

**MODELLING THE POTENTIAL IMPACTS OF CLIMATE CHANGE ON  
SNOWPACK IN THE ST. MARY RIVER WATERSHED, MONTANA**

**RYAN J. MACDONALD**

B.Sc., University of Lethbridge 2006

A Thesis  
Submitted to the School of Graduate Studies  
of the University of Lethbridge  
in Partial Fulfillment of the  
Requirements for the Degree

**MASTER OF SCIENCE**

Department of Geography  
University of Lethbridge  
LETHBRIDGE, ALBERTA, CANADA

© Ryan James MacDonald, 2008

## ABSTRACT

Climate change poses significant threats to mountain ecosystems in North America (Barnett *et al.*, 2005) and will subsequently impact water supply for human and ecosystem use. To assess these threats, we must have an understanding of the local variability in hydrometeorological conditions over the mountains. This thesis describes the continued development and application of a fine scale spatial hydrometeorological model, GENESYS (GENerate Earth SYstems Science input). The GENESYS model successfully simulated daily snowpack values for a 10 year trial period and annual runoff volumes for a thirty year period. Based on the results of these simulations the model was applied to estimate potential changes in snowpack over the St. Mary River watershed, Montana. GCM derived future climate scenarios were applied, representing a range of emissions controls and applied to perturb the 1961-90 climate record using the “delta” downscaling technique. The effects of these changes in climate were assessed for thirty year time slices centered on 2020s, 2050s, and 2080s. The GENESYS simulations of future climate showed that mountain snowpack was highly vulnerable to changes in temperature and to a lesser degree precipitation. A seasonal shift to an earlier onset of spring melt and an increase in the ratio of rain to snow occurred under all climate change scenarios. Results of mean and maximum snowpack were more variable and appeared to be highly dependent on scenario selection. The results demonstrated that although annual volume of available water from snowpack may increase, the seasonal distribution of available water may be significantly altered.

## **ACKNOWLEDGEMENTS**

The past two years have been an incredible life experience, I am grateful for having opportunity to learn and grow as an individual as well as a researcher. This project was made possible through funding from the Alberta Ingenuity Center for Water Research (AICWR). I owe so many thanks to my supervisor, Dr. Jim Byrne for his continual guidance and support through this entire process. Dr. Stefan Kienzle has also been a large part of this project since the beginning and has provided invaluable perspective and direction. I would also like to thank Dr. Larry Flanagan for being a member of my committee and providing very useful comments throughout this thesis. As an external examiner Dr. John Horel also provided useful input to this thesis.

There have been numerous other individuals at the University that have provided me with support throughout the past two years. I would particularly like to thank Dr. Craig Coburn, Dr. Hester Jiskoot, Dr. Matthew Letts and Dr. Joe Rasmussen who have shown me the perspective I couldn't see.

This past two years would have been impossible without the help of my family and friends. I owe particular thanks to my office mates Kate and Rob, it wouldn't have been the same without them. I was lucky to have Ed and Cathy Anderson who let me live in their house for almost two years. I thank my mom and Ed for always being the support I needed and my dad whom I owe my love for the outdoors to. I owe the first four years of my education to Kelly, my sister. I would particularly like to thank Jenn who has been my strongest support and has given so much to my life. For her I am forever grateful.

## TABLE OF CONTENTS

ABSTRACT .....	ii
ACKNOWLEDGEMENTS .....	iii
TABLE OF CONTENTS .....	iv
LIST OF FIGURES.....	vi
LIST OF TABLES.....	viii
CHAPTER 1 .....	1
1.1 Introduction.....	1
1.2 Research Objectives.....	4
1.3 Thesis Structure .....	5
CHAPTER 2 .....	7
2.1 Introduction.....	7
2.1.1 Background .....	8
2.2 Study area .....	13
2.3 Model description.....	14
2.3.1 Derivation of Terrain Categories .....	15
2.3.2 MTCLIM .....	16
2.3.2.1 Solar Radiation .....	17
2.3.2.2 Temperature .....	17
2.3.2.3 Humidity.....	18
2.3.2.3 Precipitation .....	18
2.3.2.3.1 <i>Precipitation-elevation relationship</i> .....	19
2.3.2.3.2 <i>Adjusting for seasonality of precipitation</i> .....	21
2.3.2.3.3 <i>Determination of monthly mean precipitation values</i> .....	23
2.3.2.3.4 <i>Application to daily precipitation data</i> .....	25
2.3.3 Snow pack and soil water model .....	26
2.3.3.1 Precipitation Partitioning .....	26
2.3.3.2 Canopy Interception .....	27
2.3.3.2.1 <i>Snow interception</i> .....	27
2.3.3.2.2 <i>Rain interception</i> .....	29
2.3.3.3 Sublimation .....	29
2.3.3.4 Snow Melt.....	31

2.3.3.5 Evapotranspiration .....	32
2.3.3.6 Soil Moisture .....	33
2.3.4 Model Application and Validation .....	34
2.4 Results .....	35
2.5 Discussion .....	43
2.6 Summary .....	47
CHAPTER 3 .....	49
3.1 Introduction .....	49
3.1.1 GCM description .....	50
3.1.2 Modelling approach .....	51
3.2 Study Area .....	54
3.3 Methods .....	55
3.3.1 Hydrometeorological model .....	55
3.3.2 GCM data .....	57
3.3.3 Downscaling .....	59
3.3.4 Data analysis .....	60
3.4 Results and Discussion .....	61
3.4.1 Climate Change Scenarios .....	61
3.4.2 Rain to snow ratio .....	61
3.4.3 Spatial change in mean annual snowpack .....	63
3.4.4 Volume of Maximum Snowpack .....	66
3.4.5 Timing of Maximum Snowpack .....	68
3.4.5 Effect of Scenario Selection on Future Snowpack Predictions .....	72
3.5 Summary and Conclusions .....	73
CHAPTER 4 .....	76
4.1 Summary and Conclusions .....	76
4.2 Recommendations .....	79
REFERENCES .....	82

## LIST OF FIGURES

Figure 1 The St. Mary Watershed in Montana and Southern Alberta. The locations of the watershed stream gauging point, Preston snow survey, Many Glacier SNOTEL site and St. Mary and Babb weather stations are shown. ....	13
Figure 2 Linear relationship between local elevation in relation to the Man Glacier SNOTEL site and dSWE at the Preston snow course. ....	20
Figure 3 Linear relationship between local elevation in relation to St. Mary and dSWE at the Preston snow course. ....	21
Figure 4 Seasonal distribution of precipitation at Many Glacier and St. Mary. ....	22
Figure 5 Mean monthly precipitation change as a function of season and elevation. ....	24
Figure 6 Varying monthly mean precipitation ratios between St. Mary and TCs at the elevation bands 1400m, 2000m and 2800m. ....	25
Figure 7 Comparison between simulated annual runoff and observed annual flow volume. ....	36
Figure 8 Comparison between daily simulated snowpack (mm SWE) and observed snowpack at Many Glacier for the 10 year trial period. ....	37
Figure 9 Comparisons of simulated daily snowpack (mm SWE) and observed daily snowpack at Many Glacier for the years 1979, 1980, and 1981. ....	39
Figure 10 Simulated April 1 SWE over the entire St. Mary watershed from 1961 to 2000. ....	40
Figure 11 Comparison of simulated April 1 SWE between 3 elevation bands. ....	41
Figure 12. Soil moisture and Precipitation plotted for 2 water years (Apr 21, 1987 to Nov 7, 1989).....	42

Figure 13 Comparison of August 31 soil water at 3 elevation bands.....	42
Figure 14 The St. Mary Watershed in Montana and Southern Alberta. The locations of the watershed stream gauging point, Preston snow survey, Many Glacier SNOTEL site and St. Mary and Babb weather stations are shown. ....	54
Figure 15 Spatial change in mean annual snowpack (mm SWE) representing the greatest (CGCM), median (CSIRO), and least (NCAR) changes relative to the 1961-90 base period. ....	64
Figure 16 Time series of maximum snowpack for the greatest and least change scenarios. ....	67
Figure 17 Time series of the Julian date of maximum snowpack for the greatest and least change scenarios.....	69

## LIST OF TABLES

Table 1 Linear relationships between mean monthly precipitation at St. Mary and mean monthly precipitation at Many Glacier (n=23).....	23
Table 2 Comparison of date of complete melt at different point melt factors (PTM). .....	32
Table 3 1961-90 water balance for 3 elevation bands, values represent a 30 year average in mm of SWE. ....	36
Table 4 GCMs and experiments available from the PCIC (after Barow and Yu, 2005)...	58
Table 5 Model experiments used in this study (after Barrow and Yu, 2005).....	58
Table 6 Description of SRES scenarios used in this study (after Barrow and Yu, 2005).	59
Table 7 Annual changes in temperature and precipitation for each of the 5 scenarios relative to the 1961-90 base period.....	61
Table 8 Changes in rain and snow as a percentage of the total precipitation for all 5 scenarios. ....	62
Table 9 Comparison of mean annual snowpack (mm of SWE) and percent change in mean annual snowpack relative to the 1961-90 historical period.....	65

# CHAPTER 1

## INTRODUCTION

### 1.1 Introduction

Environmental, economical, and societal health are dependent on an adequate supply of water resources (Arnell, 1999; Lettenmaier *et al.*, 1999). Over 40% of the world's population resides within watersheds that originate from mountains (Beniston, 2003). Climatic processes control and are controlled by the physiographic characteristics of mountains (Beniston *et al.*, 1997). Mountains are expected to be one of the areas where the impacts from climate change will be most noticed (Bonsal *et al.*, 2003; Field *et al.*, 2007). One of the most significant impacts from climate change in mountainous areas will be the anticipated reduction in snowpack and a substantial shift in the timing and availability of fresh water supply (Barnett *et al.*, 2005; Field *et al.*, 2007).

In their third assessment report, the IPCC declared that in snowmelt-dominated watersheds, adaptation may not fully offset the effects of reduced water availability (Cohen *et al.*, 2001). This poses a significant problem for much of western Canada, where a warming of 1 to 4 degrees Celsius in mean annual temperature has been observed over the last century (Schindler and Donahue, 2006). This warming is most pronounced during the winter and spring seasons (Karl *et al.*, 1993). As a result, an earlier onset of spring (Groisman *et al.*, 1994; Cayan *et al.*, 2000; Stewart *et al.*, 2005), decreased late season snowpack (Selkowitz *et al.*, 2002; Mote, 2003, Hamlet *et al.*, 2005; Mote *et al.*, 2005), and an increase in rain to snow ratios (Knowles *et al.*, 2006) have been observed in

historical climate records. These trends are expected to continue as temperatures increase (Hamlet and Lettenmaier, 1999; Leung *et al.*, 2004; Lapp *et al.*, 2005).

If these trends continue, the seasonal distribution of runoff will be altered, with an increase in winter and early spring runoff and a decrease in late summer (Stewart *et al.*, 2005). These changes in streamflow timing and magnitude have important implications for ecosystems on the Canadian Prairies (Schindler, 2001). Low flows during late summer can have serious implications for aquatic and riparian health if ecosystem water requirements are not met (Schindler and Donahue, 2006). The distribution and community composition of native aquatic species could change (Clair *et al.*, 1998), while increased drought stress could increase riparian forest mortality (Rood *et al.*, 1995).

Human populations will also be affected as agriculture and water supply for human consumption in the Western Prairie provinces are highly dependent on spring runoff that fills storage in reservoirs. Increased aridity as a consequence of climate warming will result in more frequent persistent drought conditions (Gan, 2000; Sauchyn *et al.*, 2003). With spring melt occurring earlier in the year and increased demands for water during the late season when water supply is reduced, the effects of drought will be exacerbated (Schindler and Donahue, 2006). Under these conditions, maintaining streamflow requirements for ecosystem health and supplying water for human use will be difficult. These potential changes must be quantified for appropriate management decisions to be made.

Hydrological models provide the structure to evaluate the relationships between hydrology and climate (Leavesley, 1994). To appropriately model the influence of climate change on mountain hydrology, physically based models that quantify the processes of the hydrological cycle through the application of the physical laws of hydrology are required (Leavesley, 1994; Dingman, 2002). A physical model must be capable of representing the spatial and temporal variation in physiographic characteristics of a watershed, including terrain altered micrometeorology, and associated characteristics of soils and vegetation for modelling in complex mountain terrain (Refsgaard and Knudsen, 1996; Todini, 1988). This thesis describes the ongoing development of a physically based spatial hydrometeorological model for mountainous terrain.

To assess the potential impacts of climate change and other anthropogenic disturbance on mountain regions, reliable climate data are required (Daly *et al.*, 2007). The best climate change information currently available is from GCMs (General Circulation Models), which are mathematical models that represent interactions between the atmosphere, land surface, oceans, and the cryosphere (Karl *et al.*, 2003) at a grid cell resolution of roughly 300 x 300 km (IPCC-TGICA, 1999). GCMs provide reliable estimates of large-scale climate, however, their coarse spatial resolution does not resolve the topographical effects of mountains on climatic processes. Therefore, GCMs are not suited to model fine scale processes in mountainous environments (Leung and Wigmosta, 1999; Hay *et al.*, 2000).

Through a technique known as downscaling, climate change projections from large-scale GCMs can be transferred to the local scale needed to determine the potential effects from climate change on mountain hydrology. The coupling of fine scale hydrological models with coarse scale GCMs through downscaling provides a means of integrating climate change projections into hydrological impacts assessments at an appropriate spatial resolution (Wood *et al.*, 1997). With downscaled projections of future climate, a hydrological model is able to provide physically based estimates of the potential changes in local hydrological conditions. This thesis applies a hydrometeorological model with a range of future GCM derived climate scenarios to assess the potential future snowpack conditions over the St. Mary River watershed on the eastern slopes of the Rockies.

## **1.2 Research Objectives**

1. Model hydrometeorological variables over mountainous terrain in the St. Mary River watershed;
2. Assess potential climate change impacts on mountain snowpack in the St. Mary watershed for a range of GCM derived scenarios using the GENESYS model.

The first objective involved developing and enhancing several new subroutines for the GENESYS (GENerate Earth SYstems Science input) model. The model, previously known as SimGrid (Sheppard, 1996; Lapp *et al.*, 2005) has been modified, improving the representation of terrain related

hydrometeorological processes relevant to mountain environments. The changes include: a seasonally varying precipitation-elevation function; and a reclassification of the watershed using terrain categories derived from elevation and land cover data. Improvements were made to estimate the spatial and temporal variability in canopy interception, sublimation, evapotranspiration, and soil water storage. The GENESYS model performance was assessed by comparing model output to annual streamflow at the watershed outlet gauging station and daily snowpack values at a SNOTEL site within the watershed.

The second objective is to assess changes in mountain snowpack from climate change in the St. Mary watershed using the GENESYS model. A range of future temperature and precipitation estimates are derived using five GCM climate change scenarios. The scenarios selected represent a range of plausible future climates based on IPCC emissions scenarios. Changes in temperature and precipitation from each scenario were applied by downscaling GCM output to the regional scale using the “delta” method. Changes in mean and maximum annual snowpack, the timing of snowmelt, and spatial change in snowpack were assessed for the 2020 (2010-39), 2050 (2040-69), and 2080 (2080-99) time periods relative to 1961-90.

### **1.3 Thesis Structure**

This thesis is presented as 4 chapters. Chapter 1 introduces the study rationale and objectives. Chapter 2 is the first journal paper and describes the enhancements and initial application of the GENESYS model to watershed

hydrometeorology. This chapter has been submitted to the Journal of Hydrometeorology and is currently under review. The third chapter, and second journal paper, applies the GENESYS model to project potential impacts from climate change on mountain snowpack in the St. Mary River watershed. The fourth and final chapter consists of conclusions drawn from the previous two chapters and describes future work that is needed in this area.

## CHAPTER 2

### FURTHER DEVELOPMENT OF A SPATIAL HYDROMETEOROLOGICAL MODEL FOR COMPLEX MOUNTIAN ENVIRONMENTS

#### 2.1 Introduction

Water supply in western North America is highly dependent on mountain ecosystems (Barnett *et al.*, 2005; Field *et al.*, 2007; Mote *et al.*, 2005). Mountain ecosystems are complex; this is largely due to topographical controls on meteorological variables, including air temperature, radiation, saturation deficit, and precipitation (Raupach and Finnigan, 1997). It is this complexity that makes mountain regions extremely vulnerable to changes in climatic processes (Beniston, 2003; Leung and Wigmosta, 1999; McKenzie *et al.*, 2003). In order to assess the effects of anthropogenic disturbance and changing climates on mountain regions, reliable hydrometeorological data are needed (Daly *et al.*, 2007). Unfortunately, measurements of meteorological processes in mountainous regions are sparse and do not represent the spatial and temporal variability required over entire watersheds (Diaz, 2005). It is, therefore, necessary to estimate the spatial and temporal properties of mountain hydrometeorology using detailed spatial models. This chapter describes continued developments of one such model.

A physically based method for estimating daily hydrometeorological variables at representative spatial resolution is presented. The GENESYS (GENerate Earth SYstem Science input) model developed at the University of Lethbridge under the direction of Dr. James Byrne (Sheppard 1996; Lapp *et al.*,

2002) has been applied in several studies (Lapp *et al.*, 2005, Larson 2008; Larson *et al.*, in review) including work described herein. Portions of GENESYS have previously been presented in the literature as the SimGrid microclimate model. GENESYS is a suite of models used to simulate daily hydrometeorological conditions spatially over complex terrain.

The objective of this work is to further develop the GENESYS model for increased application in mountainous environments and to assess the suitability of the model in predicting anthropogenic effects on ecosystem services. Improvements to the precipitation-elevation functions and the determination of a representative spatial scale for model application are shown. To account for a more detailed hydrological balance, routines for estimating spatial and temporal variation in canopy interception were developed, and the sublimation, evapotranspiration, and soil water storage routines were enhanced.

### **2.1.1 Background**

One of the main purposes of the GENESYS model is that it is able to operate at a high spatial resolution. With appropriate spatial input data, GENESYS can account for changes in hydrometeorology by using physical relationships between meteorological variables and physiographical characteristics. It is, however, still important to understand that increased resolution implies increased complexity and not necessarily a higher degree of accuracy (Daly, 2006). Therefore, the explanation of hydrometeorological variables at an appropriate spatial scale in mountainous terrain is important.

It is difficult to determine what an appropriate scale might be due to the complexities in ecosystem responses to the hydrological cycle (Martin, 1993). We suggest that using these ecosystem responses is likely the best means to determine the appropriate scale at which to model. Vegetative cover is highly dependent on hydrometeorological conditions (Mather and Yoshioka, 1968; Stephenson, 1990), and therefore, provides an ecologically sensitive surrogate for the spatial characteristics of hydrometeorological variables. In this chapter, we integrate the spatial distribution of land cover to represent spatial variability in mountain hydrometeorological conditions. This is achieved using a GIS overlay analysis of topography and land cover to create grids that are used to provide the spatial inputs for the GENESYS model.

