as Brown et al. (2010) considered six
teleconnection patterns, while Dai (2013) looked only at the inter-decadal Pacific oscillation.

2.1.4 Large Scale Atmospheric Processes and Climate Change

The arctic region is warming at double the rate of the rest of the northern hemisphere, due to the arctic amplification feedback previously described (Screen et al., 2010; Serreze et al., 2009). A dominant hypothesis suggests that weakening meridional temperature gradients associated with arctic amplification cause a decrease in the meridional pressure gradient. As this pressure gradient decreases, the upper level flow of the jet stream decreases, which is correlated with amplified atmospheric waves and decreasing translational velocities (Francis & Vavrus, 2012, 2015). This atmospheric setup favors persistent weather regimes that can lead to extreme flooding, drought, heat waves, and cold periods (Francis & Vavrus, 2012; James A. Screen & Simmonds, 2013). This explanation for a linkage between arctic amplification and persistent weather may help explain the large number of extreme persistent events in recent years (Coumou & Rahmstorf, 2012) and has been noted regionally in the northeastern United States (Guilbert et al., 2015). Liu et al. (2012) describe a similar physical mechanism that relates a loss sea ice to meridional meanders in the troposphere during the fall and persistent blocking patterns during the winter; while Petoukhov et al. (2013) found that amplified Rossby waves result from increased mean temperatures at the mid-latitudes. All of these hypothesized mechanisms support the notion of an increasing frequency of blocking patterns; and blocking patterns are often associated with extreme weather (eg. Barnes et al., 2014; Buehler et al., 2011; Matsueda, 2011; Whan et al., 2016). However, Hopschet
al. (2012) noted that although these mechanisms may be possible, there is currently a lack of evidence to support these findings in a statistically significant manner. Therefore, the hypothesis suggesting that shifts in geopotential heights have affected mid-latitude Rossby waves remains unresolved (Barnes et al., 2014; Francis & Vavrus, 2012; Kintisch, 2014; James A. Screen & Simmonds, 2013). Some other hypotheses suggest that increases in sea surface temperatures could play a similar role (Muller, 2013; Palmer, 2014). Palmer (2014) suggests that increases in sea surface temperatures lead to powerful storms over the western Pacific Ocean and that the release of latent energy from these storms causes propagating wave trains that amplify planetary waves. Muller (2013) suggests that increases in sea surface temperatures lead to organized squall lines in convective systems that lead to increased extreme precipitation.

2.2 General Circulation Models

General Circulation models (GCMs) are a mathematical representation of the physical processes in the atmosphere, land surface, cryosphere and ocean used to simulate the response of climate to increases in greenhouse gas concentrations. GCMs can be used when simulating climate change scenarios where the results must be geographically consistent. One major issue with these models is that their gridded scale is typically on the order of hundreds of kilometers.

The Coupled Model Intercomparison Project phase 5 (CMIP5) supplies daily data from the most current version of 20 GCMs from around the world that are forced by the most current projections of atmospheric conditions supplied by the IPCC’s Fifth Assessment Report. GCMs are often downscaled using an algorithm such as bias correction with
constructed analogue (BCCA, Brekke et al., 2013) to allow smaller spatial resolution models to be run with future projections. BCCA methods leverage observed climate data to eliminate any tendencies for a given model towards hot, cold, wet or dry; the data are then downscaled by searching the observed record for similar climatic conditions and combining the 30 closest matches at a higher spatial resolution. A full description of the BCCA methodology may be found in Hidalgo et al. (2008), Maurer, Hidalgo, et al., (2008) and Maurer et al., (2010). Relative to Coupled Model Intercomparison Project phase 3 (CMIP3), the CMIP5 models generally have higher spatial resolution and contain a number of improvements that have led to better simulations of some climate features (Knutti & Sedlávcek, 2012; Julienne C. Stroeve et al., 2012). CMIP5 models are driven by representative concentration pathways (RCPs; Moss et al., 2010) that include a set of greenhouse gas emission, aerosol, and land-use changes scenarios developed as input for climate modeling experiments. CMIP5 simulations use four RCPs distinguished by their radiative forcing at the end of the century. The radiative forcing in these scenarios ranges from 2.6 to 8.5W/m², where larger numbers represent higher emissions scenarios that result in higher net radiation. Additional details on the construction of RCPs may be found in Moss et al. (2010).

2.2.1 Downscaling: The Delta Method

The delta method has been used extensively in studies with general circulation model projections to aid in climate change impact studies (e.g., Elsner et al., 2010; Hamlet & Lettenmaier, 1999; Snover et al., 2003). The delta method is an algorithm that is capable of adding the variability of past climates to future projections. This is accomplished in a
two-step approach, which first generates a delta or delta factor and then modifies an observed series by the delta or delta factor. For temperatures, deltas are generated by subtracting historic modeled values from future projected values. Temperature deltas are then added to a historic observed time series of temperatures (Maurer et al., 2010). For precipitation, delta factors are generated by dividing future projected values by historic modeled values. Precipitation delta factors are then multiplied by a historic observed time series of precipitation. This can be performed at any temporal resolution from annually to monthly; however, using monthly deltas and delta factors allows projected seasonal trends to be maintained.

2.3 Hydrological Impacts and Climate Change

Changes in flow record statistics can result from changes in land use and land cover, changes in water resource management and changes in atmospheric conditions, among others. Globally, statistically significant trends have been found in the increasing frequency of severe floods (Milly et al., 2002), while drought has been shown to be relatively constant over the last 60 years (Sheffield et al., 2012). In the United States, normalized losses associated with extreme flooding events have increased with time in contrast with other extreme weather events whose associated losses have remained relatively constant (Changnon, 2003). Flood magnitudes have been shown to be decreasing in the Southwest and increasing in the Midwest and East (Peterson et al., 2013). Many have noted that New England has experienced increased flooding (Collins, 2009; Hodgkins, 2010; Villarini et al., 2010).
Snow plays a critical role in the hydrology of a region. Large areas of the northern hemisphere rely on snow accumulation to ensure water resource supply (Barnett et al., 2005; Viviroli et al., 2007). Many studies have looked into the reaction of a mountainous watershed to a changing climate (Bavay et al., 2009; Horton et al., 2006; Singh & Kumar, 1997; Stahl et al., 2008). Future snow fall, accumulation and melt will be greatly impacted by increases in precipitation and temperature. The ratio of snow to rain has been observed to be decreasing in time (Knowles et al., 2006). Mountainous regions in the western United States have already experienced a decrease in snowpack as a result of increasing temperatures (Mote et al., 2005). In watersheds where glaciers are present, it is expected that increased temperature shifts will result in decreased summertime streamflows as peak flows shift earlier into the spring (Stahl et al., 2008). Spring flows have been observed to occur earlier in snowmelt dominated regions with increases in temperature (Cayan, Dettinger, Kammerdiener, Caprio, & Peterson, 2001; G.A Hodgkins, Dudley, & Huntington, 2003; Thomas G. Huntington, Richardson, McGuire, & Hayhoe, 2009; Stewart, Cayan, & Dettinger, 2005). However, some have found that snow cover may be increasing due to certain large scale atmospheric processes (Cohen et al., 2012). Many have found that snow cover and accumulation trends are regionally dependent (Stewart, 2009; Ye & Mather, 1997). In general, many regions will expect decreases in snow pack, extent and duration with increases in temperature (Stewart, 2009). However, some areas that are experiencing concurrent precipitation and temperature increases may experience greater snow pack depths and extents but experience earlier high melt events (Stewart, 2009). This may lead to increasing flood risk in some areas.
The intensity of a precipitation event (Hancock et al., 2010) and the characteristics of a watershed (Del Giudice et al., 2014; Emerson et al., 2005; Wang & Yu, 2012) are known as primary factors in the generation of flood conditions. It is known that extreme flooding events are related to very recent precipitation, and antecedent conditions. Many studies have shown the importance of antecedent moisture conditions on the production of runoff from a given precipitation event (De Michele & Salvadori, 2002; Descroix et al., 2002; James & Roulet, 2007; Meyles et al., 2003; Nishat et al., 2010; Penna et al., 2011; Radatz et al., 2013; Sahu et al., 2007). Pre-event surface storage has also been shown to be a factor in the genesis of high streamflows (Ayalew et al., 2013; Hancock et al., 2010; Montaldo et al., 2004). The response of a watershed has been shown to be drastically different during wet and dry conditions, where wet conditions are characterized by horizontal flow (or runoff) and dry conditions are characterized by vertical flow (or infiltration) (Grayson et al., 1997; Meyles et al., 2003). Meyles et al. (2003) describes how water that exfiltrates to the surface in dry conditions is absorbed downslope by drier soils. Interestingly, the soil moisture threshold that separates wet and dry watershed response has been reported between 49% and 90% (Grayson et al., 1997; James & Roulet, 2007; Meyles et al., 2003; Penna et al., 2011; Radatz et al., 2013). Also, hysteresis has a strong impact on drainage in soils, where the behavior of drainage when wetting is much different than when drying (Dane & Wierenga, 1975). In the studies above that display the importance of antecedent conditions, it is important to investigate shifts in the persistence of precipitation events as an important driver in flood genesis.
Thus, with shifts in large scale atmospheric processes associated with persistent systems, it is expected that there will continue to be nonlinear increases in severe flooding frequencies and magnitudes. Although it is thought that changes in precipitation and temperature contribute to changes in antecedent conditions, which in turn will have different relative impacts on flood genesis, the role of internal variables in connecting floods to climate drivers has not yet been studied in detail.

New England has experienced changes in hydro-climatic processes associated with temperature changes. Huntington et al. (2004) found the ratio of snow to rain precipitation to be decreasing over the majority of New England. Hodgkins et al. (2002) found very significant trends (p<0.0001) in earlier ice-out dates for all five lakes in New England whose records were longer than 150 years. Consistent with less precipitation occurring as snow and earlier ice-outs, the center of volume of springtime flows was found to occur earlier at all gauge stations heavily impacted by snowmelt (Hodgkins et al., 2003). Hodgkins et al., (2011) also showed summer low-flow to be increasing for the majority of New England.
Chapter 3: Downscaled Climate Projections over the Lake Champlain Basin, Vermont, USA

IMPACTS OF PROJECTED CLIMATE CHANGE OVER THE LAKE CHAMPLAIN BASIN IN
VERMONT

Justin Guilbert1*
School of Engineering
University of Vermont
Burlington, VT 05401,

Brian Beckage
Department of Plant Biology
University of Vermont
Burlington, VT 05405,

Jonathan M. Winter
Center for Climate Systems Research
The Earth Institute
Columbia University
New York, NY 10025,

Radley M. Horton
Columbia University
NASA Goddard Institute for Space Studies
New York, NY 10025

Timothy Perkins
Department of Plant Biology
University of Vermont
Burlington, VT 05404

and

Arne Bomblies
School of Engineering
University of Vermont
Burlington, VT 05405.