The incorporation of hydrological processes relevant to mountainous environments is essential for environmental modelling (Woo and Marsh, 2005). Of the hydrological variables, precipitation is perhaps the most difficult to quantify spatially. By accounting for the physiographical controls on spatial and temporal distribution of precipitation and incorporating observed meteorological data, estimates of precipitation can be made for mountainous environments (e.g. Daly *et al.*, 2002). This paper describes how the GENESYS model estimates spatial and temporal variability in precipitation by adjusting for elevation and season.

Sublimation of snow represents a fundamental component of the mountain winter hydrologic balance (Pomeroy *et al.*, 1998b). Strasser *et al.* (2007) suggest that 15 to 90 % of winter snow pack in mountainous environments is lost to sublimation depending on exposure. Pomeroy *et al.* (1999) have shown that

sublimation can account for up to 45% of the annual snow fall in spruce forests of the Wolf Creek basin, Yukon. Hood *et al.* (1999) found that during the 1994-1995 snow year, sublimation loss accounted for 15% of the total accumulated snow pack on the Niwot Ridge, Colorado. Zhang *et al.* (2004) found similar results in the southern mountain Taiga of Siberia. Dery *et al.* (1998) have developed a sublimation model that is adopted to estimate sublimation in GENESYS.

The interception of snow and rain by the forest canopy can have a significant effect on the annual hydrological balance (Gelfan *et al.*, 2004). Coniferous canopies in particular have been shown to intercept up to 60% of the total annual snow fall, resulting in a loss of roughly 40% of the annual snow fall from the canopy (Pomeroy and Gray, 1995). Rainfall interception has also been shown to account for up to 50% of rainfall (Calder, 1990). In order to simulate and subsequently predict changes in the hydrologic balance of forested regions, canopy interception must be included (Hedstrom and Pomeroy, 1998).

Processes affecting interception are highly dependent on the physiographical and meteorological characteristics of a region (Elder *et al.*, 1991). Therefore, the selection of an appropriate interception model is very important. An empirical rainfall interception routine based on canopy leaf area index (LAI) was adapted for GENESYS (Von Hoyningenuene, 1983). In cold regions and mountainous regions, snow accounts for a large portion of the total precipitation and can remain in the canopy for several days (Hedstrom and Pomeroy, 1998). The canopy snow interception model developed by Hedstrom

and Pomeroy (1998) for the southern boreal forests of western Canada provides physically based estimates of canopy load, interception, and unloading. This model is adopted as a subroutine in the GENESYS model. This adaptation assumes that the cold climatic regime and physical properties of tree species of the boreal forest represent the east slopes of the Rocky Mountains well enough to provide realistic estimates of interception.

Evapotranspiration (ET) also plays an important role in ecosystem function as soil moisture, vegetation productivity, and the hydrological balance are all affected by ET (Wever *et al.*, 2002). Unfortunately, measurements of ET in mountainous environments are generally unavailable and are not representative spatially (Diaz, 2005). To make estimates of ET over space, a number of equations have been developed using relationships with meteorological data (Valiantzas, 2006). The Penman-Monteith equation integrates wind, air temperature, humidity and radiation to provide estimates of potential evapotranspiration (ETP). This chapter presents the addition of a Penman-Monteith evapotranspiration routine to GENESYS.

Accounting for ET enables estimates of soil moisture to be made over the watershed. Soil moisture is an important source of water for ecosystem function that is not well monitored over space or time (Oglesby and Erickson, 1989; Robock *et al.*, 2000). The inclusion of soil water processes in hydrological models enables the prediction of ecosystem response to changing meteorological conditions, allowing for more or less plant available water and runoff (Henderson-Sellers, 1996). Perhaps more importantly, soil moisture

monitoring can provide insight into the occurrence of drought (Dai *et al.*, 2004). A daily soil water budgeting routine was developed for GENESYS. The soil water routine includes estimates of ETP, ET, soil water storage, and controls on the ET/ETP ratio where water supply limits ET. Estimates of the spatial variation in soil water processes were enabled using the SSURGA soils database (NRCS, 2007).

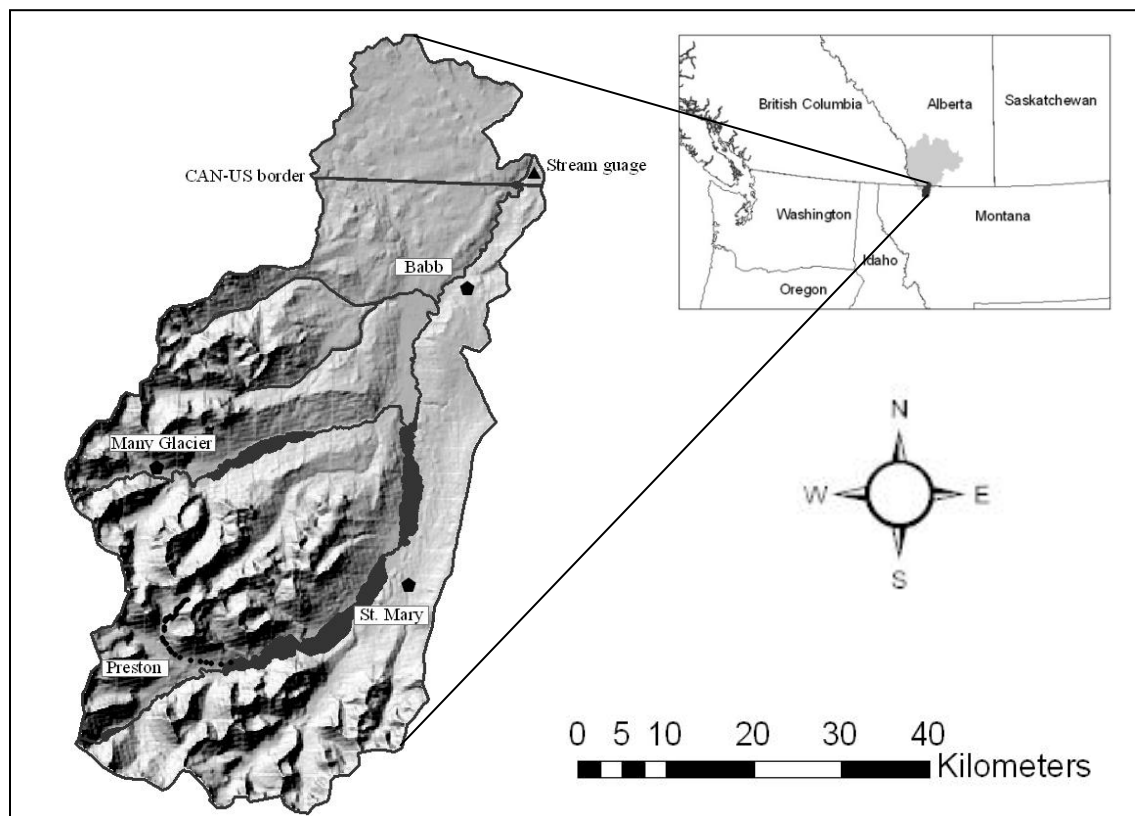
Improvements to the GENESYS model are evaluated for the St. Mary watershed in northern Montana. Daily simulations for 30 years were conducted. The suitability of the spatial surfaces produced by GENESYS is assessed using observed data within the drainage basin. Daly (2006) suggests that the best methods for assessing model estimates are those that are independent from the data and model used to derive the spatial dataset. Stream flow provides an integration of precipitation over entire drainage basins, representing regional precipitation patterns (Rood *et al.*, 2005). Using annual stream flow as an assessment of spatial data quality enables an objective model evaluation, in particular for precipitation estimates (Milewska *et al.*, 2005). To evaluate the daily winter hydrological balance simulated by GENESYS, model output is compared to daily data from a SNOTEL site within the watershed for the 10 year test period. This comparison provides insight into the applicability of the model at a daily time step.

Based on our results we suggest that the improved GENESYS Model represents the hydrological cycle at an appropriate spatial and temporal scale to permit model application in predicting and monitoring changes in hydrological

processes in complex environments. We also show that with improved data for model development and calibration that GENESYS may be applied to estimate impact from environmental change on other ecosystem services.

## 2.2 Study area

The headwaters of the St. Mary watershed lie on the eastern slopes of the Rocky Mountains with the majority of the upper watershed residing within Glacier National Park, Montana. The St. Mary River flows from the continental divide, through the upper and lower St. Mary lakes, and ends in southern Alberta where it meets the Oldman River (Figure 1).



**Figure 1 The St. Mary Watershed in Montana and Southern Alberta. The locations of the watershed stream gauging point, Preston snow survey, Many Glacier SNOTEL site and St. Mary and Babb weather stations are shown.**

The climatic regime is a transitional zone between coastal and continental climates. The region is also influenced by the orographic effect which is most noticeable during the winter months as synoptic conditions dominate (Hanson, 1982). The area receives the majority of its precipitation in the winter, with snow accounting for roughly 70% of the annual precipitation at high elevations (Selkowitz, et al., 2002).

The total drainage area of the study watershed is 1195 km<sup>2</sup>, with a mean elevation of 1745 m, ranging from 1249 m to 3031 m. The area is a relatively undisturbed, ecologically diverse region, this is largely attributed to the fact that a large portion of the drainage area is within Glacier National Park. Coniferous forests account for 24% of the land cover while deciduous trees account for 21% and herbaceous plants cover another 29% of the area. 23% of the area is barren rock or soil, and 3% of the area is water (USGS, 2000).

### **2.3 Model description**

The GENESYS model requires two separate processes to estimate spatial variability in mountain hydrometeorology. The first process involves the spatial estimation of hydro-meteorological variables. This is done using a GIS to derive modelling units over space referred to as Terrain Categories (TCs). For each TC, the SimGrid subroutine (Sheppard, 1996) is used to provide spatial estimates of temperature, precipitation, solar radiation, and relative humidity. The results of these first steps are a number of detailed spatial surfaces of hydrometeorological variables that can be used in further modelling.

The second process uses spatial hydrometeorological estimates from SimGrid to model the hydrological balance at each TC using the snow pack and soil water sub-model. This sub-model is a series of physically based equations that are used to represent mountain hydrometeorological conditions, varying both spatially and temporally.

### **2.3.1 Derivation of Terrain Categories**

Previous versions of TC delineation were derived using a combination of slope, aspect and elevation that resulted in a large number of random, discontinuous categories. A new system of classification is developed herein using land cover and elevation.

A land cover grid derived using Landsat imagery (USGS, 2000) was overlaid with a 100 meter digital elevation model (DEM) to determine TCs for the St. Mary basin. The land cover grid consisted of 9 categories: dry herbaceous, mesic herbaceous, deciduous trees/shrubs, coniferous trees/shrubs, conifer trees/open, water, snow, barren rock/soil, and shadows. Snow and shadow classes were eliminated from the land cover grid and assigned the values of the nearest land cover. The combination of elevation and land cover resulted in a number of TCs that represented less than 1% of the entire watershed. These TCs were joined to neighboring TCs to create 82 TCs which have greater contiguity than the previous method and maintain similar elevation, slope, and aspect properties. Mean values of the latter three variables are derived for each terrain category and used for input to the SimGrid model.

### 2.3.2 MTCLIM

To spatially represent hydro-climatic variables, the MTCLIM model (Hungerford *et al.*, 1989) is looped for each TC in the SimGrid subroutine. The MTCLIM model was developed to estimate climate data for discrete locations in mountainous environments based on the relationships between recorded climate data at valley stations and individual SITES (Hungerford *et al.*, 1989). MTCLIM uses two types of climatological logic; a topographic logic that determines hydro-meteorological variables by extrapolating data from a BASE climate station to the SITE, and a diurnal climatology that derives additional information from BASE input data (Hungerford *et al.*, 1989). The diurnal climatology in MTCLIM generates incident solar radiation and relative humidity, while the topographic logic extrapolates BASE data to make estimations of maximum and minimum air temperature and precipitation (Glassy and Running, 1994). MTCLIM can be driven by any BASE station that provides maximum and minimum temperatures and precipitation and physiographic information for a given SITE.

In GENESYS terminology SITES are referred to as TCs. Input data for each TC includes mean elevation, mean slope, mean aspect, mean monthly precipitation, and mean annual LAI. Variables set as constants over all TCs are surface albedo, atmospheric transmissivity, and minimum and maximum temperature lapse rates. Hydroclimatic variables are determined using these input variables and a series of physically based subroutines (Hungerford *et al.*, 1989; Glassy and Running, 1994).

### **2.3.2.1 Solar Radiation**

An algorithm relating diurnal temperature amplitude and atmospheric conditions is used to calculate incoming solar radiation. The generalized algorithm is:

$$Q_s = I_s + D_s, \quad (1)$$

where,  $Q_s$  is the total incoming solar radiation ( $\text{KJm}^{-2}$ ) and  $I_s$  is the direct beam radiation, and  $D_s$  is the diffuse radiation at the earth's surface. All variables calculated within MTCLIM assume clear sky atmospheric transmissivity is equal to 0.65 at sea level and increases by 0.008/m in elevation. Daily atmospheric transmissivity is adjusted using the diurnal temperature range at the BASE station to reflect the roles of clouds and water vapor in reducing the clear sky transmissivity (Bristow and Campbell, 1984).

### **2.3.2.2 Temperature**

Surface maximum, minimum, and daylight average temperatures are calculated daily for each TC using maximum and minimum temperature lapse rates. Lapse rates for maximum and minimum temperatures used are  $6.1^\circ\text{C}/\text{km}$  and  $5.9^\circ\text{C}/\text{km}$ , respectively. These lapse rates were derived by Pigeon and Jiskoot (2008, in press) through analysis of daily temperature data at Castle Mountain ski resort, approximately 100 km northwest of St. Mary basin.

Maximum temperatures are also adjusted to account for solar radiation (refer to Hungerford *et al.*, 1989 and Sheppard, 1996). This paper has not directly verified

temperature predictions, however, Larson (2008) showed that the temperature estimates from GENESYS compared very well to an alpine site near Waterton, Alberta. We assume that temperature estimates are adequate for the current investigation.

### **2.3.2.3 Humidity**

Daily relative humidity values are calculated from vapour pressure estimates based on dew point temperature ( $T_d$ ) and daylight average temperature ( $T_{avg}$ ).  $T_d$  values are not available for the current study; therefore, daily minimum temperatures are used as a proxy for  $T_d$  (Sheppard, 1996).

Relative humidity at each TC ( $RH_{tc}$ ) as a percentage is determined by:

$$RH_{tc} = \left( \frac{es}{esd} \right) * 100, \quad (2)$$

where,  $esd$  is saturated vapour pressure (kPa) and  $es$  is ambient vapour pressure (kPa). Vapour pressure estimates are computed using

$$\text{Vapour pressure} = 0.61078 * e^{\left( \frac{17.269 * T}{237.3 + T} \right)}, \quad (3)$$

where,  $T$  is either average daylight temperature or dew point temperature to estimate  $esd$  and  $es$  respectively (Hungerford *et al.*, 1989).

### **2.3.2.3 Precipitation**

MTCLIM logic uses a ratio between mean monthly precipitation values at each TC and the BASE station to adjust for changes in precipitation as a function of elevation and season. To make these adjustments an analysis of recorded

precipitation data within the St. Mary watershed was conducted. Data used for this investigation included the Preston snow survey (Fagre, 2006), the Many Glacier SNOTEL site (NRCS, 2007) and the St. Mary climate station (NOAA/NCDC, 2006). A series of steps were used to derive the precipitation values at each TC:

1. Derive a precipitation-elevation relationship
2. Adjust for seasonality differences between the BASE station and the mountain portions of the watershed
3. Calculate ratios between monthly mean values at the BASE station and each TC
4. Apply ratios to daily precipitation data

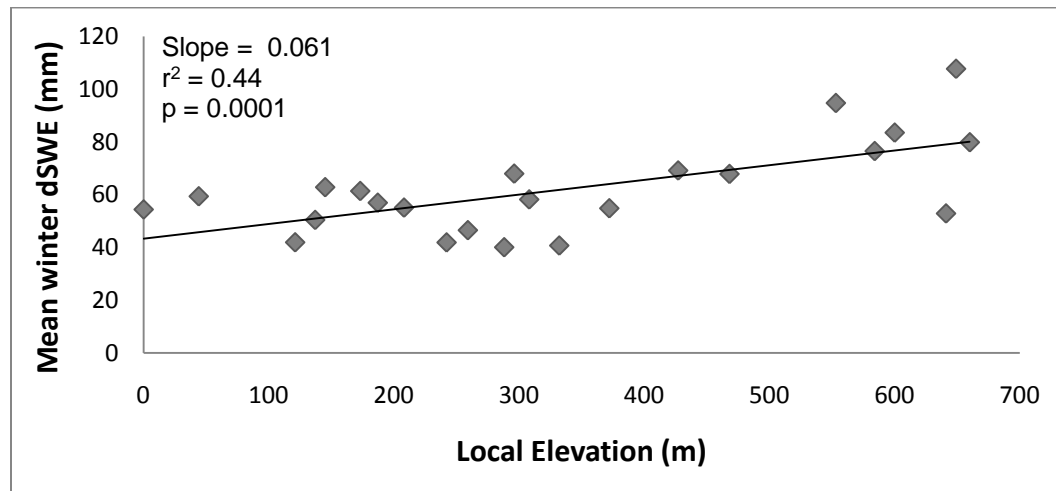
#### **2.3.2.3.1 *Precipitation-elevation relationship***

SWE data from the Preston snow survey were used to derive a precipitation-elevation relationship that will enable more accurate representation of the orographic effects on precipitation. The Preston snow survey (Figure 1) began in 1994, and continues to the present, operated by the United States Geological Survey (USGS). The survey has 32 sampling points located near the center of the watershed and spans an elevation range from 1438 m to 2290 m. SWE data have been acquired from the inception of the survey to the end of the 2006 snow year (Fagre, 2006).