Corresponding Author:
Justin Guilbert
23 Mansfield Ave., Burlington, VT 05401
jguilber@uvm.edu
Abstract

The Lake Champlain Basin is a critical ecological and socioeconomic resource of the Northeastern US and Southern Quebec. While general circulation models (GCMs) provide an overview of climate change in the region, they lack the spatial and temporal resolution necessary to fully anticipate the effects of rising global temperatures associated with increasing greenhouse gas concentrations. Observed trends in precipitation and temperature were assessed across the Lake Champlain Basin to bridge the gap between global climate change and local impacts. Future shifts in precipitation and temperature were evaluated as well as derived indices, including maple syrup production, days above 32.2 degrees Celsius (90 degrees Fahrenheit), and snowfall, relevant to managing the natural and human environments in the region. Four statistically downscaled, bias-corrected GCM simulations were evaluated from the Coupled Model Intercomparison Project Phase 5 (CMIP5) forced by two Representative Concentrations Pathways (RCPs) to sample the uncertainty in future climate simulations. Precipitation is projected to increase by between 9.1 and 12.8 mm per year per decade during the 21st century while daily temperatures are projected to increase between 0.43 and 0.49 degrees Celsius per decade. Annual snowfall at six major ski resorts in the region is projected to decrease between 46.9% and 52.4% by late-century. In the month of July, the number of days above 32.2 degrees Celsius in Burlington, Vermont is projected to increase by over 10 days during the 21st century.

Keywords: Climate Change, Climate Change Impacts, Statistical Downscaling, Northeastern United States, Lake Champlain Basin, New England
3.1 Introduction

The Lake Champlain Basin is a 21,326 km$^2$ watershed on the United States-Canada border that spans Vermont, New York, and Quebec. Economic activities associated with the watershed, including tourism, agriculture, and recreational use, are extremely important to the region, and sensitive to climate change. The region has recently been impacted by extreme weather events including significant flooding in 2011 from both heavy spring rainfall and Tropical Storm Irene in late-summer. These precipitation events resulted in extensive damage to public infrastructure and private property, and highlight the potential impacts of climate change (Pealer, 2012). Adaptation planning can be enhanced by reliable climate projections at local governance scales.

Regional changes in temperature and precipitation have already been observed in the region. Keim et al. (2005) found positive trends in precipitation across the NE US but did note that trends differ across subsets of this region. Groisman et al. (2005) found spatially divergent precipitation trends within the Lake Champlain Basin, a sub-region of the NE US, with linear trends between 1900-2002 of approximately +10% in the northern section but -10% in the southern part of the watershed. The magnitude of historical precipitation change in the Lake Champlain Basin increased by 8% to 38% at low and high elevations, respectively, over the 40 year period from 1963 to 2003 (Beckage et al., 2008).

Several studies have examined climate change projections across the NE US at broad regional scales. Hayhoe et al. (2007) used 9 general circulation models (GCMs) and 2 emissions scenarios from the IPCC SRES (Intergovernmental Panel on Climate Change Special Report on Emissions Scenarios) to project a 4.5°C and 2.9°C mean annual
temperature increase under the A2 and B1 emissions scenarios, respectively, by the end of the century (2070-2099) in the NE US. Ahmed et al. (2012) used six GCMs and four regional climate models (RCMs) that were statistically downscaled and bias-corrected to project that the annual number of frost days would be reduced by 25 days and the growing season length would increase by 20 days by mid-century (2046-2065) in the NE US. This study found that maximum 5-day precipitation increased in some areas of the Lake Champlain Basin while decreasing in others. Rawlins et al. (2012) used 9 RCM simulations forced with the SRES A2 emissions scenario from the North American Regional Climate Change Assessment Program (NARCCAP) to compare historical and mid- to late-century (2041-2070) NE US temperature and precipitation. The average annual increase in temperature and precipitation across this time period was 2.6° C and 6%, respectively.

Regional climate change impacts are potentially highly consequential for a number of industries, including maple syrup production, dairy farming, and winter recreation (e.g. skiing and snowmobiling; Horton et al. 2014). Moreover, state- and local-level policy makers increasingly consider climate data for future planning (Horton et al., 2014). For example, water resources, transportation infrastructure, and public health adaptations may rely heavily on predictions of expected future climate. This study derives projected climate metrics, detailed below, that can be used to inform decisions related to agriculture, tourism, and human health.

3.2 Methods
The range of projected climate change for the Lake Champlain Basin in Vermont, USA, was examined using statistically downscaled GCM projections. Four GCMs and two emission scenarios were utilized from the recent Coupled Model Intercomparison Project Phase 5 (CMIP5) model runs. Simulated future changes in climate were assessed with respect to historical precipitation and temperature from National Climatic Data Center (NCDC; Menne et al., 2012) stations. Projected changes in stakeholder-relevant climate metrics were derived. Metrics related to agriculture and dairy include a heat index, an aridity index, and growing season length. Human health related metrics include heating/cooling requirements, a heat index and days above 32.2°C (90°F). Days suitable for maple syrup production, freezing days and snowfall metrics were developed in relation to the maple syrup and ski industry.

3.2.1 Study region

The Lake Champlain Basin is primarily comprised of forested and agricultural land. A network of rivers and streams is distributed throughout the region, and the north to south oriented Green Mountains form the major topographic feature of the study region. Elevations in the Lake Champlain Basin range from 30 meters above sea level (masl) on Lake Champlain to 1,340 masl at the top of Mount Mansfield, the highest point in Vermont, over a distance of less than 50 km. (Figure 3.1)

3.2.2 Climate Data

Historical meteorological station data from throughout the study region was acquired from the Global Historical Climatology Network (GHCN; Menne et al., 2012). These
Data are at a daily resolution and consists of three variables: precipitation, maximum temperature and minimum temperature. The full record for 132 stations was downloaded. The periods of record were 1875-present and 1884-present for precipitation and temperature, respectively.

Data for future projections of temperature and precipitation were from four realizations of a CMIP5 multi-model ensemble which was downscaled using a method of Bias-Correction with Constructed Analogues (BCCA; Brekke, et al., 2013). BCCA was chosen for its high spatial (1/8°) and temporal (daily) resolution. BCCA methods leverage observed climate data to both bias-correct and statistically downscale GCM data. A full description of BCCA methodology can be found in Hidalgo et al. (2008) and (Maurer & Hidalgo, 2008; Maurer et al., 2010). Relative to Coupled Model Intercomparison Project Phase 3 (CMIP3), CMIP5 models generally have higher spatial resolution and contain a number of improvements that have led to better simulation of some climate features (Knutti & Sedlávcek, 2012; Stroeve et al., 2012). CMIP5 models are driven by Representative Concentration Pathways (RCPs; Moss et al., 2010), which are a set of greenhouse gas emission, aerosol, and land use changes scenarios developed as an input for climate modeling experiments. CMIP5 simulations use four different RCPs distinguished by their radiative forcing at the end of the century in W·m⁻², ranging from 2.6 to 8.5, with higher numbers denoting greater greenhouse gas emissions and radiative forcing. Additional details on the construction of RCPs can be found in Moss et al. (2010). In this study, RCPs 4.5 and 8.5 were used to capture moderate and high climate change scenarios for purposes of adaptive planning.
Daily minimum temperature, daily maximum temperature, and daily precipitation were analyzed across three time periods: 1961-2000, 2040-2069, and 2070-2099. All three time periods were assessed using four GCMs each forced by two RCPs. GCMs were selected to capture the range of potential future temperature and precipitation based on a preliminary analysis of the entire BCCA ensemble of simulations under RCP 8.5 for the 2070-2099 period. Future climate simulations (2070-2099) were ranked from wettest to driest and warmest to coolest, with the top and bottom 10% of the ranked GCMs considered for selection. Priority among GCMs ranking in the top and bottom 10% of temperature and precipitation was given to well-established models that were available for both CMIP3 and CMIP5, maximizing opportunities for comparison of results to existing and future studies. After this process the selected GCMs were 1) the Commonwealth Scientific and Industrial Research Organization csiro-mk3-6-0.1 (wettest 10%), 2) Institut Pierre-Simon Laplace ipsl-cm5a-mr.1 (driest 10%), 3) Japan Agency for Marine-Earth Science and Technology, Atmosphere and Ocean Research Institute at the University of Tokyo, and National Institute for Environmental Studies miroc-esm.1 (warmest 10%), and 4) Institute for Numerical Mathematics inmcm4.1 (coolest 10%).

Historical climate records and future climate projections for temperature and precipitation were compiled for a portion of the Lake Champlain Basin. This area spans from 72°11'15"W to 73°18'45"W and from 43°56'15"N to 45°26'15"N (Figure 3.1). The rectangular area is covered by a 10 by 13 grid of 1/8° degree cells. The outer dimensions of this area are approximately 100 km by 180 km. At the latitude of the study region 1/8° degree cells are approximately 10 km by 14 km.
3.2.3 Climate Data Analysis

3.2.3.1 Historical Trends

For each GHCN station all missing values were removed, and minimum and maximum daily temperatures were averaged to calculate daily mean temperature. Daily temperature and precipitation data were averaged across all available stations for each day by a simple arithmetic mean. Annual means and quantiles for temperature and precipitation were calculated from the distribution of average daily values for a given year. Only the years in which data from 15 or more stations were available were used. These periods were 1941-2012 for precipitation and 1958-2012 for temperature.

3.2.3.2 Deltas and Climate Projections

Daily climate data were averaged to produce a seasonal cycle (i.e., composed of mean monthly values) for every grid cell within the study region. This was repeated for each BCCA scenario using the same 4 GCMs and 2 RCPs for both temperature and precipitation across all three time periods. Temperature change factors or ‘deltas’ were calculated for each grid cell by subtracting the seasonal cycle in the historical period from the respective seasonal cycle in each of the two future periods. Each value represents the projected temperature change in degrees Celsius for a particular cell and month for a given future period. Projections of change in precipitation were calculated in a similar manner except that ratios of change (‘delta factors’) were used. Temperature deltas were added to the observed daily data (Maurer et al., 2010) for the study region for every cell to provide a future temperature projection in each grid cell. The delta factors for
precipitation were multiplied by the observed daily precipitation data for every cell to provide a projection for future precipitation. The delta method was applied to examine projected changes in mean temperature and precipitation as well as extremes of temperature (5th and 95th quantiles) and precipitation (95th and 99th quantiles) to explore changes in extremes.

3.2.3.3 Ratio of Precipitation to Potential Evapotranspiration

In order to gauge potential future impacts on vegetation and agriculture, the ratio of precipitation to potential evapotranspiration (rPPET) was calculated. Potential evapotranspiration (PET) was calculated using Thornthwaite’s equation (Thornthwaite, 1948)(1), which depends upon average daily and monthly temperature (Ta and Tαi), average day length by month (L) and number of days per month (N).

\[
PET = 16 \frac{L}{12} \frac{N}{30} \left( \frac{10T_{\alpha}}{I} \right)^\alpha
\]

Equation 3.1

where \[ \alpha = \left(6.75 \times 10^{-7}\right)I^3 - \left(7.71 \times 10^{-5}\right)I^2 + \left(1.792 \times 10^2\right)I + 0.4923 \]  

Equation 3.1.1

and \[ I = \sum_{i=1}^{12} \left( \frac{T_{\alpha i}}{5} \right)^{1.514} \]

Equation 3.1.2

3.2.3.4 Days Below Freezing, Days Above 32.2°C, Snowfall, Growing season, and Days with Maple Syrup Production

The number of freezing days per month was determined by counting the number of days per month for which the average temperature was below 0°C. These data were used to calculate the mean number of freezing days per month and year for each time period. This method was also used to calculate the number of days where the maximum daily
temperature was above 32.2°C (90°F) per month. This was calculated for the grid cell within the region that contained Burlington, VT to assess the heat wave impacts of climate change on a population center.