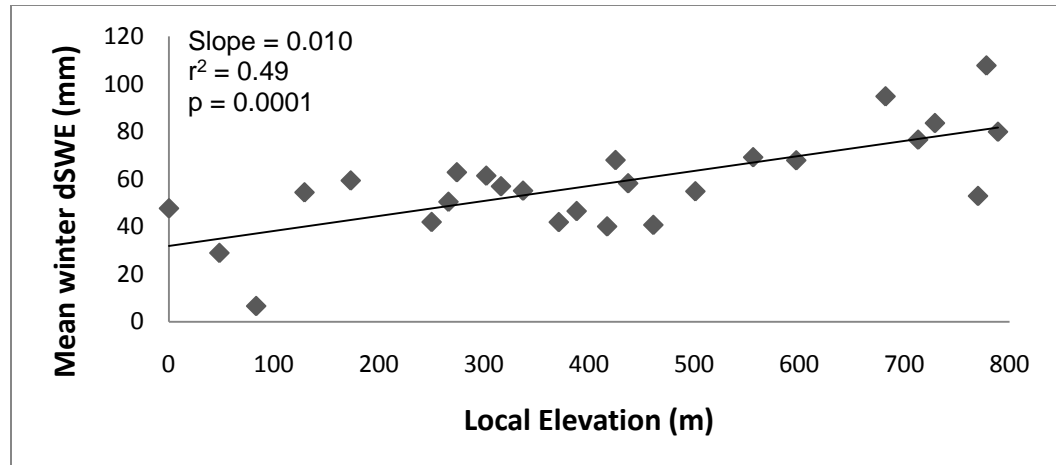
A change in winter SWE ( $\Delta$ SWE) was calculated for each monthly sampling interval for 73 months (Larson, 2008). The mean winter  $\Delta$ SWE was then determined for all sites. Sampling points were removed if they had negative

mean monthly  $\Delta$ SWE values, resulting in a total of 26 sampling points used to derive 2 precipitation-elevation relationships.

The two precipitation-elevation relationships are derived relative to the Many Glacier SNOTEL site and the St. Mary climate station. The SNOTEL site is located in a small meadow surrounded by trees at an elevation of 1519 m in the western portion of the watershed (NRCS, 2007). The St. Mary station is at an elevation of 1390 m, located near St. Mary, Montana, in the eastern portion of the watershed (NOAA/NCDC, 2006). Using precipitation-elevation relationships relative to both St. Mary and Many Glacier as BASE stations enables the model to account for seasonality differences between mountain and low land areas. The resultant relationships between local elevation and mean winter  $\Delta$ SWE are presented in Figures 2 and 3.



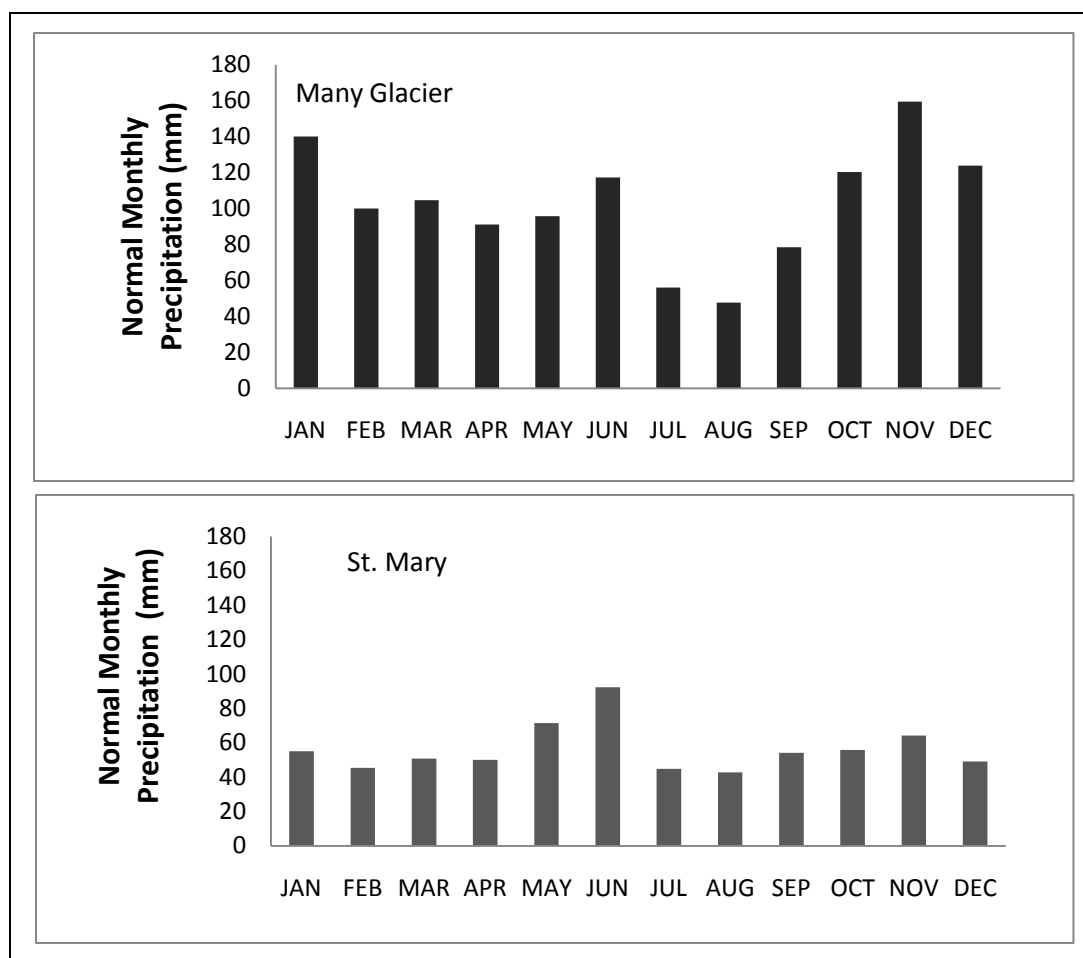
**Figure 2 Linear relationship between local elevation in relation to the Man Glacier SNOTEL site and dSWE at the Preston snow course.**



**Figure 3 Linear relationship between local elevation in relation to St. Mary and dSWE at the Preston snow course.**

#### **2.3.2.3.2 Adjusting for seasonality of precipitation**

Figure 4 shows that the seasonal distribution of precipitation differs significantly between the St. Mary climate station and Many Glacier SNOTEL site ( $r^2=0.09$ ,  $p=0.17$ ). This is consistent with the fact that mountainous regions experience very different hydrometeorological conditions relative to low elevation prairie-transitional zones (Bales *et al.*, 2006).



**Figure 4 Seasonal distribution of precipitation at Many Glacier and St. Mary.**

An adjustment was made to account for the seasonal differences in precipitation between the Many Glacier SNOTEL site and the St. Mary climate station, resulting in a better representation of precipitation over the watershed. Monthly relationships between St. Mary and Many Glacier precipitation means for the years 1982 to 2005 were derived using linear regression (Table 1).

**Table 1 Linear relationships between mean monthly precipitation at St. Mary and mean monthly precipitation at Many Glacier (n=23).**

MONTH	Mountain BASE EQUATION	r2	p
JAN	MG = 1.526(STM) + 58.575	0.83	<0.0001
FEB	MG = 1.505(STM) + 29.284	0.77	<0.0001
MAR	MG = 1.587(STM) + 28.825	0.82	<0.0001
APR	MG = 1.233(STM) + 28.189	0.57	<0.0001
MAY	MG = 0.549(STM) + 55.335	0.29	0.006
JUN	MG = 0.863(STM) + 33.151	0.83	<0.0001
JUL	MG = 0.735(STM) + 26.554	0.85	<0.0001
AUG	MG = 0.875(STM) + 13.842	0.72	<0.0001
SEP	MG = 0.904(STM) + 31.279	0.53	<0.0001
OCT	MG = 1.875(STM) + 17.199	0.84	<0.0001
NOV	MG = 1.797(STM) + 45.295	0.89	<0.0001
DEC	MG = 1.570(STM) + 46.428	0.87	<0.0001

Based on the regression results, we suggest that the St. Mary climate station can be used to predict monthly precipitation at Many Glacier, especially during the winter. This enables the model to account for changes in seasonality using a single low elevation BASE station, where the longest climate records are available.

### **2.3.2.3.3 Determination of monthly mean precipitation values**

To determine monthly precipitation means the two  $\Delta$ SWE equations are used. For TC elevations below 1500 m, the Preston  $\Delta$ SWE relationship relative to St. Mary (Figure 3) is applied with the monthly mean values from St. Mary climate station as the BASE (equation 4).

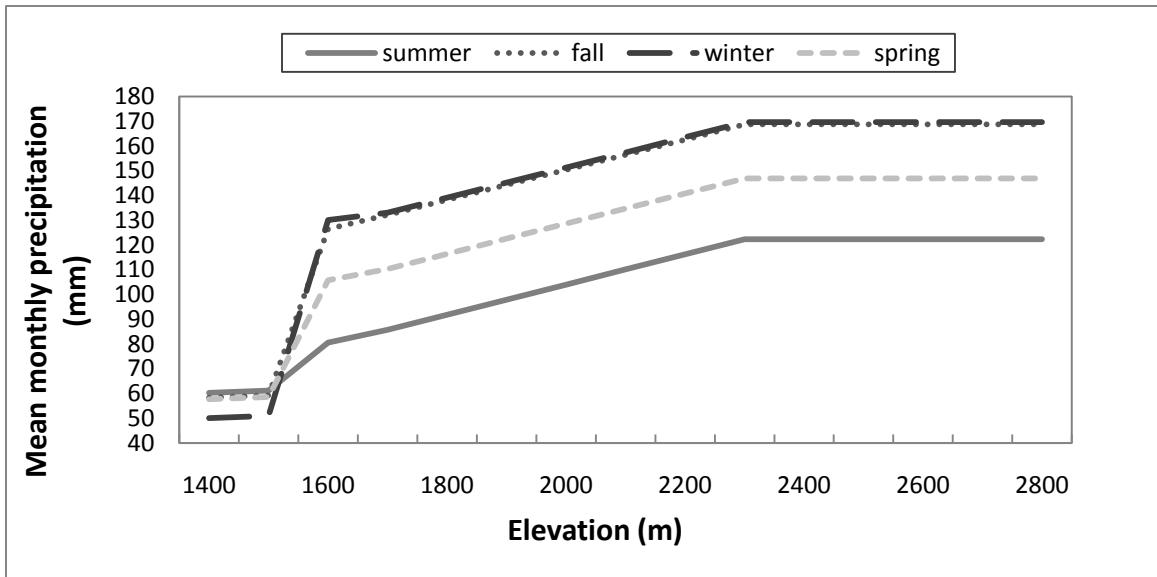
$$SWE_{(<1500)} = 0.010(\text{local elevation}) + \text{BASE} \quad (4)$$

If the TC elevation is above 1500m, monthly mean precipitation values for a mountain BASE are calculated using monthly relationships presented in table

1. Using the predicted mountain BASE, the Preston  $\Delta$ SWE relationship relative to Many Glacier (Figure 2) is applied (equation 5).

$$SWE_{(>1500)} = 0.061(\text{local elevation}) + \text{mountain BASE} \quad (5)$$

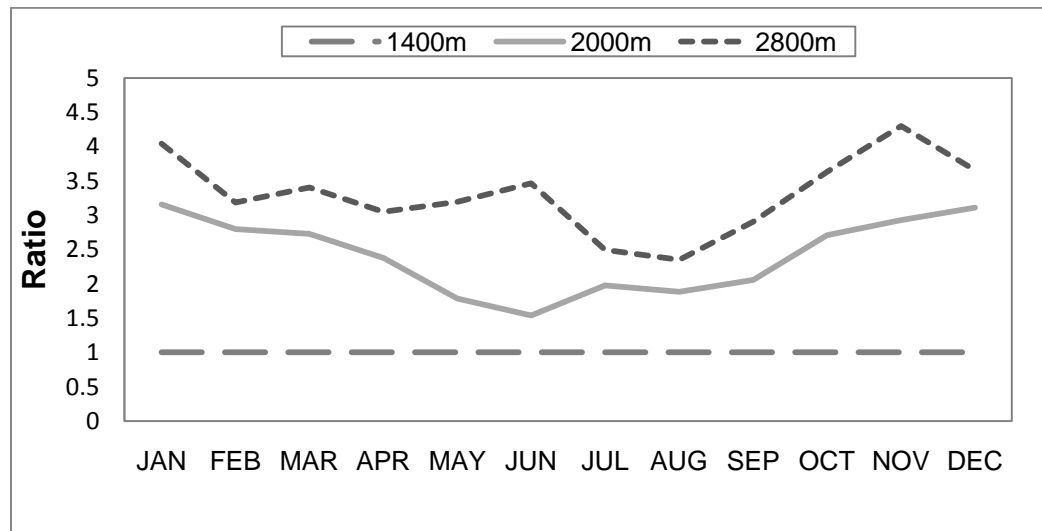
This allows for a seasonal shift in precipitation to be made for the mountainous portion of the watershed, while accounting for the effects of elevation. At elevations greater than 2300m (extent of the snow course data), monthly means are assigned the same value as the mean at 2300m, resulting in no change in SWE with elevation above 2300m. Figure 5 shows how simulated monthly mean precipitation volumes change with season and elevation.



**Figure 5 Mean monthly precipitation change as a function of season and elevation.**

#### 2.3.2.3.4 Application to daily precipitation data

In order to apply the effect of elevation and shift in seasonality to the daily historical precipitation record, ratios are calculated between monthly precipitation means at the St. Mary climate station and the monthly means at each TC. This results in the largest ratios during the winter and smallest ratios during the summer (Figure 6).



**Figure 6 Varying monthly mean precipitation ratios between St. Mary and TCs at the elevation bands 1400m, 2000m and 2800m.**

These ratios are multiplied by daily precipitation volumes at the St. Mary climate station, resulting in precipitation volumes that are adjusted as a function of elevation and season over the watershed.

### 2.3.3 Snow pack and soil water model

With the snow pack and soil water model, daily spatial hydro meteorological data are used to simulate changes in TC SWE. When snowpack is present a daily winter hydrological balance is calculated,

$$SWE_{(t)} = SWE_{(t-1)} + P_{(t)} - I_{(t)} - S_{(t)} - R_{(t)} - IF_{(t)}, \quad (6)$$

where,  $SWE$  is the amount of snow water equivalent (mm) in the snow pack,  $P$  is simulated daily precipitation as rain or snow,  $I$  is canopy interception,  $S$  is sublimation,  $R$  is runoff,  $IF$  is infiltration and  $t$  is the time step (days).

If the snowpack is completely gone, a hydrological balance is calculated that accounts for evapotranspiration ( $ET$ ) and changes in soil moisture ( $SM$ ) conditions,

$$SM_{(t)} = SM_{(t-1)} + P_{(t)} - I_{(t)} - ET_{(t)} - R_{(t)}, \quad (7)$$

allowing for two separate hydrological balances to be calculated dependent on surface conditions enables processes affecting water supply in the basin to vary both spatially and temporally.

#### 2.3.3.1 Precipitation Partitioning

The method used to partition rain and snow was developed by Kienzle (2008). This method uses an S shaped curve and two temperature variables:

$$Pr = 5 * \left( \frac{T-Tt}{1.4*Tr} \right)^2 + 6.76 * \left( \frac{T-Tt}{1.4*Tr} \right) + 3.19 * \left( \frac{T-Tt}{1.4*Tr} \right) + 0.5, \quad (8)$$

where,  $Pr$  is the proportion of precipitation that falls as rain (range from 0 to 1),  $T$  is the mean daily temperature,  $Tt$  is the threshold mean daily temperature and  $Tr$  is the range of temperatures where both rain and snow can occur. Default values for  $Tt$  and  $Tr$  are 2.6 °C and 13.3 °C respectively as suggested by Kienzle (2008).

### **2.3.3.2 Canopy Interception**

Snow ( $SnowInt$ ) (mmSWE) and rain ( $RainInt$ ) (mm) interception by the canopy are determined for each terrain category containing coniferous forests. Rain interception is only calculated for deciduous forests because a subroutine was not available that accounts for snow interception in deciduous forests.

#### **2.3.3.2.1 Snow interception**

The snow interception ( $snowInt$ ) sub-routine uses a formula derived by Hedstrom and Pomeroy (1998):

$$SnowInt = I * 0.678 \quad (9)$$

where, 0.678 is a determined unloading coefficient for pine and spruce forests and  $I$  is the intercepted snow load at the start of unloading (Hedstrom and Pomeroy, 1998),  $I$  is determined using the following:

$$I = (L - Load) * (1 - 10^{-k*P}) \quad (10)$$

where,  $Load$  is the total snow load in the canopy on the previous day ( $Kgm^{-2}$ ),  $L$  is the maximum load for the given canopy given the boundary layer conditions

( $\text{Kg m}^{-2}$ ),  $P$  is the amount of precipitation falling as snow. An approach is taken for the determination of  $K$  (the proportionality factor), where it is assumed that there is a closed canopy and interception is completely efficient. It is also assumed that snowflakes are falling vertically on the canopy. The following formula represents  $K$  when the preceding assumptions are made:

$$k = \frac{1}{L} \quad (11)$$

where,

$$L = S * LAI \quad (12)$$

$LAI$  is set to 2 to represent a mean value for coniferous forests and  $S$  is the species value at a given density,  $S$  is determined using the following:

$$S = SV * \left(0.27 + \left(\frac{46}{P_s}\right)\right) \quad (13)$$

$SV$  is a constant of  $5.9 \text{ kg m}^{-2}$  derived by Hedstrom and Pomeroy (1998) for spruce forests.  $P_s$  ( $\text{Kg m}^{-3}$ ) is the density of snow calculated as a function of mean air temperature (Hedstrom and Pomeroy, 1998).

The canopy load is calculated by adding each interception event to the canopy store and subtracting snow that is sublimated. The canopy is able to store snow until  $L$  is reached, at which time the remaining snow in the canopy will fall to the ground and is then incorporated in the snowpack.

### 2.3.3.2.2 Rain interception

$RainInt$  is calculated using the Von Hoyningenuene (1983) formula that calculates interception as a function of total rainfall and  $LAI$ .  $RainInt$  is calculated on a daily basis using the following function:

$$RainInt = 0.30 + 0.27Rain + 0.13LAI - 0.013Rain * LAI - 0.007LAI^2 \quad (14)$$

where,  $Rain$  is the precipitation that falls as rain. Von Hoyningenuene (1983) found that this function is unstable at precipitation values greater than 18 mm, resulting in anomalous interception values. Therefore, daily precipitation is set to a maximum of 18 mm for this calculation; all other precipitation is considered as through fall. The intercepted rain store is determined for each day by adding the intercepted rain to the store and subtracting evaporated and transpired water from the canopy.