A metric for snowfall was calculated for four grid cells which were determined to contain six of the region’s largest ski resorts. This calculation was based on days with an average temperature below freezing combined with projections of daily precipitation. If a particular day was below freezing, a table was used from the NCDC (NCDC, 2013) to determine the ratio of snowfall to total precipitation. This table contains values ranging from 10 to 100, depending on the temperature. This ratio was then multiplied by the spatially averaged precipitation across the four selected cells for the respective day to convert precipitation to snowfall. The values calculated through this method were averaged over each 30-year period to determine monthly projected snowfall. Monthly projected snowfall was summed from October through April to determine the projected cumulative seasonal snowfall.

The length of growing season was determined by finding the maximum time between two days whose minimum temperatures were below 0°C for each of the years within each 30 year period.

Days in which sugar maple trees are likely to exude sap suitable for maple syrup production were determined based on daily minimum and maximum temperatures. Days with a minimum temperature below -1.1°C (30°F) and maximum temperature above
2.2°C (36°F) were considered suitable for maple syrup production (Skinner et al., 2010). The number of suitable days was calculated for every month for the 30 year period.

3.2.3.5 Heating and Cooling Requirements and Heat Index

Heating and cooling requirements were determined based on daily average temperatures across the region. Days with temperatures below 20°C (68°F) were deemed to require heating and days with temperatures above 25.6°C (78°F) were deemed to require cooling based on a human comfort index (Burroughs & Hansen, 2004). To quantify the amount of heating and cooling required, the number of degrees per day above or below the cooling and heating limits were summed monthly and annually to create heating and cooling requirements in degree days (DD).

A heat index was calculated in DD. The heat index calculations utilized the same formula as NOAA, which uses temperature and relative humidity. The heat index, as per NOAA, is calculated for days whose maximum daily temperature is greater than 26.7°C (80°F) before calculation. Daily dew point temperatures were calculated using an empirical model, which requires average minimum and maximum daily temperature (Kimball et al., 1997). Relative humidity was calculated from the resulting dew point temperature and average daily temperature (Lawrence, 2005).

3.2.4 Bayesian statistics

Bayesian credible intervals were computed for the delta factors of temperature, precipitation, temperature quantiles, precipitation quantiles, and derived metrics for future periods (Lavine, 2009). Thirty monthly replicates exist for each of the eight RCP-
by-GCM combinations as well as the 30 observations for each month during the base period. It was assumed that these replicates for both the future and historic periods were normally distributed with different but unknown means and variances. The posterior distribution of the difference (e.g., for temperature) or ratio (e.g., for precipitation) of these population means were estimated and reported at the 95% credible interval on these metrics. Credible intervals were calculated using Just Another Gibbs Sampler (JAGS; Plummer, 2003), which generates samples from the posterior distribution using Markov Chain Monte Carlo simulation.

3.3 Results

3.3.1 Observed Trends

Averaged GHCN daily data across the region indicated both an increase in temperature and precipitation between 1958-2012 and 1941-2012, respectively (Figure 3.2). Average temperature increased by 0.19°C per decade, which is nearly twice the rate of 1.02°C per 100 years found by Trombulak and Wolfson (2004) for the period of 1903-2000. Annual precipitation increased by 45.8 mm per decade, which is higher than other studies that include this region (Hayhoe et al., 2007; Keim et al., 2005). Keim et al. (2005) found increases in annual precipitation between 8 and 39 mm per decade across Vermont, while Hayhoe et al. (2007) estimated increases in annual precipitation of approximately 7 mm per decade. During the time period of 1941-2012, the number of operating weather stations increased by 106% for temperature and 230% for precipitation. The 0.05 and 0.95 quantiles of temperature increase by 0.50°C and 0.05°C per decade, respectively. The 0.95 and 0.99 quantiles of precipitation increase by 0.38mm and 0.48mm per day per
decade, respectively. The record precipitation year (2011) totaled 1.46 meters of precipitation, which is 2.47 standard deviations above the period mean of 1.08 meters of precipitation from 1948-2012. Nine of the ten warmest years in the temperature record have occurred between 1990 and 2012, with the warmest year on record occurring in 2012. Trends in temperature and precipitation could be affected by changes in the spatial and topographic distributions of climate stations with time. To support the trends found for the region, Burlington, Vermont, one the longest and most reliable standing records of weather in the region was used. Burlington’s weather station supported the regional findings, showing positive trends in both temperature and precipitation (Figure 3.2).

3.3.2 Climate Projections

Temperature and precipitation projections are qualitatively consistent with the increases seen in the historical station data. Annual average temperature is projected to increase by 3.1°C by mid-century and 4.6°C by late-century, which is consistent with Hayhoe et al. (2007) estimate of between +2.1°C and +2.9°C for mid-century and between +2.9°C and +5.3°C for late-century. Similarly, all monthly estimates of temperature change were significantly different from zero using a 95% Bayesian credible interval (Figure 3.3). The 0.05 quantile of daily temperature is projected to increase 3.5°C by mid-century and 5.3°C by late-century, while the 0.95 quantile of daily temperature is projected to increase 2.8°C by mid-century and 4.0°C by late-century (Figure 3.3). (Table 3.1)

Average daily precipitation is projected to increase by 7.1% by mid-century, and 9.9% by late-century (Figure 3.4), which is in agreement with Hayhoe et al. (2007) who estimated
an increase of between 5% and 8% by mid-century and between 7% and 14% by late-century. Eight of the 12 calendar-month precipitation delta factors are significantly different from 1.0 (no predicted change) for both mid-century and late-century. The 0.95 quantile of daily precipitation is projected to increase by 8.9% by mid-century and 12.5% by late-century, while the 0.99 quantile of daily precipitation is projected to increase by 11.9% by mid-century and 16.7% by late-century (Figure 3.4). (Table 3.1)

The number of below freezing days annually is projected to decrease 29% and 39% by mid- and late-century, respectively, from an average of 116.5 days during the base period (Figure 3.5). All months in the two future periods show decreases in snowfall in comparison to the base period (Figure 3.6). Cumulative annual snowfall projections near major Vermont ski resorts show decreases from the base period average of 676 cm. Mid- and late-century projections estimate cumulative snowfall to decrease by 36% and 50%, respectively (Table 3.2).

In the base period, 5.5 days per year were warmer than 32.2°C (90°F) in the cell containing Burlington, VT. This is projected to increase by 310% (+18.6 days) by mid-century and 530% (+31.9 days) by late-century (Figure 3.7). The heat index calculated for the base period was 130 DD. This value is projected to increase to 475DD (+265%) by mid-century and to 754DD (+480%) by late-century (Figure 3.8). The length of the growing season during the base period was 140.5 days, which is projected to increase 20% by mid-century and 31% by late-century (Figure 3.9). Days suitable for maple syrup production are projected to decrease and shift in annual cycle. The two peaks found in the base period in fall and spring both decrease while shifting toward mid-winter. Annually
the number of suitable days decreases from the base period with 60.3 days to 53 and 49 by mid- and late-century. Interestingly, this decrease in annually suitable days occurs while the months of December and January see a net increase in suitable days of 4.3 and 5.4 by mid- and late-century (Figure 3.10). (Table 3.2)

Heating and cooling requirements in the study region were calculated in DD. During the 21st century heating requirements are projected to decrease by 19% and 27% annually by mid- and late-century, respectively, while cooling requirements are expected to increase by 13% and 40% over the same period (Figure 3.11). (Table 3.2)

April through October were analyzed for statistical differences in rPPET between the base period, mid-century and late-century. By mid-century, 4 of the 7 months show significant decreases in the rPPET compared to the base period, and by late-century this increases to 5 of 7 (Figure 3.12). To decrease rPPET either precipitation must decrease or temperature must increase. For the late-century, only fall precipitation is not projected to increase significantly. The range of possible precipitation change is between -0.5% and +11.7% by late-century. Spring, summer and fall temperatures are projected to increase by approximately 4°C. This indicates that any projected decrease in rPPET is likely being driven by temperature increases. (Table 3.2)

3.4 Discussion

The Lake Champlain Basin is likely to experience significant climate change during the 21st century. Temperatures are expected to increase by 0.46°C per decade through late-century, which is more than double the rate of increase, +0.19°C per decade, in the
historic station data. Precipitation is expected to increase 9.9% (+108mm per year) by late-century, less than one quarter the rate found in the historic station data. Because of large multi-decadal variability, even the long term observed precipitation trends may be strongly influenced by unpredictable natural variability. Hayhoe et al. (2007) predicts an increase in temperature of between 2.9°C and 5.3°C, and an increase in precipitation of between 7% and 14% for late-century. NECIA (2006) predicts similar increases by 2100 with projections between approximately 2°C and 7°C for temperature and approximately 10% for precipitation.

The changes in temperature and precipitation are expected to result in an increase in the growing season aridity (as measured by rPPET). By late-century, July’s rPPET is projected to decrease below 1.0, meaning potential evapotranspiration exceeds precipitation. This is in agreement with Hayhoe et al. (2007) and NECIA (2006) who both expect an increase in short duration droughts. A change in rPPET could have ecological and agricultural impacts such as increased water stress on forests, greater susceptibility of trees to insects, decreased crop productivity, and need for irrigation of agricultural areas (Lindner et al., 2010; Maracchi et al., 2005).

By late-century both annual snowfall and the number of days below 0°C are expected to decrease by 50% and 45 days, respectively. This projection of decreased snowpack is lower in magnitude than the findings of Hayhoe et al. (2007), however this may be explained by this paper’s use of only high elevation grid cells in a more northern climate. The ski industry is likely to be impacted by the decrease in snowfall and the decline in number of freezing days, particularly because temperature increases are most pronounced.
in the winter months. With less natural snow, the need for snowmaking is expected to increase while the conditions conducive to snowmaking are projected to diminish. However, finer resolution analyses of climate change would be necessary to better project changes in snowfall, particularly in higher elevation locations that are not well resolved by current climate models. The maple syrup industry is also temperature sensitive, depending on cycles of freezing nights below 1.1°C followed by above freezing days greater than 2.2°C for sap to flow in trees (Skinner et al. 2010).

One potential benefit of regional warming is a longer growing season. Projections indicate that the growing season is expected to lengthen (+43 days) by late-century. This increase in growing season length agrees with Hayhoe et al. (2007) who found an increase of between 29 and 43 days by late century. With a longer, warmer growing season a larger variety and volume of food may be able to be produced. However, some of the benefit of a longer growing season could be offset by increasing aridity and agriculture pest pressure as temperatures warm. A longer ‘warm’ season and shortened ‘cold’ season is also expected to shift the seasonal energy requirements of the region with less heating yet more cooling required. Late-century projections predict the number of days above 32.2°C to be between 35 and 40 days while the annual heat index is expected to increase to between approximately 700 and 800 DD. Projected heat index values in July for late-century will make the average day fell approximately 7°C(13°F) warmer. Projections of both days above 32.2°C and heat index largely agree with the findings of the Northeast Climate Change Assessment (NECIA, 2006). For industries, such as the dairy industry, where animal comfort can be related to production, an increase in the
severity of daily heat indices could negatively impact profits (Rosenzweig & New York State Energy Research and Development Authority, 2011).

Extreme temperature and precipitation threshold events are likely to increase. By late-century, the 95th percentile of daily maximum temperature is projected to increase by 4.0°C while the 5th percentile is projected to increase by 5.3°C. These increases in temperature quantiles could increase the occurrence of dangerous heat waves. By late-century, the 99th percentile of daily precipitation is projected to increase by 3.4 cm which is equivalent to an additional 34,000 m³ of liquid precipitation per km². This increased flow could overwhelm current infrastructure including bridges and culverts as well as increase nutrient loading to Lake Champlain through overland flow and stream bank erosion.