### 2.3.3.3 Sublimation

The vapour transfer model developed by Thorpe and Mason (1966) and later modified by Dery *et al.* (1998) is used to estimate daily sublimation losses as a function of snow properties and atmospheric conditions. Total sublimation loss is estimated by:

$$Q_{subl} = \frac{dm}{dt} N_{(z)}, \quad (15)$$

where,  $Q_{subl}$  ( $\text{kg m}^{-2}\text{s}^{-1}$ ) is the sublimation rate for a column of blowing snow over a horizontal land surface,  $dm/dt$  is the change in mass of a blowing snow

particle due to sublimation per second, and  $N(z)$  is the number of snow particles per unit volume ( $\text{m}^{-3}$ ). The number of snow particles depends on the particle shape  $\alpha$  and radius  $r$ . A mean  $\alpha = 5$  and  $r = 100 \mu\text{m}$  (Pomeroy and Male, 1998) are used. When using a  $\alpha = 5$ , Dery *et al.* (1998) suggest that  $N_{(z)} = 9.09 \cdot 10^7 \text{m}^{-3}$ . Thorpe and Mason, (1966) estimate the change in mass of a blowing snow particle by:

$$\frac{dm}{dt} = \frac{2\pi r \sigma \frac{Qr}{CN_{Nu}Ta} \left( \frac{Ls}{RvTa} - 1 \right)}{\frac{Ls}{CN_{Nu}Ta} \left( \frac{Ls}{RvTa} - 1 \right) + Rv \frac{Ta}{N_{sh} D_{ei}}}, \quad (16)$$

where,  $2\pi (m)$  is the area function of a snow particle,  $\sigma$  is the water vapor deficit :  $ei$  is the vapour pressure value at saturation over ice;  $Ta$  is air temperature (K);  $Qr$ , the radiation transferred to the particle is  $Qr = \pi r^2 (1 - \alpha_p) Q_*$  (Schmidt, 1991), with  $\alpha_p$  the shortwave particle albedo (0.5: Schmidt *et al.*, 1998) and  $Q_*$  ( $\text{Wm}^{-2}$ ) the total incident radiation;  $C$  is the thermal conductivity of air ( $2.4 \cdot 10^{-2} \text{Wm}^{-1}\text{K}^{-1}$ );  $Ls$  is the latent heat of sublimation ( $2.838 \cdot 10^6 \text{J Kg}^{-1}$ );  $Rv$  is the gas constant for water vapor ( $461.5 \text{J Kg}^{-1} \text{K}^{-1}$ ); and  $D$  is the molecular diffusivity of water vapor in air ( $2.25 \cdot 10^{-5} \text{m}^2 \text{s}^{-1}$ );  $N_{Nu}$  and  $N_{sh}$  are Nusselt and Sherwood numbers respectively:

$$N_{Nu} \text{ and } N_{sh} = 1.79 + (0.606 N_{re}^{0.5}), \quad (17)$$

where, the Reynolds number  $N_{re} = (2rVr/V)$ ;  $Vr$  is the ventilation velocity, assumed to equal horizontal wind speed (Schmidt, 1982; Dery and Taylor, 1996), and  $V$  is the kinematic viscosity of air ( $1.53 \cdot 10^{-5} \text{m}^2 \text{s}^{-1}$ ; Dery and Yau, 1999).

The adopted sublimation model requires 3 environmental inputs calculated by the MTCLIM subroutine; incident solar radiation ( $\text{Wm}^{-2}$ ); ambient air temperature (K); and relative humidity (%). Lapp *et al.* (2005) have shown the sensitivity of this sublimation routine to changes in the 3 environmental inputs. Lapp *et al.* (2005) have also shown the influence of wind on sublimation rates, wind data were not available for this study, therefore, the assumption remains that  $V_r$  approximates horizontal wind speed.

#### **2.3.3.4 Snow Melt**

The snow melt routine was adopted from Quick and Pipes (1977). In order for snow melt to occur, the snow pack must ripen; this is determined by the ability of the snowpack to store cold. The snow pack cold storage ( $TREQ$ ) is determined by:

$$TREQ_i = (MLTF * TREQ_{i-1}) + Tmean_i, \quad (18)$$

where,  $MLTF$  is a decay constant (set to 0.85) and  $Tmean$  is the mean daily temperature. When  $TREQ$  reaches 0 (enough energy is absorbed and the snowpack is ripe), melt can occur.

Daily snow melt ( $M$ ) (mm) values are calculated as a function of air temperature.

$$M = PTM * (Tmax + TCEADJ * Tmin), \quad (19)$$

where,  $PTM$  is a point melt factor ( $\text{mm/day}^\circ\text{C}$ ) and  $TCEADJ$  is an energy partition multiplier. Wyman, (1995) suggests a  $PTM$  of 1.8 for the Canadian

Rocky Mountains. However, using the timing and rate of melt at the Many Glacier SNOTEL site to calibrate *PTM* suggest a *PTM* of 1.0 is more suited to this area.

**Table 2 Comparison of date of complete melt at different point melt factors (PTM).**

Year	JDay of complete melt		
	observed	Simulated (PTM 1.0)	Simulated (PTM 1.8)
1977	134	146	125
1978	139	145	133
1979	121	127	117
1980	117	86	86
1981	142	144	135
1982	127	120	107
1983	120	120	97
1984	127	126	117
1985	105	104	81
1986	114	110	96

The rate of melt did not change significantly with changes in *PTM*; however, the timing of complete melt showed that a *PTM* of 1.0 provides the most suitable simulation of the date of complete melt, especially in years where complete melt occurs earlier (Table 2).

### **2.3.3.5 Evapotranspiration**

Daily *ETP* estimates are made only when snowpack is depleted. When the snowpack remains, sublimation is calculated. A version of the Penman-Monteith evapotranspiration equation developed by Valiantzas (2006) is adopted:

$$ETP = 0.038R_s\sqrt{T_{mean} + 9.5} - 2.4\left(\frac{R_s}{R_a}\right)^2 + 0.075(T_{mean} + 20)\left(1 - \frac{Rh}{100}\right), \quad (20)$$

where,  $R_a$  is extraterrestrial radiation,  $R_s$  is solar radiation and  $Rh$  is relative humidity. This equation is used because it does not require wind data and has

been shown to provide accurate estimates of evapotranspiration when compared with the standardized FAO-56 Penman-Monteith scheme using a global climatic dataset (Valiantzas, 2006). *ET* estimates are assumed to equal *ETP*, however, are limited by water supply. For any TC, if soil water declines below half of storage capacity a restricting equation we have developed is applied:

$$XK = \left(2 * \frac{SM}{SoilMax}\right)^{1.5}, \quad (21)$$

where, *XK* is the water supply control on *ET*, and *SoilMax* is the soil field capacity for a particular TC. When *SM* is half of the storage capacity *ET* is determined by

$$ET = ETP * XK. \quad (22)$$

### **2.3.3.6 Soil Moisture**

In order to account for changes in soil moisture, soils data for each TC are required. Soils data were only available for the eastern portion of the St. Mary watershed; therefore, land cover was used as a surrogate for soils in the western portion of the watershed. Using the same land cover grid as was used for TC delineation; relationships between land cover type and soil type were used to extrapolate soil depth and water holding capacity values from the eastern portion of the watershed to the entire basin.

Mean soil depth and field capacity values from each soil type were used for the analysis. A GIS overlay analysis, involving land cover, mean soil depth, and mean field capacity grids was used to extract mean soil depth and field capacity for each land cover type.

Soils data were sparse in the eastern portion of the watershed above an elevation of 2700 m and a slope of 45 degrees. An analysis of land cover data showed that although present, the density of vegetative cover is also low above this elevation and slope. It was, therefore, assumed that elevations above 2700 m and steeper than 45 degree slopes do not have significant water storage capacity. Therefore, the constraints above 2700 m, and more than 45 degrees were included in the analysis and assigned values of the barren rock/soil land cover type.

For each TC mean soil depth and field capacity values were determined. This enables the spatial representation of maximum soil field capacity over the entire drainage. Field capacity values over the watershed range from 0.0 to 199.4 mm, accounting for soil depth. The values associated with each TC are then used in the model to estimate changes to daily soil water loss through ET and gain from snow melt and rain as presented in equation 7.

#### **2.3.4 Model Application and Validation**

To simulate a daily hydro meteorological balance over the St. Mary watershed, daily temperature and precipitation data from the St. Mary climate station are used. Missing records from the years 1960 to 2000 were infilled by Larson (2008) using nearby climate stations and linear regression. The complete time series is divided into sections for modelling purposes, the 1961 to 1990 section was the initial run, used for model validation, and the 1961 to 2000 run is used for a trend analysis of April 1 SWE and August 31 soil water.

Due to the lack of spatially distributed hydrometeorological data over the region, direct validation of spatial variables simulated by GENESYS is not possible. To validate outputs, a measurable variable must be derived from the hydrometeorological surfaces. The validation uses a mass balance approach by comparing total annual volume of runoff from GENESYS and naturalized annual stream flow volume at the international border gauging station 05AE027 (Alberta Environment, 1998).

To evaluate daily snowpack values predicted by GENESYS a comparison is made between the Many Glacier SNOTEL site and simulated values at the TC representing the SNOTEL site. It is recognized that the SNOTEL site is used in developing the monthly precipitation-elevation relationships; therefore, precipitation volume estimates are biased. However, the objective of the comparison is to test the method of applying monthly relationships to daily data and to assess the ability of GENESYS to account for snow accumulation and ablation processes at the daily time step. The Many Glacier SNOTEL site is also the only site within the basin that could be used for this analysis.

## **2.4 Results**

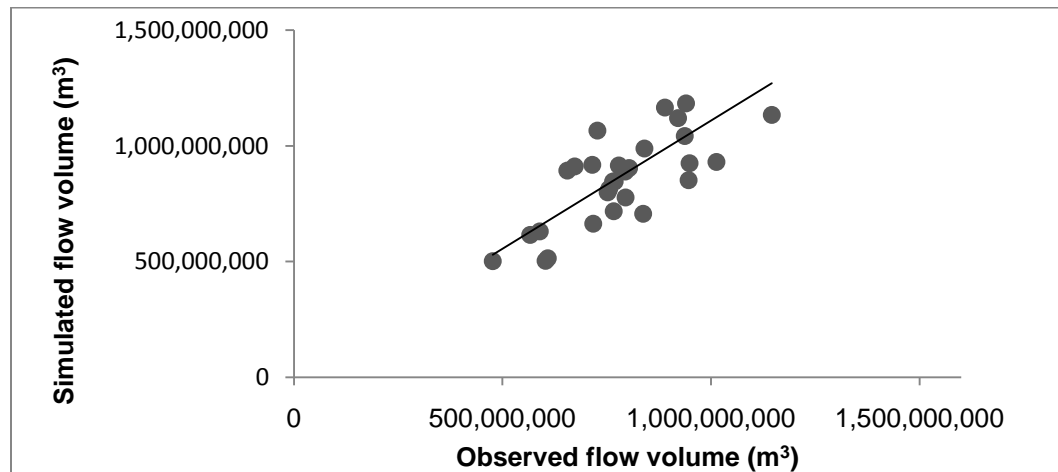
The hydrological balance for three elevation bands over the entire period between November 8, 1960 and October 30, 1990 is presented in table 3. The aforementioned dates were chosen because they represent the start of the 1961 water year and end of the 1989 water year for the 1500 m elevation band. All variables solving for runoff produced by both snow and rain are presented in

table 3. Where, *elevation* is meters above sea level, *Runoff* is the total runoff from rain and snow melt, *Precip* is both rain and snow, *Snow Int* is intercepted snow lost to sublimation and *Rain Int* is intercepted rain lost to evapotranspiration. The pack value is the snowpack occurring on October 30, 1990 averaged over the 30 year period.

**Table 3 1961-90 water balance for 3 elevation bands, values represent a 30 year average in mm of SWE.**

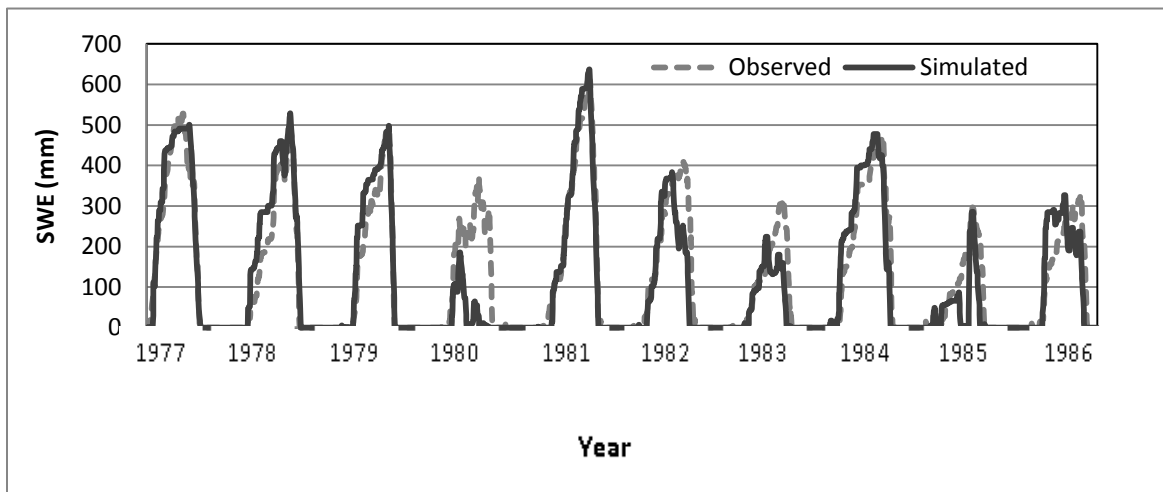
<i>Elevation</i>	<i>Runoff</i>	<i>Precip</i>	<i>ET</i>	<i>Snow Int</i>	<i>Rain Int</i>	<i>Pack</i>	<i>Balance</i>
<b>1500</b>	699	1280	268	189	125	0	0
<b>2000</b>	1101	1658	225	218	109	4	0
<b>2500</b>	1516	1969	126	229	86	13	0

Two separate analyses were used to evaluate the GENESYS model. The first analysis was a comparison of observed annual flow volume at the international border gauging station and simulated annual runoff over the whole basin (Figure 7). The second analysis was a comparison of snowpack values (mm SWE) between the Many Glacier SNOTEL site and its corresponding TC.



**Figure 7 Comparison between simulated annual runoff and observed annual flow volume.**

For the years 1960 to 1990 simulated annual flow volumes are compared to observed values using forced origin linear regression (Figure 7). Results show strong agreement between simulated and observed values ( $r^2=0.956$ ,  $p<0.0001$ ). There is an 11% overestimation of the annual volume for the years 1960 to 1990 over the watershed; this is likely a function of the precipitation estimates for the rainy seasons or underestimation in evapotranspiration (or sublimation). Observed mean annual flow volumes do not significantly differ from simulated mean annual runoff volumes at the 95% confidence level for the 1960 to 1990 period (t-test,  $p=0.122$ ). Over the 30 years on average 21% of the annual snow fall over the entire watershed is sublimated, similar to values reported by Hood *et al.* (1999) and Pomeroy *et al.* (1999). Interception values compare well with Pomeroy *et al.* (1998) accounting for an average of 34% of the total annual precipitation over the watershed.

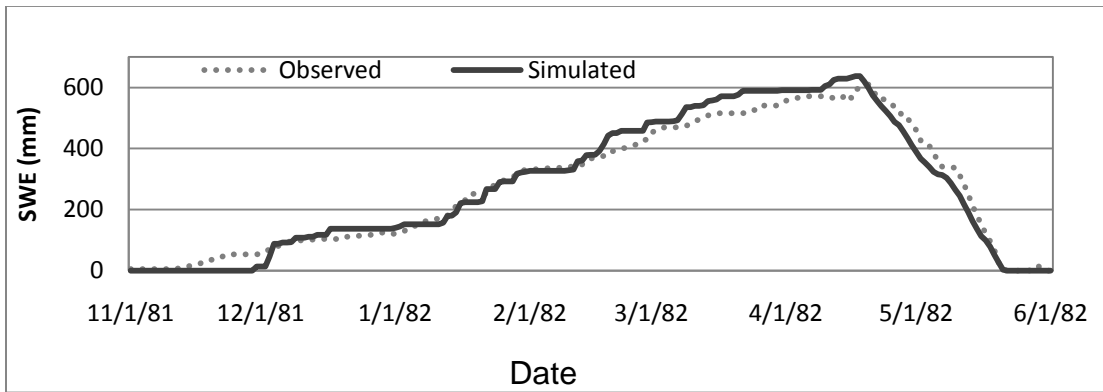
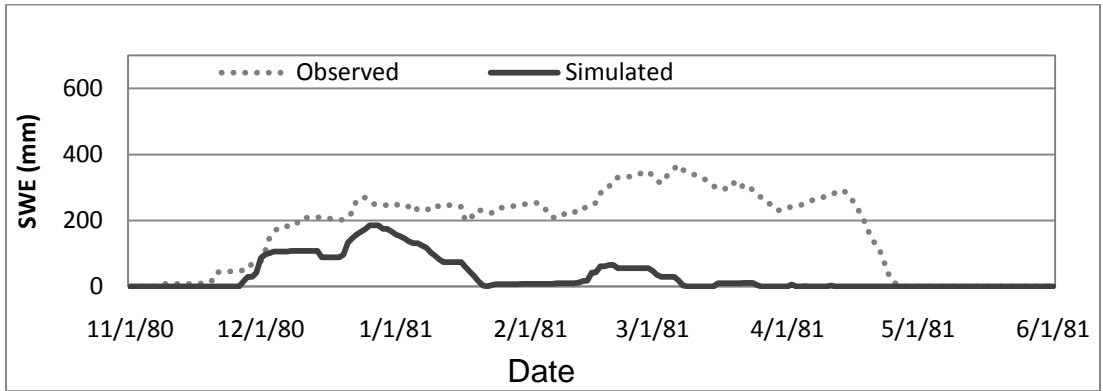
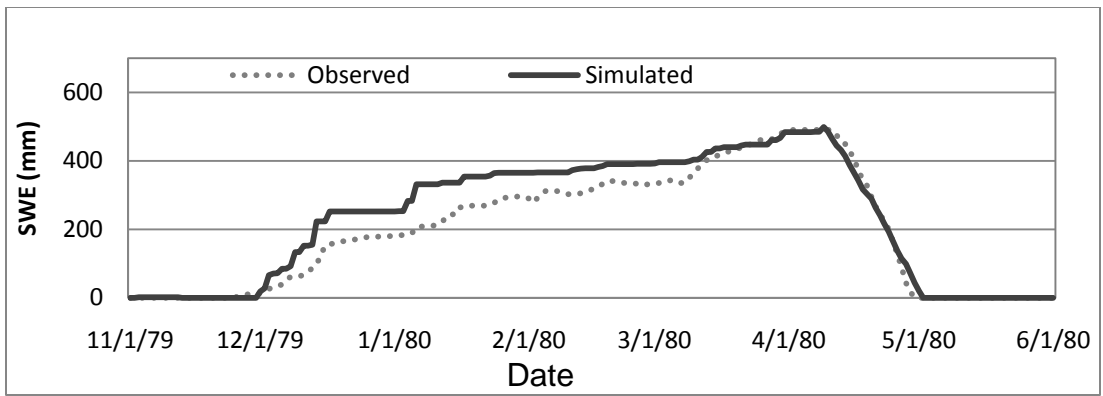


**Figure 8 Comparison between daily simulated snowpack (mm SWE) and observed snowpack at Many Glacier for the 10 year trial period.**

Simulated snowpack compares well with observed snowpack at the Many Glacier snow pillow for the 10 year validation period ( $r^2=0.824$ ,  $p<0.0001$ ).

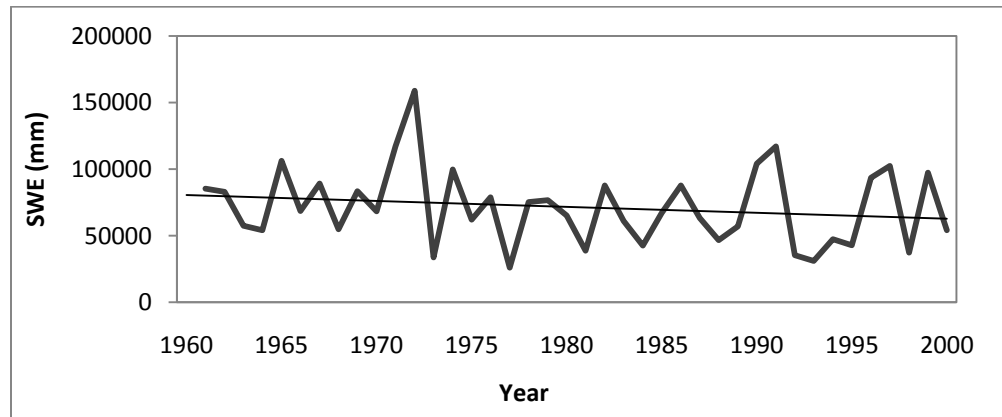
Figure 8 shows that snow melt rates are well simulated with the exception of 1980. Figure 9 provides a closer look at the behavior of the GENESYS modeled seasonal snowpack. The snow years 1979-80 and 1981-82 show a generally strong agreement between simulated and observed snowpack at the Many Glacier SNOTEL site ( $r^2=0.91$ ,  $p<0.0001$  and  $r^2=0.97$ ,  $p<0.0001$  respectively).

For snow year 1980-81, the simulated snowpack significantly deviated from the values at Many Glacier ( $r^2=0.026$ ,  $p=0.037$ ).



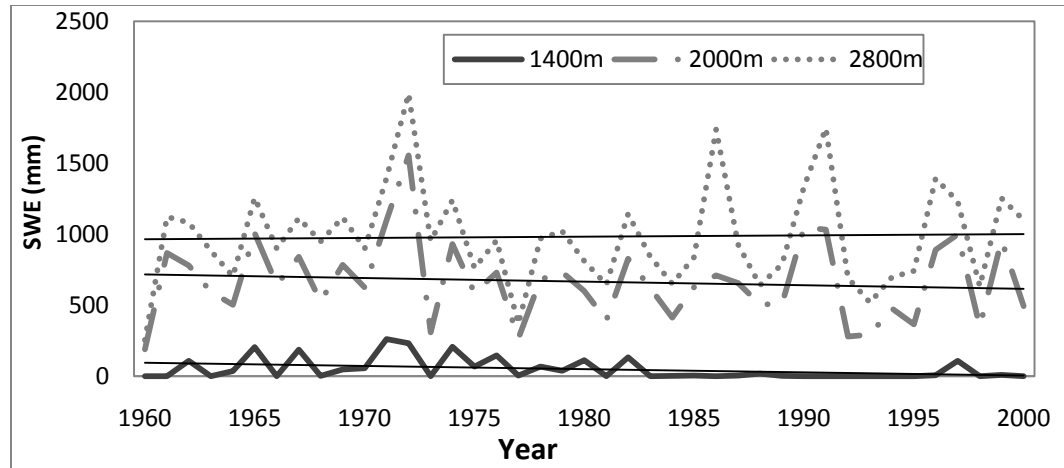
**Figure 9 Comparisons of simulated daily snowpack (mm SWE) and observed daily snowpack at Many Glacier for the years 1979, 1980, and 1981.**

During the 1980-81 snow year complete melt occurred in mid winter, likely as a function of temperature overestimation. Also, precipitation input for this year was significantly lower than the measured values at the Many Glacier SNOTEL site. This suggests that St. Mary data did not represent hydrometeorological conditions at Many Glacier for this year.



**Figure 10 Simulated April 1 SWE over the entire St. Mary watershed from 1961 to 2000.**

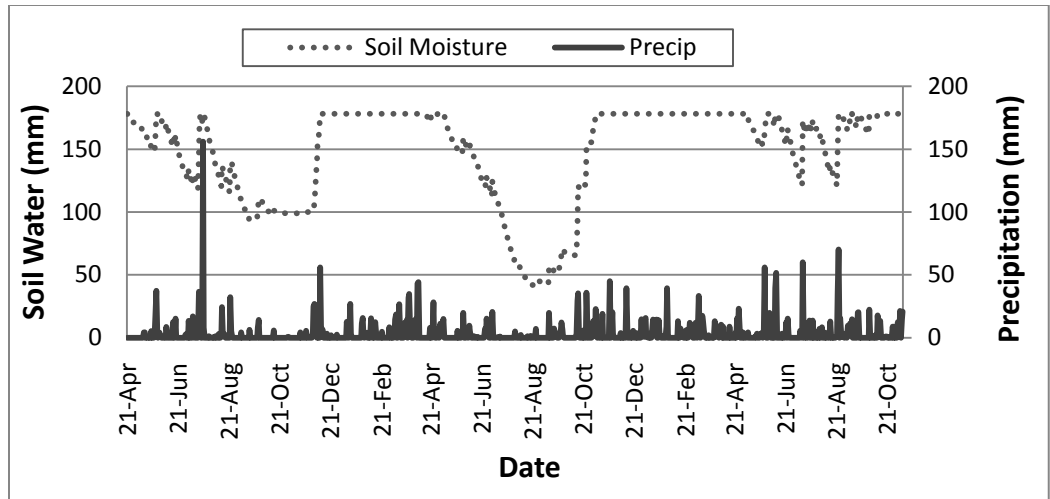
The Mann-Kendall test showed no significance in the slight decreasing trend in basin total SWE on April 1 for each year (Figure 10). To show snow line changes and the effects of elevation on April 1 SWE, 1400 m, 2000 m, and 2800 m April 1 SWE values are plotted (Figure 11).



**Figure 11 Comparison of simulated April 1 SWE between 3 elevation bands.**

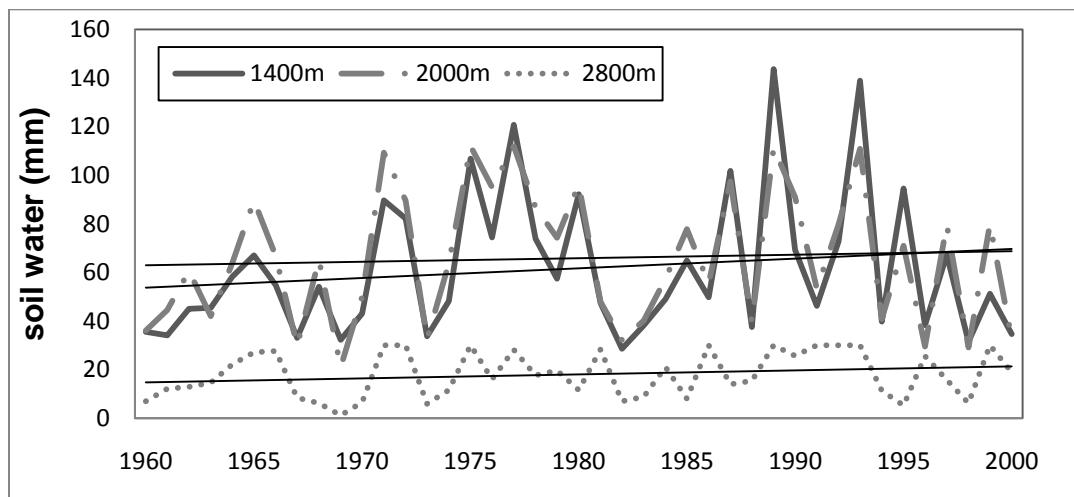
The Mann-Kendall test was used to test the significance of trends at each elevation band and resulted in a significant decreasing trend at 1400 m at the 95% confidence level, a non-significant decreasing trend at 2000 m, and a non-significant increasing trend at 2800 m. This shows that April 1 SWE as well as changes in snow line are dependent on elevation and that processes affecting snowpack differ over the drainage.

Soil moisture and precipitation values are plotted for a two year period starting April 21, 1987 and ending November 7, 1989 (Figure 12). This plot demonstrates the models ability to simulate daily soil water fluctuations at the 1500 m elevation band for a TC with a maximum soil water holding capacity of 178 mm. Precipitation is plotted to show the effect of precipitation conditions on recharging soil water storage.



**Figure 12. Soil moisture and Precipitation plotted for 2 water years (Apr 21, 1987 to Nov 7, 1989)**

Changes in late season soil moisture conditions are assessed by analyzing trends in simulated soil water on August 31 for the 1960 to 2000 period (Figure 13).



**Figure 13 Comparison of August 31 soil water at 3 elevation bands.**

Using the Mann-Kendall test the predicted August 31 soil water values for the 1400m, 2000m, and 2800m elevation bands all show non-significant

increasing trends. These results suggest that late season soil water has no dependence on elevation; however, this could be a function of the evapotranspiration routine, which is unable to account for the heterogeneous land cover.

## **2.5 Discussion**

This study has demonstrated the applicability of the GENESYS model in estimating spatially dependent variables over a mountainous watershed. This physically based method provides an alternative to spatial interpolation and can operate at a fine spatial scale. The model produces two important outputs, spatial estimates of hydro-climatic variables and predictions of detailed hydrological processes at a daily time step.

None of the spatial datasets could be objectively evaluated directly with measured data. Comparisons between annual stream flow volume and annual predicted runoff volume were made. Annual runoff volume estimates from the GENESYS model provide an integration of precipitation, temperature, solar radiation, and relative humidity, as all spatial variables are used as input for deriving annual runoff volume.

When compared to observed annual stream flow volume, the GENESYS model performs well, with a high coefficient of determination. An overestimation of 11% of the annual stream flow volume over 30 years suggests that the model is performing well in simulating the annual hydrological balance. The 11% overestimate compares well with results from Milewska *et al.* (2005) who

evaluated precipitation datasets from PRISM (Daly, 1994, 2002) and ANUSPLIN (Hutchinson, 2004) using annual stream flow. They showed that in a southern Alberta watershed with similar topography, using PRISM and ANUSPLIN precipitation datasets as inputs to an annual mass balance resulted in a 1% overestimation and 6% underestimation of annual precipitation respectively.

If the assumption that the overestimation in predicted annual stream flow volume is a function of precipitation estimates is correct, then it is likely that summer precipitation estimates are too high. Summer precipitation is less dependent on elevation and more dependent on convective storms (Hanson, 1982). Therefore, using winter relationships to estimate spring, summer, and fall volumes may result in severe overestimates in high volume rain events during these seasons.

The dependence of precipitation on topography is also only partially accounted for using elevation as the sole predictor. The precipitation-elevation relationship used in this study explains a maximum of 49% of the variability in mean winter precipitation. The model would likely explain more of the variability in precipitation, if other topographical predictors could be used. For instance, Anderton *et al.* (2004), Basist *et al.* (1994), Daly (1994, 2002) and Marquez *et al.* (2003) have shown that using a combination of variables including slope, aspect, coastal proximity, and orientation can greatly increase the predictive power of regression based estimates of precipitation. Relationships between precipitation and slope and aspect were not significant for the Preston snow course (Larson, 2008), likely because the data do not represent enough

variability in slope and aspect. We have shown, however, that the regression equations used herein provided reliable estimates of snowpack over mountainous terrain; this is evident looking at relationships presented in Figures 8 and 9.

With the exception of 1980, daily snow pack accumulation and ablation processes simulated by GENESYS are representative of actual conditions at the Many Glacier SNOTEL site for the 10 year trial period. Estimates of water loss through interception and sublimation compare well with other studies in mountainous regions (eg., Hood *et al.*, 1999). Evapotranspiration estimates carried out using the Penman-Monteith equation are limited in that the equation was derived for grassland regions; however, the equation provides reasonable estimates given the available data (Table 3).

By showing differences in simulated snow pack for the 10 years, this study has demonstrated the difficulties in using low elevation climate records for driving spatially distributed hydrological models in complex terrain. Using a single low elevation station to determine a watershed scale hydrological balance is subject to significant influence from local hydrometeorological conditions (Daly, 2007). We've shown that the physiographical and climatological characteristics of the St. Mary climate station differ significantly from the mountainous portion of the watershed. We have accounted for this difference with a seasonal shift in precipitation. However, using only St. Mary climate data as input still leads to a bias in estimates that are influenced by the individual characteristics of the climate station. GENESYS has the capability to mitigate this by including

additional observations which would likely result in better spatial representation of meteorological conditions.

A process to include that would enable more accurate estimations of sublimation, interception, and melt would be wind redistribution of snow. Snow redistribution could not be included in the GENESYS model due to the lack of observed records representing spatial and temporal variability in wind speed. This is likely not a problem in the forested portions of the watershed but results in an underestimation of redistribution and sublimation in the exposed areas.

Simulated April 1 SWE trends provide insight into possible changes in mountain late season snowpack and spring snow melt (Cayan, 1996). Significant trends in April 1 SWE were not found over the entire basin or at high elevations. However, these trends are likely not noticeable, due to the dependence of April 1 SWE on spring temperature (Mote *et al.*, 2005). With temperatures remaining cold over the upper watershed, the influence of earlier melt may not be observable in the historical record.

Snow cover at lower elevations is, however, more susceptible than higher elevations to changes in temperature because of their differences in temperature regime (Nolin and Daly, 2006). Simulations show that there is a significant decreasing trend in April 1 SWE at lower elevations; demonstrating changes in snow line and the effects of increased temperature on the timing of spring snow melt. These results agree with Mote (2003) and Laternser and Schneebeli (2003) who found that high elevations experience different trends in late season

snow pack from low elevations. The detection of trends presented in the literature supports the models ability to simulate potential changes in mountain hydrology and subsequent impacts on water resources.

Results from a two year simulation demonstrate the ability of the GENESYS model to estimate changes in soil moisture, however, significant annual trends in simulated August 31 soil moisture were not found. The ability to estimate soil water conditions spatially allows the model to predict drought conditions, a severe problem in agricultural regions of North America (Sauchyn *et al.*, 2006). It is hoped that future developments of the GENESYS model will enable us to investigate the impacts of wide scale drought and fire on a range of ecosystem services including hydrology and ecosystem productivity.

## **2.6 Summary**

This chapter has demonstrated the ability of the GENESYS Model to account for spatial and temporal changes in hydro meteorological conditions in mountainous environments. This study supports suggestions by Daly (2007) and Bales *et al* (2006) showing that the key limitation to modelling in remote regions is the lack of observed data for model calibration and validation. With increased monitoring and physical studies describing important processes like wind redistribution of snow (e.g. Anderton *et al.*, 2004) and cold air drainage (e.g. Lundquist and Cayan, 2007) modelling efforts will be more suitable for making management decisions.

Future work with GENESYS will involve incorporating wind data, enabling more representative estimates of sublimation, evapotranspiration, and interception. The GENESYS model has much potential in spatially estimating mountain ecosystem change in response to anthropogenic and natural disturbance as appropriate data for model development, calibration, and verification become available.

## CHAPTER 3

### ASSESSING THE POTENTIAL IMPACTS FROM CLIMATE CHANGE ON MOUNTAIN SNOWPACK IN THE ST. MARY RIVER WATERSHED, MONTANA

#### 3.1 Introduction

Mountains play a key role in the global hydrological cycle and are a main source of water for many of the world's river systems (Beniston *et al.*, 1997). It is expected that climatic change will enhance the hydrological cycle (Arnell *et al.*, 1999; Thomson *et al.*, 2005; Huntington, 2006), resulting in higher rates of evaporation (Thomson *et al.*, 2005), increased proportion of rain to snow (Huntington *et al.*, 2004; Knowles *et al.*, 2006), and potential changes in the amount and seasonality of precipitation (Arnell, 1999; Field *et al.*, 2007). Changes in the amount of precipitation, in particular maximum snow accumulation, affect the volume of runoff, while changes in temperatures affect the timing of runoff, leading to earlier melt and reduced late season streamflow (Barnett *et al.*, 2005). These changes, augmented by other anthropogenic disturbance, will have a significant impact on mountain snowpack, and subsequently, snow-derived water supply (Beniston, 2003).

Water supply on the western Prairies of Canada is highly dependent on snow melt from the eastern slopes of the Rocky Mountains (Schindler and Donahue, 2006). Mountain snowpack provides temporary storage of water, released at important times of the year (Hamlet *et al.*, 2005). Mountain snow accumulations are likely to decline with continued atmospheric warming (Hamlet and Lettenmaier, 1999), resulting in a reduction of available water from

snowpack in mountainous regions (Barnett *et al.*, 2005; Lapp *et al.*, 2005).

Numerous studies have already shown hydrological changes in snow dominated regions, with earlier onset of melt (Cayan *et al.*, 2001; Stewart *et al.*, 2004; Stewart *et al.*, 2005), decreases in low-elevation snow cover (Nolin and Daly, 2006), and decreases in mean annual streamflow (Zhang *et al.*, 2001; Rood *et al.*, 2005). This study investigates the effects of climate change on mountain snowpack in the St. Mary drainage basin, Montana, a watershed that supplies water for approximately 200,000 ha of irrigation for Southern Alberta, Canada.

To assess potential changes in mountain snowpack in the St. Mary watershed we apply a fine scale hydrometeorological model driven by General Circulation Model (GCM) derived scenarios of future climate. This combination of GCMs and fine scale hydrometeorological modelling provides a structure for conceptualizing and investigating the relationships between climate and water resources (Leavesley, 1994; Xu, 1999).

### **3.1.1 GCM description**

GCMs currently provide the most sophisticated, physically based approach to simulate large-scale responses of the climatic system to projected changes in greenhouse gas (GHG) emissions (Laprise *et al.*, 2003). GCMs depict climate for a three dimensional grid over the entire globe (IPCC, 2007) using mathematical representations of the physical climatic processes that link the radiation budget of the atmosphere and land surface to global circulation and the hydrological cycle (McFarlane *et al.*, 1992). The main components of a GCM

account for radiative transfer, cloud formation, precipitation, boundary layer physics, and GHG concentrations (Giorgi and Mearns, 1991). GCMs are run with initial boundary conditions that account for sea surface temperature, sea ice, snow cover, and GHG abundance. The more recent GCMs are able to account for both oceanic and atmospheric circulation, enabling more accurate assessments of the impacts of increased greenhouse gas emissions (McFarlane *et al.*, 1992).