Increasing greenhouse gas concentrations in the atmosphere is expected to impact climate across the Lake Champlain Basin with significant hydrologic and ecological consequences. For example, increases in temperature could open the region to biological invasion (Ammunet et al., 2012; Bellard et al., 2013), while shifting the ranges of northern species of trees (McNulty & Aber, 2001). Predicting the impacts of climate change is essential for adaptation throughout the Lake Champlain Basin. Results from analyses will inform hydrological, ecological, and lake models with the VT EPSCoR project, as well as provide the basis for adaptation planning in the region. It is noted that this study has focused on quantiles and means of precipitation but aspects of climate change must also be carefully considered including tropical and post-tropical cyclones such as Hurricane Irene, and the potential for changes in mid-latitude climate due to
factors such as reductions in Arctic sea ice (Liu et al., 2012; 2013), which are both poorly simulated by GCMs. Future work includes finer-scale downscaling in this topographically-diverse region; for example, using statistical lapse-rate relationships and regional climate model simulations to explore the future climate of the high elevation portions of the LCB domain and by better resolving processes influencing regional climate in the focal region.

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References


Brekke, L., B. Thrasher, E. Maurer, and T. Pruitt, 2013: Downscaled CMIP3 and CMIP5 Climate Projections: Release of Downscaled CMIP5 Climate Projections,


Figure 3.1 Study region including northern Vermont and southern Quebec, with Lake Champlain and the Champlain Valley bordering the western edge of the study region and the Green Mountains across the eastern portion of the study region. 1/8° degree cells are shown to display the resolution of the Bias-Correction Constructed Analogues data.
Figure 3.2 Historic trends in annual A) temperature and B) precipitation from weather stations across the study region. The dotted black lines indicate the number of weather stations that are reporting temperature or precipitation values. 1958 and 1941 were selected as the beginning of the historical period for temperature and precipitation, respectively, because these were the first years with 15 or more weather stations reporting temperature or precipitation. The solid black line is the linear regression line showing the increase in temperature from 1958 to 2012, with the slope indicating an increase of 0.19°C per decade and 0.12 mm per day per decade.
Figure 3.3 Projected increase in A) mean, B) the 0.05 quantile and C) the 0.95 quantile for 2040-2069 and 2070-2099 for RCP 4.5 and 8.5 compared to the 1970-1999 base period and four GCMs. The 0.05 quantile is representative of the coldest days of a month and the 0.95 quantile is representative of the warmest days of a month. The furthest right column shows the average annual temperature change for mean, 0.05 quantile and 0.95 quantile. The summer months, June through August, are projected to generally warm less than months in other seasons.
Figure 3.4 Projected changes in A) mean, B) the 0.95 quantile and C) the 0.99 quantile of daily precipitation for 2040-2069 and 2070-2099 for four GCMs under RCP 4.5 and 8.5 compared to the 1970-1999 base period. Precipitation delta factors above 1.0 indicate that the intensity of daily precipitation is expected to increase. There is large uncertainty in the precipitation projections: the change in mean precipitation for July in the period 2070-2099, for example, varies from a ~25% increase to a ~22% decrease in daily rainfall. The choice of GCM often has a greater influence than the climate scenario, for example, in October where both the minimum and maximum change in mean precipitation occurs in the same scenario run by two different GCMs. The two quantiles
shown are representative of the upper tail of a daily precipitation distribution. Nearly all delta factors averaged across the year show an increase in large (0.95) daily precipitation events for both future periods while all delta factors averaged across the year for the largest (0.99) daily precipitation events are above 1.0.

**Figure 3.5** The projected number of freezing days per month for A) 2040-2069 and B) 2070-2099 for RCP 4.5 and 8.5 compared to the 1970-1999 base period. Circular points represent individual GCM runs while the horizontal bars represent the mean of the GCM runs for each RCP. The vertical dashed lines are visual connectors between the individual runs and averages within a month. Freezing days were defined as days with an average temperature below 0°C. The number of mid-winter (December through February) thaws in the study area will increase as the number of freezing days decreases.
Figure 3.6 The projected monthly snowfall for A) 2040-2069 and B) 2070-2099 for RCP 4.5 and 8.5 compared to the 1970-1999 base period. Circular points represent individual GCM runs while the horizontal bars represent the mean of the GCM runs for each RCP. The vertical dashed lines are visual connectors between the individual runs and averages within a month. Monthly snowfall shows a decreasing trend with time across both emissions scenarios. All months show decreases in snowfall for both time periods and all scenarios. By late-century, a 42 percent decrease in January snowfall is projected in comparison to the observational period.
Figure 3.7 The projected number of ‘hot’ days above 32.2°C (90°F) for 2040-2069 and 2070-2099 for RCP 4.5 and 8.5 compared to the 1970-1999 base period. Circular points represent individual GCM runs while the horizontal bars represent the mean of the GCM runs for each RCP. The vertical dashed lines are visual connectors between the individual runs and averages within a month. The number of hot days per year increases by 18.6 and 31.9 days by 2040-2069 and 2070-2099 with increases in July accounting for 7.3 and 11.2 days per year, respectively.
Figure 3.8 Projected heat index for A) 2040-2069 and B) 2070-2099 for RCP 4.5 and 8.5 compared to the 1970-1999 base period. Circular points represent individual GCM runs while the horizontal bars represent the mean of the GCM runs for each RCP. The vertical dashed lines are visual connectors between the individual runs and averages within a month. In degree days the annual heat index is expected to increase by over 600 and 1100 degree days by mid- and late-century, respectively.
**Figure 3.9** The projected growing season length for 2040-2069 and 2070-2099 for RCP 4.5 and 8.5 compared to the 1970-1999 base period. Circular points represent individual GCM runs while the horizontal bars represent the mean of the GCM runs for each RCP. The vertical dashed lines are visual connectors between the individual runs and averages within a month. The growing season length increases by 23.4 and 32.5 days by 2040-2069 and 2070-2099 under RCP 4.5 and by 32.8 and 53.6 days under RCP 8.5.
Figure 3.10 Projected days with maple syrup production for A) 2040-2069 and B) 2070-2099 for RCP 4.5 and 8.5 compared to the 1970-1999 base period. Circular points represent individual GCM runs while the horizontal bars represent the mean of the GCM runs for each RCP. The vertical dashed lines are visual connectors between the individual runs and averages within a month. Days with maple syrup production are projected to shift from 2 distinct peaks in fall and spring towards one peak in mid-winter. The number of days suitable for maple syrup production is expected to decrease by 7.3 and 11.5 days per year by mid- and late-century.
Figure 3.11 Projected heating and cooling requirements for A) 2040-2069 and B) 2070-2099 for RCP 4.5 and 8.5 compared to the 1970-1999 base period. Circular points represent individual GCM runs while the horizontal bars represent the mean of the GCM runs for each RCP. The vertical dashed lines are visual connectors between the individual runs and averages within a month. The degree days of cooling are expected to increase by 12.7 and 40.0 while the degree days of heating are projected to decrease by 987 and 1409 by mid- and late-century, respectively.
Figure 3.12 Projected rPPET for A) 2040-2069 and B) 2070-2099 for RCP 4.5 and 8.5 compared to the 1970-1999 base period. Circular points represent individual GCM runs while the horizontal bars represent the mean of the GCM runs for each RCP. The vertical dashed lines are visual connectors between the individual runs and averages within a month. Projected future rPPET usually brackets historic rPPET indicating uncertainty in rPPET projections.
Table 3.1 In comparison to the base period, the change in the mean and quantiles of temperature and precipitation are shown for mid- and late-century.

<table>
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### Table 3.2 Relevant climate change metrics for the base period and mid- and late-century.

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<td>1.10</td>
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Chapter 4: Characterization of increased persistence and intensity of precipitation in the northeastern United States

Characterization of increased persistence and intensity of precipitation in the northeastern United States

Justin Guilbert\textsuperscript{1}
School of Engineering
University of Vermont
Burlington, VT 05405,

Alan K. Betts
Atmospheric Research,
Pittsford, VT 05763,

Donna M. Rizzo
School of Engineering
University of Vermont
Burlington, VT 05405,

Brian Beckage
Department of Plant Biology
University of Vermont
Burlington, VT 05405,

and

Arne Bomblies
School of Engineering
University of Vermont
Burlington, VT 05405.

\textsuperscript{1} Corresponding author: jguilber@uvm.edu
Main point #1: Precipitation in the northeastern United States is becoming more persistent

Main point #2: Precipitation in the northeastern United States is becoming more intense

Main point #3: Observed trends constitute an important hydrological impact of climate change

Abstract

We present evidence of increasing persistence in daily precipitation in the northeastern United States that suggests global circulation changes are affecting regional precipitation patterns. Meteorological data from 222 stations in 10 northeastern states are analyzed using Markov Chain parameter estimates to demonstrate that a significant mode of precipitation variability is the persistence of precipitation events. We find that the largest region-wide trend in wet persistence (i.e., the probability of precipitation one day, given precipitation the preceding day) occurs in June (+0.9 percent probability per decade over all stations). We also find that the study region is experiencing an increase in the magnitude of high intensity precipitation events. The largest increases in the 95th percentile of daily precipitation occurred in April with a trend of +0.7 mm per day per decade. We discuss the implications of the observed precipitation signals for watershed hydrology and flood risk.

Index Terms: Regional climate change, Climate variability, Hydrology, Climate impacts, Extreme events

4.1 Introduction
Concurrent with the global increase of temperature is a change in precipitation, which varies widely in magnitude and direction depending on the region considered. In general, dry areas have become drier and wet areas have become wetter (Dore, 2005). Warming temperatures increase the potential intensity of precipitation, as saturation vapor pressure increases steeply with temperature (Berg et al., 2013; Durack et al., 2012). Changing global circulation patterns may also have pronounced local impacts on the distribution of precipitation, influencing watershed hydrology as well as human and natural systems. However, spatial and temporal variability in precipitation is very high, and for many regions, including the northeastern United States (NE US), the connection of local-scale precipitation changes to global climate change remains elusive.

Recent research on global circulation changes suggests that arctic amplification and sea surface temperatures are drivers of changes in jet stream wave amplitude and propagation speed (eg. Francis & Vavrus, 2012; Petoukhov et al., 2013; Screen & Simmonds, 2013; Tang et al., 2013). One hypothesis (Francis & Vavrus, 2012) is that changing meridional temperature differences reduce jet stream intensity, resulting in higher amplitude waves and slower velocities, both of which can affect storm tracks and resulting local weather impacts. However, the proposed role of arctic amplification in regulating weather patterns resulting from jet stream meanders has been criticized (Kintisch, 2014). Other hypotheses suggest that changing sea surface temperature (Muller, 2013; Palmer, 2014) plays a similar role. Palmer (2014) proposes a mechanism that links increased sea surface temperatures (SSTs) to larger amplitude planetary waves. In this mechanism, increased SSTs generate more powerful storms in the western tropical Pacific, and the release of
latent energy excites propagating wave trains that interact with and amplify the mid-latitude planetary waves. Muller (2013) suggests that warming SSTs may also contribute to the organization of squall lines in convective systems that can lead to increases in extreme precipitation.