GCMs operate at an hourly time scale over several centuries, requiring very high computational power. Because of this high computational load, GCMs operate at a spatial scale on the order of hundreds of square kilometers (Laprise *et al.*, 2003). GCMs accurately simulate annual and seasonal climate over large regions, however, due to their coarse spatial resolution, lack the ability to model local climatic conditions, particularly in regions of diverse terrain (Xu, 1999; Hay *et al.*, 2000). Appropriate representation of local climatic variability is needed to make suitable assessments of impacts of climate change on fresh water ecosystems (Hauer *et al.*, 1997).

### **3.1.2 Modelling approach**

To account for local climatic variability we apply the GENESYS (GENerate Earth SYstems Science input) model. The GENESYS model, developed at the University of Lethbridge, is a physically based, spatial hydrometeorological model that operates at a daily time step and can be applied in complex terrain. GENESYS has been successfully used to simulate the hydrometeorological

conditions, and impacts of climate change for the St. Mary (Larson *et al.*, in review; MacDonald *et al.*, in review) and Oldman River watersheds (Sheppard, 1996; Lapp *et al.*, 2002; Lapp *et al.*, 2005). The process based nature of GENESYS is useful for the assessment of impacts of climate change on mountain hydrometeorology.

MacDonald *et al* (in review) applied GENESYS to simulate daily spatial hydrometeorological variables and annual runoff volume for the 1960-90 time period. In that study, the GENESYS Model successfully simulated daily snow accumulation and ablation for a ten year period at the Many Glacier SNOTEL site within the St. Mary watershed. In this chapter we use the GENESYS model to assess the impacts of climate change on spatial and temporal characteristics of mountain snowpack in the St. Mary watershed. However, in order to perturb model input for assessing the impacts of climate change on mountain hydrometeorology, GCMs still provide the most reliable information.

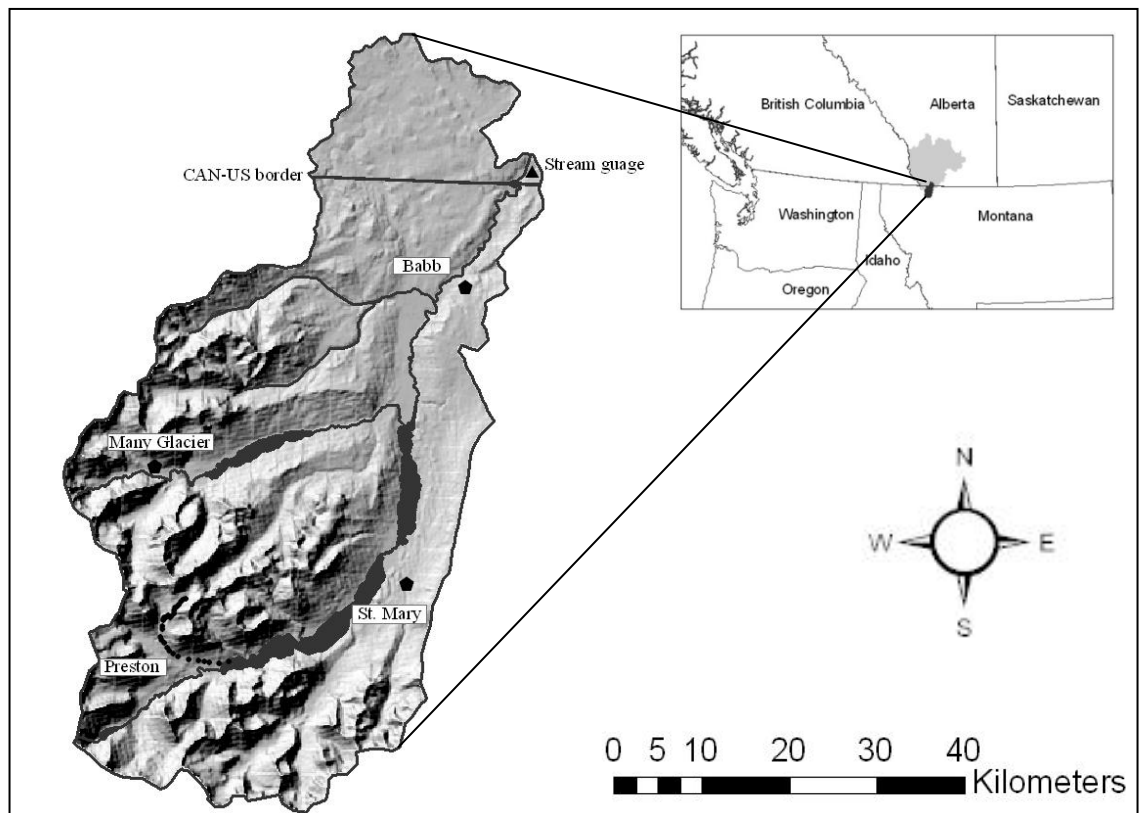
Regional downscaling methods exist that enable the use of GCM outputs at the sub-grid scale (Mearns *et al.*, 2001), thereby making GCM output suitable for environmental modelling in complex terrain. We apply the “delta” method of downscaling to couple the GENESYS model to GCM output. This method applies monthly changes derived from GCM output to the observed climate record (Hay *et al.*, 2000; Wood *et al.*, 1997; Xu, 1999). The “delta” method has been used to downscale GCM output in numerous hydrological impacts studies (Hamlet and Lettenmaier 1999; Morrison *et al.*, 2002; Andreasson *et al.*, 2004; Loukas *et al.*, 2004; Cohen *et al.*, 2006; Merritt *et al.*, 2006). The limitation to this method is that

changes in the variability of climatic conditions are not accounted for (Leavesley, 1994). However, the local variability of the driving climate station is preserved (Hamlet and Lettenmaier, 1999). Given the uncertainty in future variability and range of plausible future climates, the delta method provides a conservative estimate of the impacts of climate change on water resources (Merritt *et al.*, 2006).

We use a number of GCM scenarios to perturb the GENESYS model and test the sensitivity of winter snow hydrology to a range of possible future climates, similar to the method used by Barrow and Yu (2005). The objective of this work is to assess potential changes in snowpack timing, volume, and spatial coverage in the St. Mary drainage basin for the 2020s (2010-39), 2050s (2040-69), and 2080s (2070-99). To determine available water from snow we use estimated annual maximum basin scale snowpack (mm of SWE), as this is a good predictor of spring runoff (Barnett *et al.*, 2005). We also assess changes in rain to snow ratios, an important measure of the impact from climate warming on snow accumulation (Knowles *et al.*, 2006). To evaluate the impact from climate change on spatial coverage of snowpack we analyze the change in long-term mean annual snow cover. The temporal change in future snowpack is assessed using the Julian date of maximum snowpack over the watershed which is a surrogate for the onset of spring melt.

### 3.2 Study Area

The headwaters of the St. Mary watershed lie along the east slopes of the Rocky Mountains with the majority of the upper watershed residing within Glacier National Park, Montana. The St Mary River flows from the continental divide, through the upper and lower St. Mary lakes, and ends in southern Alberta where it meets the Oldman River (Figure 14).



**Figure 14 The St. Mary Watershed in Montana and Southern Alberta. The locations of the watershed stream gauging point, Preston snow survey, Many Glacier SNOTEL site and St. Mary and Babb weather stations are shown.**

The climatic regime is a transitional zone between coastal and continental climates. The area receives the majority of its precipitation in the winter with

snowfall accounting for roughly 70% of the total precipitation at higher elevations (Selkowitz *et al.*, 2002). Chinooks are prevalent on the eastern slopes of the study area and can induce rapid snowmelt in low lying areas.

The total drainage area of the study watershed is 1195 km<sup>2</sup>, with a mean elevation of 1745 m, and ranging from 1249 m to 3031 m. The basin is a relatively undisturbed and ecologically diverse region. This is largely attributed to the fact that a large portion of the drainage is within Glacier National Park. Coniferous forests account for 24% of the land cover, deciduous forest covers 21%, low herbaceous plants cover 29% of the area, 23% is barren rock or soil, and 3% is water (USGS, 2000).

### **3.3 Methods**

#### **3.3.1 Hydrometeorological model**

The GENESYS model integrates GIS and a series of physical subroutines to estimate spatial hydrometeorological variables on a daily time step over a study watershed. To apply the physical subroutines over the watershed, terrain categories (TCs) are used which enable representative physiographic information about the watershed to be integrated efficiently into process based equations.

TCs are determined by combining a digital elevation model (DEM) classified into 100 m elevation bands and a classified land cover grid. The land cover grid used consists of 9 land cover classes: dry herbaceous, mesic herbaceous, deciduous trees/shrubs, coniferous trees/shrubs, conifer trees/open, water, snow, barren rock/soil, and shadows. Snow and shadow classes were

eliminated from the land cover grid and assigned the values of the nearest land cover. For the St. Mary watershed, the combination of land cover and elevation, where TCs representing less than 1% of the watershed were joined to the neighboring TC, resulted in 82 individual TCs.

To estimate temperature, precipitation, solar radiation and humidity, GENESYS applies the MTCLIM model (Hungerford *et al.*, 1989) for each TC. This is achieved by applying physical subroutines that use meteorological information from a BASE climate station and characteristics of each TC. Daily meteorological estimates are used as inputs to subroutines that determine winter and open season hydrological processes, including snow accumulation and ablation, canopy interception, sublimation, evapotranspiration, soil water, and runoff.

Sublimation is estimated using a formula derived by Déry *et al* (1998). Potential evapotranspiration (ETP) is determined using humidity, solar radiation, and average temperature with a modified Penman-Monteith equation derived by Valiantzas (2006). The partitioning of precipitation into snow and rain is achieved using the temperature based method developed by Kienzle (2008), where we use mean annual values for threshold temperatures. Snow melt is calculated with a temperature based melt routine developed by Quick and Pipes (1977). Canopy interception routines differ between snow and rain. For snow interception, we use the Hedstrom and Pomeroy (1998) equation, for rain interception we use an equation derived by Von Hoyningen-huene (1983). Both equations require leaf area index (LAI) input. Runoff and soil water storage are

determined in a mass balance where soil water is depleted by evapotranspiration (ET) and runoff is determined by rain and snow melt. For a detailed description of the most recent version of GENESYS refer to MacDonald *et al* (in review).

### **3.3.2 GCM data**

The monthly GCM data used were made available by the Pacific Climate Impacts Consortium (<http://pacificclimate.org/>). Model experiments available (Table 4) represent models recommended by the *IPCC Data Distribution Center Task Group on Data and Scenario Support for Impact and Climate Analysis* (IPCC-TGICA, 1999). Following the recommendations of the IPCC, a number of scenarios are used to capture a range of possible future climates. To select the range of scenarios, a method adapted from Barrow and Yu (2005) is used, where scenarios are selected using annual mean temperature change and percent precipitation change for the 2020s. The 2020 time period is selected because it is likely the most accurate estimate of potential change given the high degree of uncertainty in future GCM simulations (Merritt *et al.*, 2006).

**Table 4 GCMs and experiments available from the PCIC (after Barow and Yu, 2005).**

Modeling Center	Country	Model	SRES simulations
Canadian Center for Climate Modeling and Analysis	CAN	CGCM2	A2, B2
Hadley Centre for Climate Modeling and Research	UK	HadCM3	A1F1, A2, B1, B2
Max Planck Institute for Meteorology	GER	ECHAM4	A2, B2
Commonwealth Scientific and Industrial Research Organization	AUS	CSIRO-Mk2	A1, A2, B1, B2
Geophysical Fluid Dynamics Laboratory	USA	GFDL-R30	A2, B2
National Centre for Atmospheric Research	USA	NCAR-PCM	A2, B2, A1B
Centre for Climate Research Studies	JPN	CCSR/NIES	A1F1, A1T, A1B, A2, B1, B2

The variability in temperature change projected by all GCMs was low relative to the variability in precipitation. Therefore, the selection of each scenario was more reliant on the predicted precipitation change. The objective was to select output from models that represented a range of GCM simulations for the future periods. Following Von Storch *et al.* (1993), regional estimates of temperature and precipitation are derived as the average of the 4 grid cells closest to the study site. We chose 5 models representing a range of plausible future climates (Table 5).

**Table 5 Model experiments used in this study (after Barrow and Yu, 2005).**

Scenario	GCM	Emissions Scenario	Resolution (°)
ECHAM	ECHAM4	A2 (1)	2.8 x 2.8
CSIRO	CSIRO-Mk2	A1 (1)	5.6 x 3.2
CGCM	CGCM2	A2 (3)	3.75 x 3.75
CCSR	CCSR/NIES	A1T	5.62 x 5.62
NCAR	NCAR-PCM	B2 (1)	2.8 x 2.8

Each of the 5 models is driven by a future greenhouse gas emission scenario that is determined based on future population, economic, societal, and environmental change (Nakicenovic *et al.*, 2000). Descriptions of the emissions scenarios used in this study are presented in Table 6.

**Table 6 Description of SRES scenarios used in this study (after Barrow and Yu, 2005).**

Emissions Scenario	Description
A1	A future world with rapid economic growth, and a mix of technological developments and fossil fuel use
A1T	A future world with rapid economic growth, and introduction of new and more efficient technology
A2	A future world with moderate economic growth, more heterogeneously distributed and with a higher population growth rate than in A1
B2	A world where the emphasis is placed on local solutions to economic, social and environmental sustainability and with intermediate levels of economic development and a lower population growth rate than A2

### 3.3.3 Downscaling

We apply the “delta” method, where changes in GCM derived temperature and precipitation for the 2020s, 2050s, and 2080s relative to the 1961-90 time series are made. To determine daily minimum and maximum absolute temperature changes a Fourier transform is applied (Epstein and Ramirez, 1994; Morrison *et al.*, 2002). This is done using the 12 monthly values predicted for maximum and minimum temperature for each time period from each GCM scenario. The Fourier transform is applied to these 12 values creating 365 continuous predicted maximum and minimum temperature changes. These values are then added to observed temperature values for every day. Percent change in precipitation

volumes are determined relative to the 1960-90 period, where the mean monthly percent change is multiplied by the observed values on days when precipitation occurred. New 31-year datasets representing changes in temperature and precipitation predicted by the GCMs are used to perturb the hydrometeorological model for 15 simulations of the future climate.

### **3.3.4 Data analysis**

To evaluate future maximum snowpack (mm of SWE) and the date of spring snowmelt, we use the Mann Kendall non-parametric test. This test is frequently used for trend detection of hydrological variables (Burn and Hag Elnur, 2002). The non-parametric Sen's slope test is used to determine the slope of the trend line (change in the variable per year). Complete time series from 1961 to 2099 could not be created due to data limitations, therefore, the time series analyzed does not contain the years 2006 to 2009. To assess spatial change in mean annual snowpack (mm of SWE), 30 year mean surfaces are created for each time slice using a GIS. Surfaces are compared visually and percent reduction in mean annual snowpack is calculated relative to the 1961-90 base period.

### 3.4 Results and Discussion

#### 3.4.1 Climate Change Scenarios

There is a general agreement among the climate change scenarios that there will be increases in mean annual temperature. However, predictions of changes in mean annual precipitation are more variable (Table 7). The magnitude of predicted temperature change for the 2020s ranged from 1.0 °C to 1.7 °C relative to the 1961-90 base period. The predicted relative precipitation change for the 2020s ranged from a 4.0% decrease to a 5.0% increase (Table 6). The variability in temperature and precipitation predictions increased for later time periods, demonstrating a greater degree change and uncertainty as a function of time.

**Table 7 Annual changes in temperature and precipitation for each of the 5 scenarios relative to the 1961-90 base period.**

Time period	ECHAM		CSIRO		CGCM		CCSR		NCAR	
	Temp (°c)	Precip (%)	Temp (°c)	Precip (%)	Temp (°c)	Precip (%)	Temp (°c)	Precip (%)	Temp (°c)	Precip (%)
<b>2020s</b>	1.4	-4	1.7	3	1.4	1	1.0	-3	1.2	5
<b>2050s</b>	2.6	-1	3.3	4	2.7	-1	4.3	3	1.5	7
<b>2080s</b>	3.8	-2	4.8	10	4.6	3	6.3	8	2.0	14

#### 3.4.2 Rain to snow ratio

Predicted changes in the timing and development of snowpack resulting from climate change are affected by the projected increase in temperature and subsequent effects on determining the phase (rain or snow) in which precipitation occurs in the GENESYS model. We applied the algorithm developed by Kienzle

(2008) to partition precipitation into rain and snow. This method uses the mean daily temperature estimates for each TC to determine whether precipitation falls as rain or snow.

For the simulated historical period, 70% of the total annual precipitation was snow and 30% was rain. The GENESYS model predicts that as temperatures increase, the proportion of snow to total annual precipitation decreases from the historical period in every scenario (Table 8). Merritt *et al* (2006) found similar results in the Okanagan Basin, BC. They predict that a greater proportion of rain during transitional months will lead to declines in snowpack under projected climate change. There is very little variability between all scenarios for the 2020 time period, however, variability increases between all scenarios by the 2080s (Table 8).

**Table 8 Changes in rain and snow as a percentage of the total precipitation for all 5 scenarios.**

	ECHAM		CSIRO		CGCM		CCSR		NCAR	
	%snow	%rain	%snow	%rain	%snow	%rain	%snow	%rain	%snow	%rain
<b>Historical</b>	70	30	70	30	70	30	70	30	70	30
<b>2020</b>	67	33	67	33	67	33	66	44	65	35
<b>2050</b>	63	37	60	40	63	37	54	46	64	36
<b>2080</b>	61	39	54	46	40	60	45	55	63	37

The CSIRO, CGCM, and CCSR scenarios present a case where a significant reduction in snow as a portion of the annual precipitation will likely result in less storage of water and increased winter runoff. Knowles *et al* (2006)

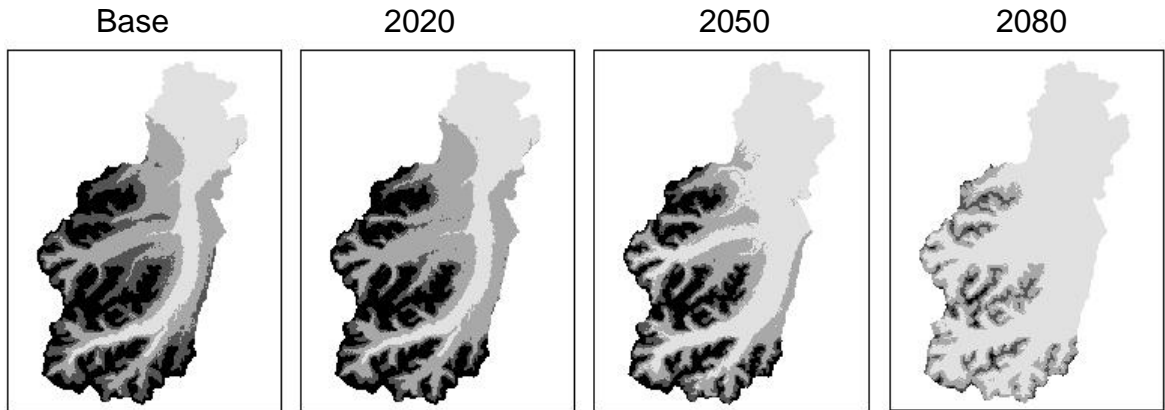
and Leung *et al* (2004) also project that if warming continues to increase, as GCMs predict, resulting decreases in snowfall will affect freshwater supply in the western United States. Even if greater precipitation volumes occur, increased winter runoff will significantly impact management strategies, as flood control and water storage provide contrasting situations (Knowles *et al.*, 2006).