The NE US has experienced an increase in precipitation of approximately 10 mm per decade and the greatest increases in extreme precipitation in the United States (Horton, et al., 2014). For example, the return period of daily rainfall intensity greater than 101.6mm (4 inches) has decreased in the last century from 26 to 11 years in the NE US, and the frequency of the upper 10 percent of rainy days has increased in the NE US (Groisman et al., 2005b, 2001). Under the recently proposed mechanisms that yield slower-moving planetary waves, storms are expected to propagate more slowly resulting in more persistent weather patterns. Changes in the persistence of precipitation in the NE US have not been studied in detail. However, NE US precipitation magnitudes show little dependence on large-scale climate variability (Brown et al., 2010; Dai, 2013). Brown et al. (2010) considered six teleconnection patterns, while Dai (2013) looked only at the inter-decadal Pacific oscillation.

Understanding the nature of precipitation variability in the NE US is critical especially with respect to severe flooding, which has become more frequent with time in this region (M. J. Collins, 2009). In this study, we provide a statistical analysis of regional trends in
the median and 95\textsuperscript{th} percentile of daily precipitation, and trends in wet and dry persistence. We focus on these metrics because as global temperatures continue to increase, shifts in these metrics are expected due to the dynamics of the jet stream and increasing vapor pressure of water in the atmosphere. Also, if there are continued positive trends in these metrics, we expect significant hydrologic implications including the magnitude and return intervals of severe flooding and problematic nonstationarity (Milly et al., 2008) in precipitation and river discharge.

4.2 Methods

We characterized statistical trends in regional precipitation believed to have the greatest hydrological implications: the median and 95\textsuperscript{th} percentile of daily precipitation and wet and dry persistence. We used daily data from the Global Historical Climatology Network (GHCN), retrieved from the National Climatic Data Center (NCDC) and covering the entire NE US as defined by the National Climate Assessment. The NE US as defined for this study thus includes the states of Connecticut, Delaware, Maine, Maryland, Massachusetts, New Hampshire, New York, Pennsylvania, Vermont, West Virginia, and the District of Columbia. However, no climate stations from the District of Columbia or Maryland satisfied our selection criteria. Daily precipitation from 222 stations was analyzed with record lengths varying between 51 and 174 years and a mean record length of approximately 84 years. Stations were selected such that each had over 50 years of data and the last data point was recorded after January 1, 1990. We removed any station that was missing 10 continuous years of data; and daily precipitation values were rounded
to the nearest 1 mm. Station names and locations are included as supplemental information.

### 4.3 Characterization of Changes in Precipitation Extremes

For each station, depths of daily precipitation were subdivided and modeled using two distributions to better represent the extreme events of the distribution, that is, to better account for rare but important events. The first distribution was best fit to all daily precipitation depth values up to the 75th percentile and the second distribution was fit to the remaining upper tail. The lower values were fit utilizing an exponential distribution, while the upper values were fit with a generalized Pareto distribution. Both distributions were fit using the method of maximum likelihood estimation. The two distributions were fit for moving 30 year windows by month and annually. A 30-year window was chosen because it was found to generate enough samples within the upper 25 percent of the distribution to minimize noise in the Pareto fitting parameters without overly smoothing the signals. For each window the 95th percentile and median of daily precipitation were calculated from the two distributions. This was completed for each month and annually. The 95th percentile and median of daily precipitation were selected to represent heavy and average daily precipitation respectively. A linear model was fit to determine trends these metrics over time. Trend magnitudes were calculated using the slope of the best-fit linear model. Interquartile ranges were calculated for the trend magnitudes of each metric for the whole region by combining all 222 stations. Comparisons were performed between the number of positive trends and negative trends, and significant (p<0.01) positive and negative trends using the Mann-Kendall test.
4.4 Characterization of Changes in Wet and Dry Persistence

The Markov-chain parameters in this study represent the probability of transition from dry day to dry day ($P_{00}$) and the probability of transition from wet day to wet day ($P_{11}$). $P_{00}$ is used as an analogue for dry persistence while $P_{11}$ is used as the analogue for wet persistence. Wet days are defined as days that record ≥0.5mm of precipitation. For each station, a moving average of $P_{00}$ and $P_{11}$ was calculated by month and annually using a 30-year window. A 30-year window was used to be consistent with the window size used to characterize the precipitation extremes. Again, the slope of a best-fit linear model was used to calculate trend magnitudes in the metrics and comparisons were performed on the trends in $P_{00}$ and $P_{11}$ across the study region as described in the previous section.

4.5 Results and Discussion

The observation records show precipitation to be non-stationary in time. Of the four statistics computed, only median daily precipitation remained largely unchanged. The 95th percentile of daily precipitation for the study region generally increases over the observed record (Figure 4.1). More than 148 (two-thirds) of the 222 stations show positive trends for the 95th percentile of daily precipitation in the months of October through May and at least half of the stations display significant (p<0.01) positive trends during every month except July and September (Table 4.4). The strongest regional trend in the 95th percentile of daily precipitation was observed in April when the average trend was +0.7 mm per day per decade. It should also be noted that the interquartile range of the observed trends for the 95th percentile of daily precipitation is largest in September. Trends in the median of daily precipitation are much less pronounced with October being
the only month with more than half of the stations showing significant (p<0.01) positive trends; and there are no months in which more than half of the stations show significant negative trends for the median of daily precipitation (Table 4.3). These results are representative of the 10 NE US states. However, these trends are not spatially uniform. The entire region experienced an average trend of +0.5mm per decade in annual 95th percentile daily precipitation while Connecticut was found to have the greatest increase with a trend of +1.1mm per day per decade in annual 95th percentile daily precipitation. No trend was found for West Virginia in annual 95th percentile daily precipitation.

Figure 4.2 shows trends in both Markov-chain parameters, wet persistence ($P_{11}$) and dry persistence ($P_{00}$). However, the trends in dry persistence are generally smaller in magnitude with some seasonal variation, small increases in spring and small decreases in fall. For trends in dry persistence, the most positive trends (151) and significant (p<0.01) positive trends (117) occur in March; the most negative trends (152) occur in October, and the highest number of significant (p<0.01) negative trends (121) occur in September (Table 4.2). The wet persistence of events increases throughout the entire year with the greatest number of increasing trends occurring in May and June with 179 and 178 stations displaying positive trends, respectively, and 145 and 146 significant (p<0.01) positive trends, respectively (Table 4.1). May and June show the strongest trends with an average regional trend in the probability of a wet day following a wet day of +0.8 and
+0.9 percent per decade, respectively. The trends in Markov-chain parameters vary spatially. Vermont and Massachusetts displayed the greatest trends in wet persistence with the annual-averaged probability of a wet day following a wet day increasing by 0.013 per decade while Pennsylvania and Connecticut showed the smallest trend in annual wet persistence with increases of 0.003 per decade.

For daily precipitation events, the warmer months show the greatest increase in wet persistence, the colder months show larger increases in the magnitude of extremes, and dry persistence increases in early spring and decreases in early fall. Annually the interquartile ranges of the trends in both $P_{11}$ and the 95th percentile of daily precipitation are above zero. Therefore, on an annual basis, it is likely that the study region will experience increasingly persistent and intense precipitation events.

Our results are largely consistent with previous work on precipitation trends in the NE US. Wet and dry persistence, however, have not been studied in detail for the NE US. Studies of precipitation persistence have been performed in areas such as Europe where it has been observed that precipitation is trending toward longer wet spells with higher intensities (Zolina et al., 2010). Intense precipitation has been studied in the NE US (Douglas & Fairbank, 2011; Walsh et al., 2014). The National Climate Assessment reported that in the NE US more precipitation is falling annually and a higher percentage of rainfall is occurring in the upper 1 percent of daily events with time (Walsh et al.,
Our results are consistent with increases in total annual precipitation because, with increases in wet persistence and the 95th percentile of daily precipitation, and minimal trends in dry persistence and median daily precipitation, there would be more annual precipitation. Also, our results are consistent with an increased amount of precipitation occurring in the upper 1 percent of events. Our results are consistent because we found that the 95th percentile of daily precipitation was increasing which can be translated as a greater percentage of daily precipitation events falling above a stationary threshold in time.

Increases in the 95th percentile of daily precipitation indicate that the upper tail of the distribution of daily precipitation is increasing in magnitude, thus higher probability density in the upper percentiles of the distribution. If the probability of persistent precipitation is increasing along with the probability of observing a given high intensity event, then the probability of an intense event following a persistent pattern is likely increasing with time, which has significant flooding implications. High magnitude flooding can result even when long periods of time pass between a persistently wet regime and an intense precipitation event due to hysteresis within soils and watershed memory. All of this is consistent with an intensification of the water cycle and large amplitude, slow moving planetary waves. Another possible explanation for the observed increases in wet persistence during the spring months is that more moisture may be available earlier for evaporation as a result of earlier spring thaws. Similarly, if arctic regions that had previously stayed frozen are now thawing during summer months, this
could increase moisture fluxes into the northeastern US. These linkages would need further study, but it is possible that long-term satellite imagery of the northern hemisphere could be used for this.

Acknowledgments

The data for this paper is available from the National Climatic Data Center’s Global Historical Climatology Network – Daily (GHCN-Daily). This work was supported by Vermont EPSCoR through NSF Award EPS-1101317.

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Walsh, J. et al. (2014), Ch. 2: Our Changing Climate. Climate Change Impacts in the United States: The Third National Climate Assessment, doi:10.7930/J0KW5CXT.

**Figures and Tables**

**Figure 4.1** Regional trends in the median and 95th percentile of daily precipitation over the period of record for 222 Global Historical Climate Network stations. The dots represent the monthly or annual mean trend, the rectangle represents the interquartile range of the trend, and the whiskers represent the full range. Outliers are not shown for viewing purposes. This figure shows the trends in the 95th percentile of daily precipitation are most significant during December, March and April and are generally increasing at a greater rate than the median. However, there is much greater variability in the trends of the 95th percentile.
Figure 4.2. Regional trends in the Markov Chain parameters of daily precipitation over the period of record for 222 Global Historical Climate Network stations. The dots represent the monthly or annual mean trend, the rectangle represents the interquartile range of the trend, and the whiskers represent the full range. Outliers are not shown for viewing purposes. This figure displays the trends in $P_{11}$, the greatest increases in wet persistence occurred during the months of May and June, while trends in $P_{00}$ show decreasing dry persistence during September and October and increasing dry persistence in March.
Table 4.1. Statistical analysis of regional trends in the probability of a wet day following a wet day, $P_{11}$

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Table 4.2. Statistical analysis of regional trends in the probability of a dry day following a dry day, $P_{00}$

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Table 4.3. Statistical analysis of regional trends in median daily precipitation

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Table 4.4. Statistical analysis of regional trends in the 95th percentile daily precipitation

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Chapter 5: The Role of Snowpack in a Changing Hydrologic Regime

THE ROLE OF SNOWPACK IN A CHANGING HYDROLOGIC REGIME

Justin Guilbert¹
School of Engineering
University of Vermont
Burlington, VT 05405,

Ibrahim Mohammed
School of Engineering
University of Vermont
Burlington, VT 05405

Beverley Wemple
Rubenstein School of Environment
and Natural Resources
University of Vermont
Burlington, VT 05405

and

Arne Bomblies
School of Engineering
University of Vermont
Burlington, VT 05405

¹Corresponding author: jguilber@uvm.edu
Abstract

The impacts of climate change on watershed hydrology remain uncertain. In the northeastern United States, which has experienced increases in both precipitation and temperature, it is unclear if snowpack and snowmelt-driven streamflows will generally increase in response to higher precipitation or generally decrease in response to higher temperature. Prior observations showed a mixed response signal with no clear guidance for the future. This study disentangles the complex relationship between trends in temperature and precipitation and trends in seasonal streamflow in Vermont using a physical watershed hydrology. We find that increased temperature can lessen the impacts of increased precipitation to an extent at an annual level. However, we find that during snowmelt temperature plays a larger role than precipitation in determining the timing and daily peak magnitude of flows. We find that in scenarios where both temperature and precipitation increase that the snowmelt driven daily streamflow peak is likely to decrease in future climate scenarios during the winter and spring, when snowpack plays a dominant role in the translation of climate drivers to hydrologic response.

5.1 Introduction

As a likely result of continued global warming, the water cycle is changing and trends in hydrology must be considered, in stark contrast to the usual assumptions of stationarity in hydrology (Milly et al., 2008, Durack, et al., 2012; Groisman et al., 2005; Huntington, 2006; Milly et al., 2002). Changes in snowpack depth and extent associated with changes in climate are of great interest because of the snowpack’s role in surface energy balance and water storage, both of which regulate the hydrologic flow regime and affect flood
risk. Many studies have observed and predicted the impacts of climate change on snow
(Barnett, Adam, & Lettenmaier, 2005; Campbell et al., 2010; Krasting et al., 2013; Mote,
2006), but these changes have proven heterogeneous and difficult to generalize. For
example, Burakowski et al. (2008) showed that snow coverage has decreased in the
northeastern United States with increases in temperatures, while Cohen et al. (2012)
found that snow cover is increasing in some locations in the Northern Hemisphere due to
the increased availability of moisture in early winter. While the overall trend in snowpack
behavior remains spatially variable and uncertain, its hydrological consequences may be
significant. For example, increased snowpack may store more water for future release,
but changing temperature variability may also influence intra-seasonal melt patterns. The
balance of such effects can influence both the timing and amount of snowmelt discharge,
with serious implications for flood risk. In this paper, we decipher the impacts of
changing climate on wintertime watershed dynamics within the northeastern United
States and the resulting impact on spring and summertime flow using a representative
study watershed in central Vermont. We focus on the linkages of changing snowpack and
hydrologic extremes. This includes tracking the impact of a changing snowpack on
spring and summer flows using a distributed hydrological model. Snowmelt has
historically played a large role in the hydrology of the northeastern United States where a
large portion of annual streamflow volume occurs during the spring months (Hodgkins et
al., 2003). Streamflow in the northeastern United States follows a seasonal pattern
highlighted by high flows in spring and autumn, and low flows in summer and winter.
This seasonal cycle may be greatly impacted in future climates (Huntington et al., 2009).
Coincident with a warming climate, Huntington (2003) predicts that runoff will decrease significantly in the northeastern United States, and Hodgkins et al. (2003) found that the timing of peak spring melt is occurring earlier. In contrast to the temperature-driven signals discussed in Hodgkins et al. (2003) and Huntington (2003), the precipitation signals in the observed record may affect the streamflow that results from snowmelt in the opposite direction. It is unclear if the magnitude of the spring peak will increase or decrease, because extreme precipitation and precipitation persistence are increasing at the greatest rate during springtime in the northeastern United States (Guilbert et al., 2015) and the interaction of coincident precipitation and temperature anomalies may exacerbate rain-on-snow flooding. Although we explore primarily winter and spring hydrological responses, the changing winter and spring climate may also affect summer base flows through changes in infiltration that affect groundwater dynamics at longer time scales.

In this study, we investigate the individual and combined impacts of temperature and precipitation changes on streamflow and snow coverage to gauge sensitivity of hydrologic regime to climate change and variability.

5.2 Methods

5.2.1 Study Region

The Lake Champlain Basin is primarily composed of forested and agricultural land. A network of rivers and streams is distributed throughout the region, and the north–south-oriented Green Mountains form the major topographic feature of the study region. Elevations in the Lake Champlain Basin range from 30 meters above mean sea level
(MSL) on Lake Champlain to 1340 MSL at the top of Mount Mansfield, the highest point in Vermont, over a distance of less than 50 km. One of the many watersheds within the Lake Champlain Basin is the Mad River watershed, which is a sub basin of the Winooski watershed. The Mad River watershed is relatively undisturbed and mostly forested with two major ski resorts, agricultural flood plains and seasonal snow cover. In an average year with no major tropical depressions impacting the region, snowmelt tends to dictate the maximum streamflow magnitude of the year.

5.2.2 Climate and Streamflow Data

Three data sources were required to support this work: streamflow, observed climate data, and snow coverage data. Streamflow data were sourced from the USGS gauge on the Mad River near Moretown, Vermont as seen in Figure 5.1.
Figure 5.1. Mad River watershed located in central Vermont within the Winooski Watershed, which drains into Lake Champlain. Elevations range from 433 meters to 4083 feet.

Historical precipitation and temperature data were sourced from a reanalysis product produced by Maurer et al. (2002) of Santa Clara University, California (http://www.engr.scu.edu/~emaurer/data.shtml). Predictions of temperature change were derived from the Coupled Model Intercomparison Project (CMIP5), while changes in
precipitation intensity were derived from Guilbert et al. (2015), because CMIP5 precipitation fails to capture observed trends in Vermont (Mohammed et al., 2015). For snow coverage data, we used the Snow Data Assimilation System (SNODAS) dataset, which is offered at a 1km resolution (http://nsidc.org/data/G02158). SNODAS is a product generated by the National Oceanic and Atmospheric Association’s National Weather Service’s National Hydrologic Remote Sensing Center.

5.2.3 Model

We used the Regional Hydro-Ecologic Simulation System (RHESSys; Tague & Band, 2004), version 5.18.r2, to develop a spatially explicit, physically based hydrology model for the Mad River (Mohammed et al., 2015). The model explicitly routes water across the land surface. The model accepts many atmospheric inputs but only requires daily minimum and maximum temperature and daily precipitation. However, precipitation may be provided to the model at an hourly resolution to run the hydrology model at an hourly time step. The soil model of RHESSys is a two layer model which is not able to model frozen soils. Each time step, net canopy through fall and snowmelt are added to the surface (detention store) where excess water is either infiltrated, evaporates or becomes overland flow. Once infiltrated, water can move vertically through the water column where it can flow either to the saturated zone, the capillary zone, the root zone, or further into groundwater storage. The groundwater model is a simplistic representation which receives water from the saturated zone and moves water to the outlet in a semi-physical manner dependent on the parameters provided to the model upon initiation. These are discussed in the calibration section (5.2.4) below.
RHESSys’s snow model is a single layer, quasi-energy budget model that depends on precipitation and temperature to determine accumulation and melt. Snow accumulation is determined by two parameters that determine the maximum temperature at which snow can occur and the minimum temperature at which rainfall can occur. These temperatures are often non-zero due to the dependence of precipitation at the surface on temperatures at the elevation of the precipitation producing cloud. RHESSys simply interpolates a percentage when the temperature falls between the two provided values above. Radiation, latent and sensible heat, and advection all contribute to snowmelt. Melt from temperature and advection are only allowed to occur when the snow pack is ripe. The snow pack is considered ripe in this model when the snowpack energy deficit equals zero. energy deficit is decreased (negative increase) when air temperatures are below zero and are increased from temperatures above zero, rain and radiation. When the snowpack energy deficit is above zero melt can occur due to radiation in the form of sublimation or rain on snow.

5.2.4 Calibration

In addition to discharge at the watershed outlet, the model was also calibrated to the Snow Data Assimilation System (SNODAS) snow coverage maps (National Operational Hydrologic Remote Sensing Center, 2004) for accurate representation of snowpack extents. Snow water equivalent was not used because of great uncertainty in connecting highly variable point measurements to the gridded data products, which are also uncertain estimates. Calibration to SNODAS was accomplished with a Monte Carlo approach that was used to vary the key climatic parameters within the RHESSys model: lapse rate of
atmospheric transmissivity, dew point lapse rate, maximum effective leaf area index, temperature lapse rate, minimum precipitation to initiate cloud cover, sea level clear sky transmissivity, daylight temperature coefficient, wind speed, maximum temperature at which precipitation falls as snow, and minimum temperature at which precipitation falls as rain.

The climatic calibration parameters were varied within realistic ranges via 2,000 Monte-Carlo simulations. Calibration was performed on the period December 2006 through April 2007 with the goal of replicating the snow coverage area from SNODAS for eight dates throughout the snow cycle. The eight dates were evenly distributed throughout the snow cycle, the 1st and 15th of each month from December through April with the 7th and 21st of April being used to increase the resolution of the calibration during the snowmelt period. In each simulation, an output file was generated containing spatially and temporally varying values of snow coverage. These values were interpolated from the center point of RHESSys patches to the geo-referenced center of SNODAS cells. Cells were considered to have snow coverage if snow water equivalence exceeded 1mm. The correctness of any simulation was determined in time and space for the snow cycle. For each of the eight dates, there were 600 calibration points, and the number of correctly predicted cells (snow/no snow) was calculated by summing the total number of true positives and negatives and then subtracting the total number of false positives and negatives. Snow calibration was considered acceptable above a threshold of 85% agreement between RHESSys and SNODAS. As expected, the calibration of snow coverage was most sensitive to two parameters, minimum temperature for rainfall and
maximum temperature for snowfall which RHESSys utilizes to determine if precipitation becomes snow, rain or a mixed event. These two parameters were held constant for the remaining calibration iterations, while all other calibration parameters were allowed to vary slightly in order to simultaneously maximize both flow and snow calibrations.

When calibrating RHESSys to streamflow, the values of the climatic calibration parameters were fixed and assumed to be correct, while the following five hydrologic parameters were varied: (1) the proportion of water moving from the saturation store to the ground water store (GW1), (2) the proportion of water moving from the ground water store to the stream (GW2), (3) the decay of hydraulic conductivity with depth (m), (4) the hydraulic conductivity at the surface (k) and (5) the soil depth (z). The model was calibrated to the water year 1970, because it was closest to the average water year for the watershed. Calibration was again performed on the five parameters via Monte Carlo simulations. Once streamflow and snow coverage calibration values were deemed acceptable, selecting a threshold value for Nash Sutcliffe Efficiency (NSE) of 0.6 or greater for the calibration year, the model was run for an additional 49 years to ensure that streamflow and snow calibrations were acceptable for multiple years with respect to NSE.

5.2.5 Analysis of Snowpack and Streamflow

The RHESSys model was validated for the water years 1961 to 2009. The water years from the validation period that were best fit to streamflow (NSE > 0.6) were used for subsequent analysis of sensitivity to climatic change and will be referred to as the
selected years through the remainder of this paper. A threshold value of 0.6 for NSE was selected to allow for more years of analysis, while still remaining in an acceptable range for model fit (Anderson & Rounds, 2010). SNODAS data are only available from 2003 to present (2016) so calibration and validation of the snow model could not be performed on the best fitting years from the streamflow model. Calibration of the snow model was performed on the winter of 2006. The model was validated for the water years of 2004, 2005, 2006, 2008 and 2009. In the validation of the snow model the percentage of correctly identified (snow or no snow) cells was determined both spatially and temporally for the duration of the snow season.