### **3.4.3 Spatial change in mean annual snowpack**

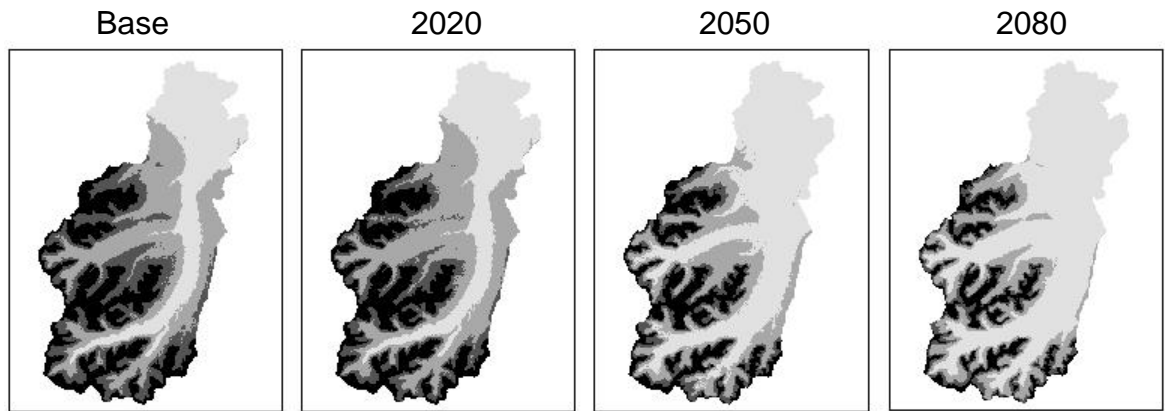
Three scenarios are selected to represent the spatial change in 30 year mean annual snowpack (mm of SWE). The CGCM scenario represents the greatest change, the NCAR scenario represents the least change, and the CSIRO scenario represents a median change in 30 year mean annual SWE (Figure 15).

Figure 15 illustrates spatial changes in snowpack over the watershed for three time slices. The CSIRO and CGCM scenarios both result in a decrease in mean annual snowpack. There is a substantial increase in the area with a mean annual snowpack of 0 to 100 mm and a large decrease in the area covered by greater than 300 mm SWE. Under these scenarios, increases in temperature will reduce the spatial extent and depth of snowpack over the St. Mary watershed. The NCAR scenario results in very little spatial change in mean annual snowpack even by the 2080s. This is expected given that this scenario has lowest increase in temperature and greatest increase in precipitation relative to the other scenarios. Based on our results, the St. Mary watershed under the NCAR scenario is likely to experience little decline in annual snowpack in the future.

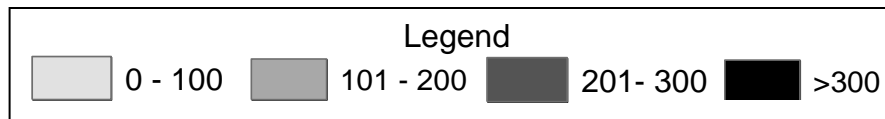
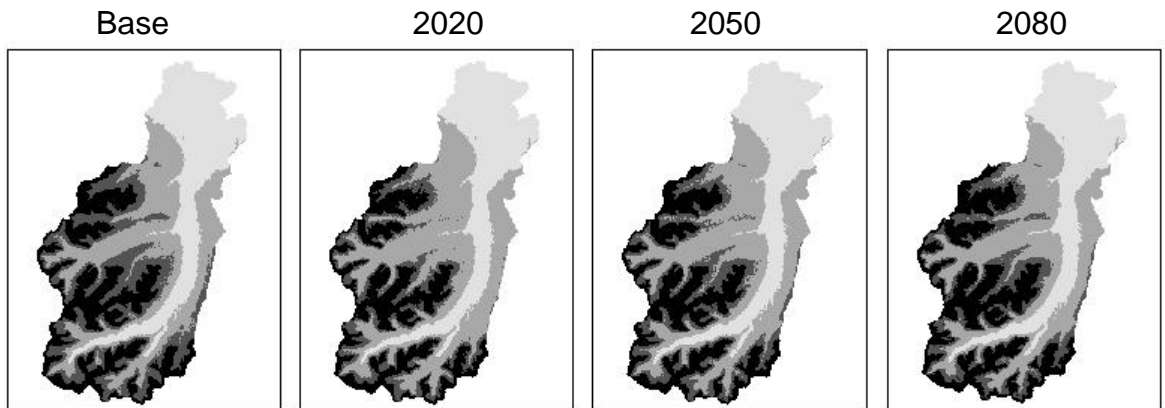
CGCM



CSIRO



NCAR



**Figure 15 Spatial change in mean annual snowpack (mm SWE) representing the greatest (CGCM), median (CSIRO), and least (NCAR) changes relative to the 1961-90 base period.**

Independent of scenario selection and consistent with the findings of Nolin and Daly (2006) and Regonda *et al.* (2005), the greatest change in mean annual snowpack is at low elevations (Table 9), with as much as 79% reduction in mean annual snowpack by the 2080s below the 1600m elevation (CGCM). Higher elevations may not experience significant loss of mean annual snowpack with the greatest reduction of 44% by the 2080s (CGCM) with some simulated scenarios resulting in a modest increase in mean annual snowpack above 2200m (Table 9). Hamlet *et al* (2005) found similar results for the western United States and showed that increases in April 1 SWE at high elevations were a function of increased precipitation above elevations where winter temperatures stay below freezing.

**Table 9 Comparison of mean annual snowpack (mm of SWE) and percent change in mean annual snowpack relative to the 1961-90 historical period.**

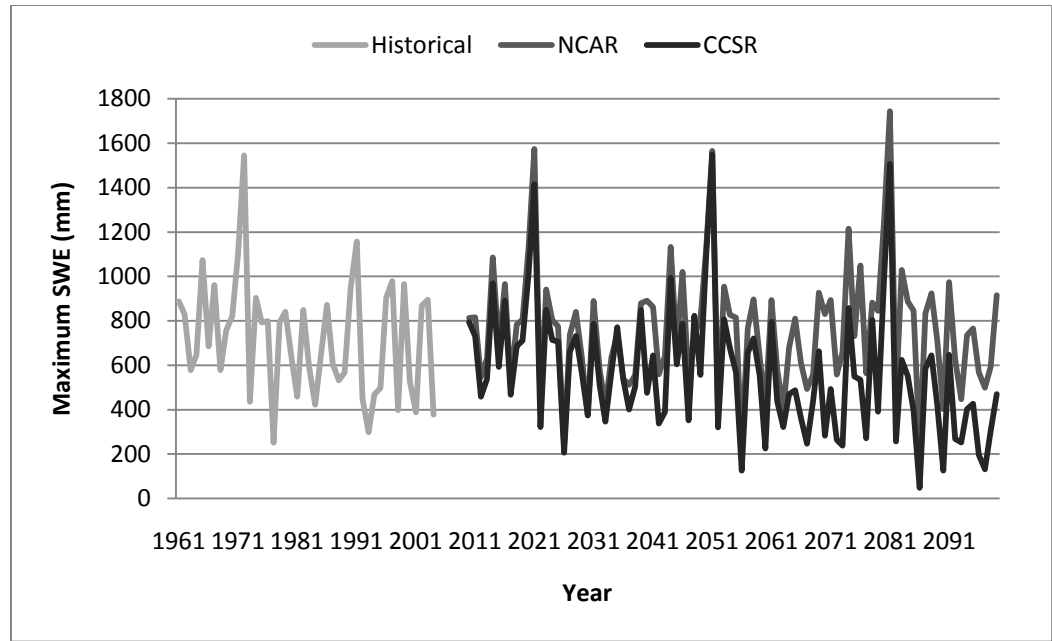
Elevation band	BASE	CGCM			CSIRO			NCAR		
		2020s	2050s	2080s	2020s	2050s	2080s	2020s	2050s	2080s
	61-90									
1400-1600	106	87 (-18%)	65 (-38%)	22 (-79%)	85 (-20%)	57 (-46%)	42 (-60%)	89 (-16%)	91 (-14%)	88 (-17%)
1600-1800	205	185 (-10%)	147 (-28%)	55 (-73%)	187 (-9%)	135 (-34%)	100 (-51%)	184 (-10%)	189 (-8%)	194 (-6%)
1800-2000	267	251 (-6%)	209 (-22%)	92 (-66%)	261 (-2%)	205 (-23%)	162 (-39%)	245 (-8%)	254 (-5%)	270 (1%)
2000-2200	356	347 (-2%)	299 (-16%)	154 (-57%)	361 (1%)	316 (-11%)	280 (-21%)	331 (-7%)	341 (-4%)	373 (5%)
2200-2400	458	449 (-4%)	395 (-14%)	233 (-49%)	467 (2%)	425 (-7%)	411 (-10%)	424 (-7%)	436 (-5%)	485 (6%)
2400-2600	514	506 (-2%)	466 (-9%)	280 (-46%)	523 (2%)	485 (-6%)	485 (-6%)	474 (-8%)	486 (-5%)	546 (6%)
2600-2800	615	586 (-5%)	511 (-17%)	346 (-44%)	600 (-2%)	554 (-10%)	569 (-7%)	549 (-11%)	557 (-9%)	621 (1%)
Basin	213	192 (-2%)	166 (-22%)	77 (-63%)	200 (-1%)	162 (-12%)	138 (-17%)	192 (-2%)	197 (-8%)	208 (-3%)

The GENESYS results show the decrease in mean annual snowpack at lower elevations is in large part a function of temperature change, similar to results found by Mote (2003). As climate warms, low elevation portions of the watershed with historical temperatures near the freezing point will warm past the freezing point, changing both the phase of precipitation and the rate of snow melt (Regonda *et al.*, 2004; Nolin and Daly, 2006).

Although not included in the current version of GENESYS, however, we recognize that as the area of high albedo snow cover decreases, a positive feedback occurs where less heat is reflected from the earth's surface (Fyfe and Flato, 1999; Karl and Trenberth, 2003), thereby increasing warming and further reducing the spatial extent of snow cover over the watershed. It is plausible that as snow cover decreases, the effects of this positive feedback will be pronounced under climate warming. To account for this positive feedback, a temperature algorithm that accounts for albedo will be a useful addition to future versions of GENESYS.

#### **3.4.4 Volume of Maximum Snowpack**

The two scenarios selected to represent the least and greatest changes in maximum annual snowpack (mm of SWE) are the NCAR and CCSR scenarios respectively. Figure 16 presents the maximum snowpack time series for the watershed for these two scenarios.



**Figure 16 Time series of maximum snowpack for the greatest and least change scenarios.**

The Mann Kendall test detects a decreasing trend over time in maximum annual snowpack in the CCSR scenario at the 99% confidence level (Sen's slope estimate = -4.10), while no significant trend is detected in the NCAR scenario (Sen's slope estimate = -0.10). The difference between the scenarios shows the sensitivity of the system to the range of possible future conditions and demonstrates the impact of emissions scenario selection on results.

The lack of trend in the NCAR scenario is likely due to the relatively large increases in precipitation and a low increase in temperature (Table 4). Future conditions under the NCAR scenario could actually see an increase in annual maximum snowpack and, therefore, water supply. This result would, however,

require a significant change in current societal structure given the rate of economic and population growth occurring at the present.

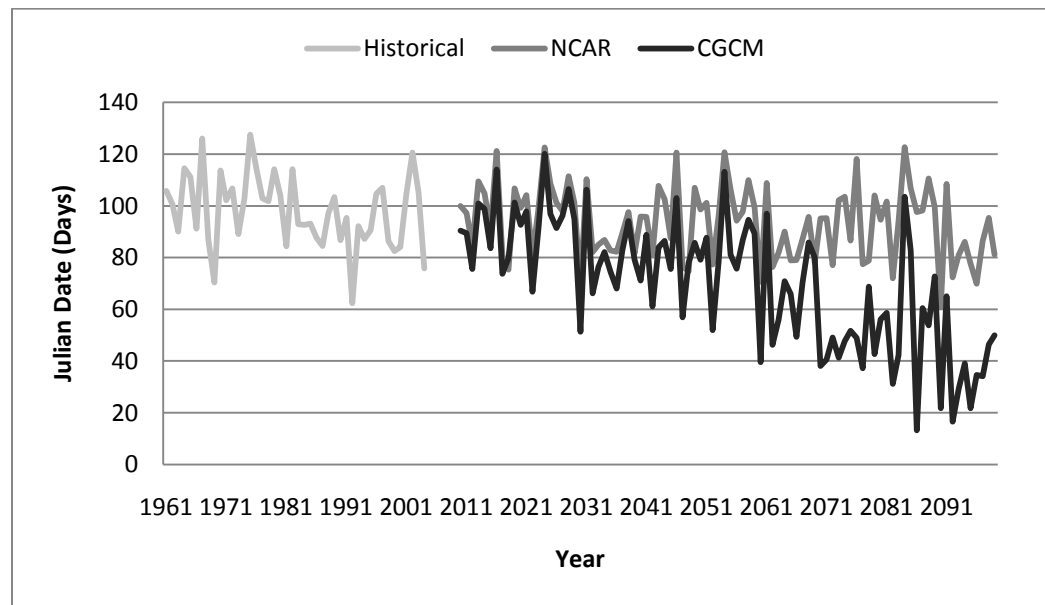
In the CCSR scenario, a modest change in precipitation and increases in temperature result in a significant reduction overall in maximum snowpack, concurrent with observed historical trends in SWE (Hamlet *et al.*, 2005; Mote *et al.*, 2005; Mote, 2006). Positive temperature trends have been shown to be correlated with decreasing snowpack (Mote *et al.*, 2005). Therefore, even with a general increase of precipitation, our modelling under the CCSR scenario suggests that higher temperatures will likely result in a decrease in maximum annual snowpack.

The time of year when the effects of temperature change are perhaps the most noticeable is during the transitional months (Karl *et al.*, 1993). Spring temperatures over western North America have increased by 1 to 3 degrees Celsius since the late 1970s (Cayan *et al.*, 2001). This trend in temperature is expected to continue as atmospheric greenhouse gas concentrations increase (IPCC, 2007). A warmer climate will affect the timing and seasonal distribution of snowmelt (Stewart *et al.*, 2004).

#### **3.4.5 Timing of Maximum Snowpack**

An earlier onset of spring has already been recorded in numerous studies over Western North America (Burn, 1994; Cayan *et al.*, 2001; Regonda *et al.*, 2005; Stewart *et al.*, 2005). An earlier timing of spring runoff will affect water management and ecosystem health, and provides a seasonally and spatially

integrated signal of the impacts of climate change (Stewart *et al.*, 2004). We use the Julian date of maximum snowpack as a surrogate for the onset of snowmelt. The two scenarios selected to represent the least and greatest changes in the Julian date of maximum snowpack over the St. Mary watershed are the NCAR and CGCM scenarios respectively (Figure 17).



**Figure 17 Time series of the Julian date of maximum snowpack for the greatest and least change scenarios.**

The Mann Kendall test identifies a significant decreasing trend in the simulated date of maximum snowpack (mm of SWE) in the CGCM scenario at the 99% confidence level (Sen's slope estimate = -0.66). A significant decreasing trend is also shown in the NCAR scenario at the 90% confidence level (Sen's slope estimate = -0.09). Under the CGCM scenario, earliest dates of maximum snowpack could approach early January by the 2080s, while under the NCAR scenario earliest dates of maximum snowpack could occur in February.

The mean Julian date of maximum snowpack over the watershed is April 8<sup>th</sup> for the historical period. Both scenarios are consistent in predicting an earlier mean Julian date of maximum snowpack over the 2080 period; with the mean Julian date of maximum SWE occurring on April 2<sup>nd</sup> in the NCAR scenario and the March 6<sup>th</sup> in the CGCM scenario by the 2080s. This shift in means along with significant decreasing trends towards earlier Julian dates of maximum snowpack infers an overall earlier onset of spring, and an alteration in the timing of runoff from snowpack in the St. Mary watershed.

An earlier date of maximum snowpack in all scenarios is, again, likely a function of temperature. Studies of trends in the onset of spring support this, showing the important role of increased temperatures on spring snowpack (Hamlet *et al.*, 2005; Mote, 2006). The influence of temperature can be seen by looking at the NCAR scenario, where there is no significant trend in maximum snowpack volume; however, there is a decreasing trend for the Julian date of maximum snowpack. This result is consistent with Clair *et al* (1998) and shows that even an annual increase of 14% in precipitation by the 2080s can be offset by a 2 degree increase in temperature, resulting in earlier melt.

An earlier onset of snowmelt has important implications for water supply and could result in a significant impact on water resources for human and ecosystem use (Barnett *et al.*, 2005; Field *et al.*, 2007). Spring snowmelt is relied upon to provide up to 80% of the annual flow volume in snow dominated basins (Stewart *et al.*, 2004) and has been the most predictable source of water for storage and human use (Stewart *et al.*, 2005). An earlier onset of melt will disrupt

current management practices by modifying the predictability of seasonal deliveries of streamflow (Regonda *et al.*, 2005; Stewart *et al.*, 2005). Reservoir operations and flood management will have to adapt to changes in the onset of peak runoff.

Shifts in peak runoff to earlier in the spring or late winter lead to decreased water availability later in the summer months, when water demand is highest (Frederick and Major, 1997; Barnett *et al.*, 2005). With melt occurring earlier, soil moisture will be depleted sooner, drought potential will be accentuated, and summer flow volumes will decrease (Stewart *et al.*, 2005). Reduced late summer flows will create management problems for human uses, and will likely have adverse impacts on aquatic ecosystems (Schindler, 2001).