In order to better understand the impacts of changing temperature and precipitation on the watershed response variables (i.e., daily streamflow and peak snowmelt streamflow magnitudes) at the outlet of the Mad River, the watershed was run under 16 different climate scenarios. The IPCC projects an increase in temperature of 2°C by the middle of the century, assuming the world continues along the highest emission scenario, relative concentration pathway 8.5 (RCP8.5) which assumes additional radiative forcing of 8.5 W m⁻². Therefore, scenarios of projected temperature change 0% (0°C), 50% (1°C), 100% (2°C) and 150% (3°C) for mid-century were generated. Likewise, change scenarios of 0%, 50%, 100% and 150% of the observed trends in daily precipitation intensity discussed in Guilbert et al. (2015) were extrapolated to generate precipitation scenarios. Changes in precipitation intensity were calculated for a period of 50 years using the month that the precipitation occurred and the percentile of daily precipitation. Each of the four temperature scenarios were combined individually with each precipitation scenario.
to generate 16 unique combinations ranging from zero change in temperature and precipitation (zero change/observed scenario) to 150% of projected temperature and precipitation change (high change scenario). The scenarios are labeled using two-letter combinations in which the first letter represents temperature change and the second letter represents precipitation change. The four letters used for reference scenarios are Z, zero change scenario, L, low change scenario, M, moderate change scenario, and H, high change scenario. Values are shown in Table 5.1.

**Table 5.1. Temperature and precipitation trend scenarios**

<table>
<thead>
<tr>
<th>Temperature Change Scenario (°C)</th>
<th>Precipitation Change Scenario (% of observed trend)</th>
<th>0 (Z)</th>
<th>50 (L)</th>
<th>100 (M)</th>
<th>150 (H)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0 (Z)</td>
<td>ZZ</td>
<td>ZL</td>
<td>ZM</td>
<td>ZH</td>
<td></td>
</tr>
<tr>
<td>1 (L)</td>
<td>LZ</td>
<td>LL</td>
<td>LM</td>
<td>LH</td>
<td></td>
</tr>
<tr>
<td>2 (M)</td>
<td>MZ</td>
<td>ML</td>
<td>MM</td>
<td>MH</td>
<td></td>
</tr>
<tr>
<td>3 (H)</td>
<td>HZ</td>
<td>HL</td>
<td>HM</td>
<td>HH</td>
<td></td>
</tr>
</tbody>
</table>

For example, HM represents the scenario with the highest temperature change (3°C) and moderate precipitation change (100% of observed trend).

### 5.3 Results

#### 5.3.1 Model Calibration and Validation

For the calibration year 1970, a Nash-Sutcliffe efficiency value of 0.71 was achieved, surpassing the calibration threshold of 0.6. The parameter set used to achieve this was 4.1(m), 36.8 (k), 1.21(z), 0.184 (GW1), and 0.893 (GW2). The validation showed a Nash-Sutcliffe efficiency value (NSE) of 0.59 for cumulative monthly streamflows for the duration of the validation period and $R^2$ for the correlation of modeled to observed is 0.83 (Figure 5.2).
The annual NSE values varied from -0.56 to 0.76 with three values below 0.0, four above 0.6, and the remaining 42 values between 0.0 and 0.6. The four years in which NSE value was greater than 0.6 were selected for further study. The selected years are 1970, 1972, 1983, and 1993.

With respect to the validation of the snow model, for the validations years described above, percentage values ranged from 73% to 97% with a mean of 90% during the validation period and full results are shown in Table 5.2 for the validation period in comparison to streamflow performance.

**Figure 5.2.** Correlation between observed monthly streamflow and RHESSys modeled streamflow for water years 1961 through 2009. $R^2 = 0.83$.
Table 5.2 Comparison of years used for snow coverage calibration and validation and corresponding Nash-Sutcliffe efficiency (NSE) value. This table shows that the snow coverage model performs well for all of the years in which data is available from the Snow Data Assimilation System with just one year with a percent correct value below 90%.

<table>
<thead>
<tr>
<th>Water Year</th>
<th>2004</th>
<th>2005</th>
<th>2006</th>
<th>2007</th>
<th>2008</th>
<th>2009</th>
</tr>
</thead>
<tbody>
<tr>
<td>NSE Value</td>
<td>0.40</td>
<td>0.53</td>
<td>0.50</td>
<td>0.44</td>
<td>-0.51</td>
<td>0.07</td>
</tr>
<tr>
<td>Snow Cover Correct (%)</td>
<td>92</td>
<td>97</td>
<td>73</td>
<td>93</td>
<td>96</td>
<td>90</td>
</tr>
</tbody>
</table>

We tested for bias in the selected years from the validation period.

Table 5.3 presents the percentiles of where the calibrated years’ streamflow statistics fall with respect to the rest of the validation period. For example, for the statistic labeled Total Summer Flow, the four calibrated years cover a large range of historic total summer flow values both (shaded green) above and below (red) the historic summer flow median ranging from the 5th percentile to the 75th percentile. The results (Table 5.3. Percentiles of selected years’ streamflow statistics as compared to entire validation period, water years 1961 through 2009. The parameters selected favor late spring melts and low to average fall conditions. Green represents above the 50th percentile and red represents below the 50th percentile for each respective streamflow statistic.)
Table 5.3) show that for the majority of observed streamflow statistics, the parameter set has low performance bias. For six of the nine streamflow statistics, the calibrated years fall in both the lower and upper 50th percentiles, which demonstrates good coverage and minimal bias for those six statistics. However, the parameter set performs well on years with low cumulative fall streamflows and late spring melts.

**Table 5.3.** Percentiles of selected years’ streamflow statistics as compared to entire validation period, water years 1961 through 2009. The parameters selected favor late spring melts and low to average fall conditions. Green represents above the 50th percentile and red represents below the 50th percentile for each respective streamflow statistic.

<table>
<thead>
<tr>
<th>Year</th>
<th>Total Annual Flow</th>
<th>Total Winter Flow</th>
<th>Total Spring Flow</th>
<th>Total Summer Flow</th>
<th>Total Fall Flow</th>
<th>Maximum Spring Flow</th>
<th>Timing of Maximum Spring Flow</th>
<th>Timing of ½ of Spring Volume</th>
<th>Number of Above 90% Flows</th>
</tr>
</thead>
<tbody>
<tr>
<td>1970</td>
<td>50</td>
<td>63</td>
<td>71</td>
<td>5</td>
<td>46</td>
<td>55</td>
<td>85</td>
<td>85</td>
<td>57</td>
</tr>
<tr>
<td>1972</td>
<td>57</td>
<td>38</td>
<td>90</td>
<td>75</td>
<td>5</td>
<td>90</td>
<td>90</td>
<td>100</td>
<td>71</td>
</tr>
<tr>
<td>1983</td>
<td>65</td>
<td>57</td>
<td>98</td>
<td>61</td>
<td>13</td>
<td>50</td>
<td>85</td>
<td>93</td>
<td>88</td>
</tr>
<tr>
<td>1993</td>
<td>25</td>
<td>40</td>
<td>36</td>
<td>28</td>
<td>44</td>
<td>38</td>
<td>65</td>
<td>56</td>
<td>32</td>
</tr>
</tbody>
</table>

A set of five of the nine streamflow statistics shown in
Table 5.3 which display the extent of bias are compared to the NSE value for the respective year in Figure 5.3.

Figure 5.3 Comparison of Nash-Sutcliffe efficiency value to percentile values of select streamflow statistics. The selected years for subsequent climate change sensitivity analysis are shown as filled in black circles. The empty blue circles represent the remaining 45 years in the validation period for streamflow. The grey line is a reference to the median value of 50. The highest bias in the selected years appears in panels C and E which show that the model performs well in years in which total autumn flow is low to moderate and spring melt is late as previously discussed.

5.3.2 Simulated Climate Scenarios
Precipitation drives streamflow and in all scenarios in which precipitation is increasing, cumulative annual streamflow also increases. However, when temperature effects are isolated, increases in temperature lead to reduced annual streamflows, presumably due to increases in evapotranspiration. Peak flows also are decrease because of the lessened role of snowpack in the spring with a rising temperature. To study this effect, the relative impacts of changing precipitation scenarios and temperature scenarios were evaluated using the model. An example (Figure 5.4) reveals a number of changes that could result in varying climatic conditions. First, temperature increases reduce the intensity of the spring melt period due to lower availability of snow. The melt period also lasts longer. Similarly, as precipitation increases, the magnitude of the spring peak increases. Therefore, the spring meltwater pulse becomes larger and more defined. We investigated these results (reported below) in much greater detail.
Figure 5.4. Cumulative streamflow and snow depth for multiple scenarios. The top left panel shows inter-annual variability of the four select years for this study in the baseline scenario. The top right panel shows the model forced for one year (1970) under four changing temperature scenarios while holding precipitation at baseline conditions. The bottom left panel shows one year forced by four precipitation scenarios while holding temperature at baseline conditions. The bottom right panel shows one year forced by four scenarios in which precipitation and temperatures increase at comparable rates. The solid and dashed black lines represent the same scenario-year combination (ZZ-1970) in all four subplots for comparison between subplots.
For every increasing temperature scenario defined above, the peak snowmelt streamflow magnitude decreases by an average of 5.5mm; and half of the streamflow volume between January and June occurs an average of 2.3 days earlier when precipitation is held constant. If temperature is held constant as shown in Figure 5.5, for every increasing precipitation scenario the peak snowmelt magnitude increases by an average 2.2 mm and half of the streamflow volume between January and June occurs 1.3 days later.

**Figure 5.5.** Response of peak daily streamflow during snowmelt by temperature change scenario. As temperatures increase, the marginal impact of precipitation changes on snowmelt decreases. This effect becomes clear past a 1°C increase in temperature.

As shown in Figure 5.5, the magnitude of the response of snowmelt streamflow magnitudes is not consistent across temperature change scenarios as the slope of the
response decreases with increasing temperatures. Therefore, temperature dominates the relationship between snowmelt and precipitation for spring flow generation. If the ZZ, LL, MM, and HH scenarios are subtracted from each other respectively and averaged (which reduces to $\frac{1}{3}(\text{Mean Peak Snowmelt}_{HH} - \text{Mean Peak Snowmelt}_{ZZ})$) the snowmelt streamflow peak decreases by 3.4mm on average, and half of the volume of streamflow between January and June occurs an average of 1.1 days earlier (which reduces to $\frac{1}{3}(\text{Mean Date of Half Volume}_{HH} - \text{Mean Date of Half Volume}_{ZZ})$). The stepwise comparison of this scenario progression from ZZ to LL to MM to HH will be referred to as “relative increases in precipitation and temperature” for the remainder of the results section.

Snowpack statistics, as would be expected, change in accordance with snowmelt statistics. On average the last time the watershed is 80% covered by snow shifts 4.5 days earlier when only temperature increases, 0.4 days later when only precipitation increases and 4.2 days earlier when stepping through relative increases in precipitation and temperature. Similarly, the watershed will be barren of snow an average of 3.3 days earlier, 0.4 days later and 2.8 days earlier per temperature scenario, precipitation scenario, and per step in relative increases in precipitation and temperature, respectively. The implications of these changes on water resources may be of significant interest, which we also explored. When comparing seasons, DJF represents winter, MAM represents spring, JJA represents summer, and SON represents autumn. Cumulative winter flows increase in all scenarios compared to baseline (ZZ), and the highest change

95
scenario (HH) shows the greatest increases. Temperature and precipitation both increase cumulative winter flows. Similar patterns to winter were found in summer and fall. A comparison of cumulative seasonal flows is presented in Figure 5.6. The slope of the curves in Figure 5.6 become steeper as the temperature scenarios increase for winter, summer and fall, showing increased streamflow sensitivity to precipitation shifts with increasing temperatures. The opposite is true for spring, where the steepest slope is found for no temperature change.
Figure 5.6. Cumulative seasonal flows by temperature change scenario. From left to right and top to bottom, seasons are ordered winter, spring, summer, and fall, respectively.

The similar pattern found for both winter and summer (where streamflow increases for increases in temperature with precipitation held constant) is as expected for winter, but counterintuitive for summer. We found no suitable explanation of this summertime result.
of the RHESSys model. Summer cumulative and minimum 7-day streamflow values were found to increase with increases in temperature, while precipitation is held constant which is paradoxical as greater levels of potential evapotranspiration should occur with increases in temperature. The increased streamflow was traced back to an increase in base flow as generated from the model. However, the ultimate source of this higher groundwater contribution is still unresolved.

Simulated sensitivities varied greatly by season. The average increase in cumulative summer flow was 0.05mm/day and 0.08mm/day with each increase in temperature and increase in precipitation, respectively. The changes in winter, summer and fall cumulative flow are much smaller than those found in spring by roughly an order of magnitude.

5.4 Discussion and Conclusions
Changes in climate can greatly impact the hydrology of a seasonally snow dominated region. Snowpack is governed by temperature and precipitation and with both likely to continue to increase in future climates in the northeastern United States, the relative impact of these two controlling variables on watershed response is dominated by precipitation, with a secondary effect of temperature. Precipitation response varies with temperature, and shows a threshold effect for sensitivity. With increasing temperatures, the duration of snow cover shrinks and overall depth decreases. Temperature-related factors of such changes include decreased percentage of precipitation falling as snow, frequent snowmelt events, and shortened snow cover duration. The lower frequency of precipitation falling as snow also may allow the surface of any existing snowpack to age,
which decreases its albedo and leads to more rapid and frequent melts (not simulated in our study). Similarly, when there is no snow the albedo drops further for bare ground, which can lead to increased ground and near surface temperatures that decreases the chance that a snowpack will redevelop. However, with increasing precipitation intensities and overall amount, our model results show a possibility that these increases may mitigate some of the impacts that increasing temperatures have on decreasing snow packs or lessen the shift toward decreasing spring flow volumes and magnitudes. This all translates to a shift in hydrologic regime especially during the spring-melt season. One notable result is that increased temperatures lead to earlier melts which when combined with limited snowpack leads to lower peak spring flow magnitudes. In other words, risk of destructive snowmelt-generated spring floods appears to be declining and will likely continue to decline, despite increasing precipitation intensity. Floods generated by anomalous precipitation in summer and fall when snowpack is absent (such as tropical storms) will not be affected by these changes, and we did not include such events in our study. It may be that the annual maximum flood becomes more dominated by such events (as opposed to spring melt flows) in the future and that the annual maximum flow will occur more frequently outside of the snow-melt season.

With decreasing snowmelt the overall volume of spring flows is likely to decrease in future periods. It would seem plausible that summer low flows would decrease as well. However, this study found that the magnitude of cumulative 7-day summertime low flows increases in every scenario compared to the base scenario. This is in contrast to the findings of Campbell et al. (2011) who projected increased drought risk in the future and
Hayhoe et al., (2006) who found decreasing summertime low flows in the northeastern United States. However, decreasing summer flows have not been observed historically (Hodgkins et al., 2005). Increased summer low flows could benefit stream ecology as the effect will reduce the frequency of extreme low flows, which have been shown to negatively impact fish abundance and species composition (Smakhtin, 2001). Also, increased low flows may lower in-stream temperatures resulting from increased atmospheric temperatures, which has been shown to stress fish populations (Mantua et al., 2010). We are unsure of the exact mechanism of higher modeled summer flows, but the result may be related to increased infiltration during more frequent episodes of snowmelt under a warming climate. This study does not consider increasingly persistent multi-day rainfall events as found in Guilbert et al., (2015), which could further increase summertime flood risk as a result of generally wetter antecedent conditions.

There are some limitations to this study. The model was very difficult to calibrate and once calibrated, as shown above, only performed well (NSE>0.5) for 9 of the 49 validation years. Poor model performance could be a result of one of two reasons. First, inability of climate input data to resolve characteristics of rainfall adequately could be a major limitation. Climate data is sparse in this region and Maurer et al. (2002) heavily relies on climate station data. Also, some assumptions of the Rhessys model could negatively impact performance. For example, the model assumes that daily precipitation values fall evenly throughout the day (Tague & Band, 2004), and due to the medium size of the watershed, daily resolution may not be ideal. A possible future solution for
this relies on a relatively new climate station that has been placed within the watershed (installed post-2010).

Acknowledgements

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References


Chapter 6: Summary and Conclusions

The work presented in this dissertation expands on the research done by others on the response of a watershed in the northeastern United States to changes in climate, specifically, increases in temperature and precipitation. This work builds on the work of Mohammed et al. (2015), who investigated the impacts of climate change on the same study site using CMIP5 data to drive model projections and showed that the CMIP5 models failed to replicate trends in precipitation. One primary difference is that the research in this dissertation used historical trends in precipitation to inform climate scenarios. A second difference is that I calibrated to snow cover because I expected snow to be a large driver of hydrology in our region with respect to storing water and increasing albedo. The first goal of this work was to better understand the range of possible outcomes for temperature and precipitation change. Then to better understand the relative contributions of shifts in temperature and precipitation to shifts in hydrologic statistics with a focus on the role of snow as an internal variable that is impacted by shifts in climate, while also modulating streamflow during the melt period.

In Chapter 3, we investigated the projections of general circulation models from the Coupled Model Intercomparison project phase 5 forced under the IPCC’s AR5 moderate and high relative concentration pathways. We determined likely shifts in mean climate for mid- and late-century, with respect to temperature and precipitation. Consistent with many other studies of climate change in the northeastern United States, we found that temperature and precipitation are expected to increase with radiative forcings. We used the projected changes to generate relevant impact metrics that can help describe the
impacts of climate change. These metrics included days above 80°F Fahrenheit (+13.4 days by late century), length of the growing season (+43.1 day by late century), cumulative seasonal snowfall (-50% by end of century), and days suitable for maple syrup production (-11.5 days by end of century).

In Chapter 4, we investigated the observed climate record of the northeastern United States for trends in precipitation. All climate stations with long records (>50 years) and minimal breaks in data (<10 years) were considered. We found that intensities of high percentile daily precipitation events are increasing, while median daily rainfall is largely remaining constant. This is consistent with other studies that found an overall intensification of the water cycle (Huntington, 2006). However, the most significant discovery of this study was that the northeastern United States are experiencing increasingly persistent rainfall events especially during the months of May and June where 145 and 146 out of 202 climate stations showed significant (p<0.01) positive trends in wet persistence, respectively. Dry persistence was found to be decreasing significantly (p<0.01) in October at 121 climate stations and increasing significantly (p<0.01) in March at 117 stations. It is unclear which mechanism(s) may be generating persistent weather systems, but it is believed to be a mechanism related to larger scale atmospheric processes, especially those related to amplified planetary waves. We believe that persistent precipitation plays a larger role in the genesis of floods outside of the time of the year in which snowmelt dominates hydrology.

In Chapter 5, we investigated the combined and individual impacts of increased temperature and precipitation on the study region’s hydrology, especially the effects of
climate on snow and snowmelt driven streamflows. We found that snow cover duration is decreasing, and that snowmelt streamflow peaks are occurring earlier and are of lower magnitude.

As expected, we found that projections of climate change for the northeastern United States show that increased temperature and precipitation are likely to occur under the expected emission scenarios (moderate and high). We identified a significant new signal of climate change in the observed record in that the persistence of wet periods is increasing throughout the year while the persistence of dry periods is increasing in spring and decreasing in the fall. As expected, snow is an important driver of flow. When looking at future scenarios, we concluded that for the competing mechanisms of increasing temperature and precipitation that temperature plays a dominant role in controlling streamflow during snowmelt. However, outside of the snowmelt period, peak magnitude streamflows are likely to increase with increasingly intense precipitation events.

There is a general physical-based understanding of why precipitation intensities are increasing based on the Clausius-Clapeyron relationship. However, what is not well understood is the larger-scale atmospheric variability and the generation of persistent systems. As is well displayed in Brown et al., (2002), there have been times during the last 10,000 years with clusters of high magnitude floods in the northeastern United States. This suggests that anthropogenic greenhouse gases and related temperature increases may not be the only driver of persistent systems associated with high magnitude flooding in
our region. Further study is needed to determine the impacts of anthropogenic forcings versus natural variability when considering extreme persistence and associated flooding.

The Vermont EPSCoR project Research on Adaptation to Climate Change (RACC) project which will be continued to some extent with another Vermont EPSCoR project Basin Resilience to Extreme Events (BREE), both focus on impacts of climate change on hydrology and the resultant impacts on Lake Champlain water quality. One major impact of climate that is a focus of RACC is the resulting transport of sediments and nutrients to the lake. Only inferences can be made here as this research focused on just one relatively small watershed of Lake Champlain. The snowmelt peak streamflow magnitude will likely decrease in future climates while peak streamflow magnitudes outside of the snowmelt season may increase. This trend in streamflow may alter the timing of the largest accumulations of sediments and nutrients at the outlets of the various watersheds on Lake Champlain. Another concern is what may happen during the winter with respect to sediments and nutrients. There are many interactions that point to the importance of the timing of manure spreading in the region. With increasing temperatures, there will likely be bare ground throughout the basin at a higher frequency during the winter season. When this occurs, it is possible that the ground will freeze and then thaw from the surface down. On manure sprayed fields, the surface contains the nutrient rich soils of the manure application. These nutrients will be readily available for transport across the frozen soil layers below the surface during a precipitation event or in the more frequent shift from snow covered to bare soil. This may cause nutrient pulses to the lake during
times when the sections of the lake are still ice covered which may cause a number of new issues in nutrient management.

When considering wintertime nutrient fluxes, a limitation of the RHESSys model is obvious as there is no mechanism for modeling frozen soil hydrology and soil temperatures are modified the same if there is snow cover or not. Trends in the sub-daily intensity of precipitation were not modeled in this study. This is where there is room for improvement and a limitation. When provided with daily data for precipitation, the model divides that value by 24 and spreads the precipitation evenly across the day. For the study region this may underestimate flow especially for short duration, high intensity precipitation event. This leads further work that could be performed. We suggest using hourly data with this model to further enhance understanding of shifts in peak flow magnitudes since peaks in streamflow occur at temporal resolutions much finer than the daily level for our study region.

To achieve this we suggest utilizing data from the climate station that was installed in the Mad River watershed in July of 2013. This will provide higher temporal resolution precipitation data to the model while likely improving the climate data quality over the gridded product that was used (Maurer et al., 2002). Also, to better understand the system, we suggest adding increased persistence to the precipitation trend. The addition of persistence could be very important when considering the likelihood of annual peak flow magnitudes shifting outside of the snowmelt season by mid-century. We omitted persistence in this study as we used modified observational climate data (Maurer et al., 2002) to force the model. We thought that if we were to change the days with or without
precipitation that we would in effect be losing the joint distribution with temperature. In addition to persistence being a consideration, we also suggest further research into the RHESSys code or that an alternative model be considered, as we were unable to identify the source of streamflow increases during the summer season with increasing temperatures when precipitation was held constant.

We are unsure why calibration was so difficult for this model. When compared to the work of Mohammed et al. (2015), who used just 5500 calibration parameter sets and achieved multiple sets of NSE values that were higher than those that we achieved. Calibration took a much longer time for this work and required many tens of thousands of parameter sets. The only difference being the version of RHESSys that was used. A part of the calibration issue may be resolved by driving the model with better climate data from the weather station as described above.

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