Reduced late season flow can result in warmer water temperatures, forcing native cold-water species like the bull trout (*Salvelinus confluentus*) into isolated headwater portions of mountain streams (Schindler, 2001). Changes in late season migration opportunities from reduced passage of low order streams could also alter fish population structures, spawning behaviour, and species distributions (Clair *et al.*, 1998). Riparian ecosystems could be impacted by late season low flows and associated low soil water, leading to greater stress on riparian forests and more frequent die off of young trees (Rood *et al.*, 1995). Overall, aquatic and riparian ecosystems will suffer if late season flows are significantly reduced. It is important to adapt management strategies in order to account for climate change and increased water demands by agriculture, industry, and urban areas, while maintaining ecosystem health.

### **3.4.5 Effect of Scenario Selection on Future Snowpack Predictions**

All climate change scenarios used in this study predict an earlier onset of spring melt. However, there are differences between scenarios with respect to long-term mean annual and maximum snowpacks. These differences are illustrated in Figures 15 and 16 and Table 7. These differences are largely attributed to the GCM used as each GCM produces different predictions of future climate (Barrow and Yu, 2005).

The scenarios used in this study demonstrate that there is considerably less variability between GCM predictions of future temperature relative to predictions of future precipitation. Hence, we are relatively confident in temperature predictions by the GCMs but have less confidence in the predicted changes in precipitation. Given the 1 to 4 degree increase in mean temperature already observed over the last century in North America (Schindler and Donahue, 2006), it is likely that warming will be more than 2 degrees by the 2080s. We, therefore, suggest that the lack of change in snowpack resulting from the NCAR scenario is highly unlikely even with substantial emission controls. Similar caution is important to consider in interpreting the CSIRO or CGCM2 outputs.

Each GCM is also subject to an emissions scenario, which has a significant impact on future projections. These emissions scenarios provide insight into the type of adaptation that might be required to mitigate the effects of climate change. The CSIRO and CCSR scenarios represent a case where there

would be very rapid economic growth, a growing global population that peaks in mid-century and subsequently declines, and the rapid introduction of new and more efficient technologies. The CGCM scenario represents a heterogeneous world with continually increasing global population and regionally oriented economic growth that is fragmented and slower relative to other emissions scenarios (Nakicenovic *et al.*, 2000). Based on results from the CSIRO, CCSR and CGCM scenarios we suggest that even under substantial adaptation, mean and maximum annual snowpack over the St. Mary River watershed will likely decline.

The NCAR scenario represents a world where the emphasis is on local solutions to economic, social and environmental sustainability, with intermediate levels of economic development. It also assumes increases in population, however, at a lower population growth rate than the A2 scenario (Nakicenovic *et al.*, 2000). Model results from the NCAR scenario show that significant changes to current social and environmental policies may mitigate the effects of climate change on available water from snowpack in the St. Mary River watershed.

### **3.5 Summary and Conclusions**

Future predictions of the potential impacts of climate change on snowpack in the St. Mary River watershed were made using the GENESYS model. Five scenarios of future climate were derived representing a range of plausible future conditions. These simulations using the GENESYS Model predict that snowpack in the St. Mary watershed is highly vulnerable to slight changes in temperature,

and to a lesser extent, changes in precipitation. We have demonstrated that future predictions are highly dependent on scenario selection. This is shown with differences between projections of mean annual and maximum annual snowpack.

Predictions of future mean annual snowpack volume differ between scenarios, with the CGCM scenario predicting a 63% decline, the CSIRO scenario predicting a 17% decline, and the NCAR scenario predicting a 3% decline in mean annual snowpack by the 2080s. Maximum snowpack values also differ between the CCSR and NCAR models with a significant decreasing trend in the CCSR prediction and no significant trend in the NCAR prediction.

Significant decreasing trends are observed in both the NCAR and CGCM scenarios, showing that the timing of the spring melt will likely change towards earlier in the season. If spring melt occurs earlier, the St. Mary watershed could experience significant reductions in late season flow. Human adaptation may help mitigate these effects of climate change on water resources. However, ecosystems remain extremely vulnerable to even slight shifts in the climatic regime (Schindler, 2001).

The high sensitivity of snow processes to changes in climate pose important questions about water resources in the future. This modelling effort provides insight into future snow conditions in the watershed but leaves much to be resolved. Current monitoring of changes in mountain ecosystems exists at a spatial and temporal scale that is inadequate. In order to properly quantify and

adapt to future changes in mountain hydrometeorology, increased monitoring in meteorological variables over time and space is required.

## **CHAPTER 4**

### **Summary, Conclusions and Recommendations**

#### **4.1 Summary and Conclusions**

The GENESYS hydrometeorological model was adapted and enhanced for estimating potential impacts of climate change on mountain snowpack in the St. Mary River watershed, Montana. The first journal paper presented the continued development of the GENESYS model and application of the model in predicting daily snowpack accumulation and ablation as well as annual runoff volume. The second journal paper described the application of GENESYS in predicting potential change in mountain snowpack under a range of climate change scenarios.

Spatial surfaces of hydrometeorological variables were estimated at a fine spatial scale over the entire St. Mary River watershed for 82 individual TCs derived from elevation and land cover data. For each of the TCs daily meteorological variables and daily water balance for both a snow covered and open surfaces were calculated. The inclusion of canopy interception, sublimation, evapotranspiration, soil water, and improved precipitation elevation functions enabled more representative estimates of hydrometeorological variables over the watershed.

The hydrometeorological surfaces were assessed using an integrated annual water balance. Simulated annual runoff volume compared well with the annual streamflow volume at the international border gauging station with an

overestimate of 11% over the 1961-90 time period. Daily snowpack (mm SWE) values at the Many Glacier SNOTEL site were well simulated, with strong agreement over the 10 year trial period ( $r^2=0.824$ ). The simulations show that the GENESYS model is able to estimate daily snowpack accumulation and ablation with reasonable accuracy. The verification of GENESYS outputs provided confidence in the model and showed that the model could be applied to make estimates of future snowpack under climate change scenarios.

Monthly GCM outputs of temperature and precipitation were downscaled using the “delta” method for five scenarios. The 5 scenarios selected represent a broad range of plausible future climates. All scenarios were consistent with an increase in temperature over the region but predictions of precipitation were considerably more variable. For each of the 5 scenarios, the 1961-90 daily temperature and precipitation record at the St. Mary climate station was perturbed to create future climate datasets for the thirty year period centered on the 2020s, 2050s and 2080s.

The perturbed climate record was used to drive the GENESYS model and provide estimates of potential change in snowpack over the St. Mary basin for each of the 5 climate change scenarios. There is agreement among all scenarios that snowpack in the St. Mary basin is highly sensitive to changes in temperature. This shows the susceptibility of snowpack to climate change even where GCMs predict increases in precipitation.

All scenarios predict an increase in the ratio of rain to snow, demonstrating the effect of temperature on winter precipitation. The effects of temperature are also shown with a comparison of mean annual snowpack, where the CSIRO, CGCM, and NCAR scenarios are all consistent with a substantive decrease in mean annual snowpack at lower elevations. This is an expected change as warming occurs in large areas of the watershed at lower elevations where historical air temperatures were just below freezing for much of the winter. Climate warming converts much of what was historically snow to rain, causing substantial changes in snowpack, and runoff.

An important measure of the effect of climate change on mountain snowpack is the timing of the onset of spring, as this could have important implications for both ecosystems and hydrology (Barnett et al., 2005). An earlier date of maximum snowpack in all scenarios shows that as temperatures increase the model predicts there will be an earlier onset of spring in the future. With an earlier onset of spring, the growing season will be longer and it is likely that late summer water supply will be reduced.

Results from mean annual snowpack changes show the sensitivity of climate change projections to scenario selection and the subjectivity in assessments of future conditions. The NCAR scenario shows no significant change in mean annual snowpack, while the CSIRO and CGCM scenarios both show significant changes in mean annual snowpack. This can be explained by the relative differences between scenario projections of temperature and precipitation. Projections of temperature and precipitation from each scenario

can be attributed to the emissions scenario used to drive the GCMs, which provides insight into what can be expected under different levels of adaptation.

Results show that even in a world where significant adaptation occurs, population increases and subsequent increases of greenhouse gases will lead to changes in the winter hydrological regime of the St. Mary River watershed. If the timing of available water changes, managers will need to adopt strategies to handle larger peak flows and lower late season flows. Greater variability may create more management problems that reduced water supply. If an earlier onset of spring and changes in the flow regime of the watershed result in reduced late season water supply, there will be significant impact to ecosystems. Ecosystem thresholds could be breached where populations are significantly altered and native species are replaced by introduced or invading species. It is likely that human adaptation could help mitigate the effects of these changes on human populations; however, ecosystems will be heavily impacted.

#### **4.2 Recommendations**

This study has demonstrated improvements made to the GENESYS model and the models strength in determining winter snowpack. However, further development of the model should include a streamflow routing routine and groundwater routines and a more detailed soil routine. Future versions of the model should also include temperature routines that are able to account for the effects of albedo.

The most significant limitation to modelling in mountainous environments is the lack of representative data. There is a lack of measurement of important hydrometeorological variables over both time and space, with a bias in measurements at low elevation, populated regions. The measurement of hydrometeorological variables at a wider range of spatial and temporal scales is essential. Improvements that may be instituted soon include:

- Snow course measurements – these are cost effective and should be included over a wider range of mountain watersheds to increase the understanding of the effects of elevation on precipitation in the winter.
- Snow telemetry sites – are needed in study watersheds where all researchers have access to the data.
- A temperature sensor network – can be installed at a relatively low cost to improve the understanding of cold air drainage and inversions in mountain regions. These processes are important for accurate modelling in mountainous terrain.
- Remote sensing has been widely used to provide spatial measurements of land cover, soils, and snow extent. Further research should be directed at including remotely sensed data in mountain hydrology studies.

Current impacts research is also limited by the unavailability of future climate data at appropriate spatial and temporal scales. In order to assess the impacts of climate change on mountain hydrology, improved downscaling methods and the incorporation of changes in climatic variability are required. Using regional climate model (RCM) output at a daily time step may provide a

means to assess future variability at a scale that is more representative.

However, these data are not always available for a wide range of climate models and scenarios. With the advent of a data network that includes a range of RCM scenarios, these studies could be conducted.

The historical time periods currently used for impacts assessment are 1961-90 or 1971-2000. This is a relatively short time period in a climatic context and does not account for long-term variability. The change in the mean climatic condition poses one challenge for ecosystems and water supplies. However, climatic variability and extremes may create the worst impacts on water users. Paleoclimatic datasets derived from tree rings are currently available for mountainous regions. These datasets can be used to verify GCM performance in simulating long-term variability (Sauchyn, personal communication). If GCMs are able to capture the long-term variability in paleoclimatic datasets then relative change from a long-term historical record could provide insight into the types of adaptive strategies that would be required to ensure human and ecosystem health.

## REFERENCES

- Alberta Environment. (1998). South Saskatchewan River Basin Historical Weekly Natural Flows 1912 to 1995, Main Report. 193pp.
- Anderton, S. P., White, S. M., and Alvera, B. (2004). Evaluation of spatial variability in snow water equivalent for a high mountain catchment. *Hydrological Processes*, **18**, 435-453.
- Andreasson, J., Lindstrom, G., Grahn, G., Johansson, B. (2004). Runoff in Sweden-mapping of climate change impacts on hydrology. *Nordic Hydrological Programme*, NHP Report., **48**, 625-632.
- Arnell, N.W. (1999). Climate change and global water resources. *Global Environmental Change*, **9**, 32-49.
- Bales, R. C., Molotch, N. P., Painter, T. H., Dettinger, M. D., Rice, R., and Dozier, J. (2006). Mountain hydrology of the western United States. *Water Resources Research*, **42**, 13-21.
- Barnett, T. P., Adam, J. C., and Lettenmaier, D. P. (2005). Potential impacts of a warming climate on water availability in snow-dominated regions. *Nature*, **438**, 1-7.
- Barrow, E., and Yu, G. (2005). Climate Change Scenarios for Alberta a report Prepared for the Prairie Adaptation Research Collaborative (PARC) in co-operation with Alberta Environment. Regina, Saskatchewan. 73pp.
- Basist, A., Bell, G. D., and Meentemeyer, V. (1994). Statistical relationships between topography and precipitation patterns. *Journal of Climate*, **7**, 1305-1315.
- Beniston, M., Diaz, H. F., and Bradley, R.S. (1997). Climatic change at high elevation sites: an overview. *Climatic Change*, **36**, 233-251.
- Beniston, M. (2003). Climatic change in mountain regions: A review of possible impacts. *Climatic Change*, **59**, 51-31.
- Bonsal, B. R., Prowse, T. D., and Pietroniro, A. (2003). An assessment of global climate model-simulated climate for the western cordillera of Canada (1961-90). *Hydrological Processes*, **17**, 3703-3716.

- Bristow, K. L., and Campbell, G. S. (1984). On the relationship between incoming solar radiation and daily maximum and minimum temperature. *Agriculture and Forest Meteorology*, **31**, 159-166.
- Burn, D. H. (1994). Hydrologic effects of climate change in west-central Canada. *Journal of Hydrology*, **160**, 53-70.
- Burn, D. H., and Hag Elnur, M. A. (2002). Detection of hydrologic trends and variability. *Journal of Hydrology*, **255**, 107-122.
- Calder, I. R. (1990). *Evaporation in the Uplands*. Wiley, Chichester, 144 pp.
- Cayan, D. R. (1996). Interannual climate variability and snowpack in the western United States. *Journal of Climate*, **9**, 928-948.
- Cayan, D. R., Kammerdiener, S. A., Dettinger, M. D., Caprio, J. M., and Peterson, D. H. (2001). Changes in the onset of spring in the western United States., *Bulletin of the American Meteorological Society*, **82**, 399-415.
- Clair, T. A., Ehrman, J., and Higuchi, K. (1998). Changes to the runoff of Canadian ecozones under a doubled CO<sub>2</sub> atmosphere. *Canadian Journal of Fisheries and Aquatic Science*, **55**, 2464-2477.
- Cohen, S. J., Neilsen, D., Smith, S., Neale, T., Taylor, B., Barton, M., Merritt. W., Alila, Y., Shepherd. P., McNeil, R., Tansey, J., Carmichael, J., and Langsdale, S. (2006). Learning with local help: Expanding the dialogue on climate change and water management in the Okanagan region, British Columbia, Canada. *Climatic Change*, **75**, 331-358.
- Daly, C., Smith, J.W., Smith, J.I., and McKane, R.B. (2007). High-resolution spatial modeling of daily weather elements for a catchment in the Oregon Cascade Mountains, United States. *Journal of Applied Meteorology and Climatology*, **46**, 1565-1586.
- Daly, C., Halbleib, M., Smith, J.I., Gibson, W.P., Doggett, M.K., Taylor, G.H., Curtis, J., and Pasteris, P. P. (2008). Physiographically sensitive mapping of climatological temperature and precipitation across the conterminous United States. *International Journal of Climatology*, DOI: 10.1002/joc.1688.
- Daly, C. (2006). Guidelines for assessing the suitability of spatial climate data sets. *International Journal of Climatology*, **26**, 707-721.

- Daly, C., Gibson, W.P., Taylor, G.H., Johnson, G.L., and Pasteris, P.P. (2002). A knowledge-based approach to the statistical mapping of climate. *Climate Research*, **22**, 99-113.
- Daly, C., Neilson, R. P., and Phillips, D. L. (1994). A statistical-topographic model for mapping climatological precipitation over mountainous terrain. *Journal of Applied Meteorology*, **33**, 140-158.
- Déry, S. J., Taylor, P., and Xiao, J. (1998). The thermodynamic effects of sublimating, blowing snow in the atmospheric boundary layer. *Boundary-Layer Meteorology*, **89**, 251-283.
- Déry, S. J., and Yau, M. K. (1999). A bulk blowing snow model. *Boundary-Layer Meteorology*, **93**, 237–251.
- Déry, S. J., and Taylor, P. A. (1996). Some aspects of the interaction of blowing snow with the atmospheric boundary layer. *Hydrological Processes*, **10**, 1345-1358.
- Diaz, H. F. (2005). Monitoring climate variability and change in the western United States. Bugmann, H. K. M., and Reasoner, M. A. (Eds.), *Global Change and Mountain Regions*, Springer, Dordrecht, The Netherlands, 636-656.
- Dingman, S. (2002). *Physical Hydrology*, 2<sup>nd</sup> edition. Prentice-Hall, Upper Saddle River, NJ. 726 pp.
- Elder, K., Dozier, J. M., Michaelsen, J. (1991). Snow accumulation and distribution in an alpine watershed. *Water Resources Research*, **27**, 1541–1552.
- Epstein, D., and Ramirez, J.A. (1994). Spatial disaggregation for studies of climatic hydrologic sensitivity. *Journal of hydraulic Engineering*, **120**, 1449-1467.
- Fagre, D. (2006). Unpublished USGS data from the Climate Change in Mountain Ecosystems project at Glacier National Park. *in*. United States Geological Survey, West Glacier Field Station.

- Field, C.B., Mortsch, L.D., Brklacich, M., Forbes, D.L., Patz, J.A., Running, S.W. and Scott, A. (2007). North America. Parry M.L., Canziani, O. F., Palutikof, J. P., van der Linden, P. J., and Hanson, C. E. (Eds.), *Climate Change 2007: Impacts, Adaptation and Vulnerability. Contribution of Working Group II Fourth Assessment Report of the Intergovernmental Panel on Climate Change*, Cambridge University Press, Cambridge, Uk., 617-652.
- Fontaine, T. A., Cruickshank, T. S., Arnold, J. G., and Hotchkiss, R. H. (2002). Development of a snowfall-snowmelt routine for mountainous terrain for the soil water assessment tool (SWAT). *Journal of Hydrology*, **262**, 209-223.
- Frederick, K. D., and Major, D. C. (1997). Climate change and water resources. *Climatic Change*, **37**, 7-23.
- Fyfe, J.C., and Flato, G.M. (1999). Enhanced climate change and its detection over the Rocky Mountains. *Journal of Climate*, **12**, 230-243.
- Gelfan, A. N., Pomeroy, J. W., and Kuchment, L. S. (2004). Modeling forest cover influences on snow accumulation, sublimation, and melt. *Journal of Hydrometeorology*, **5**, 785-803.
- Giorgi, F. N., and Mearns, L. O. (1991). Approaches to the simulation of regional climate change: A review. *Reviews of Geophysics*, **29**, 191-216.
- Glassy, J. M., and Running, S. W. (1994). Validating diurnal climatology logic of the MT-CLIM model across a climatic gradient in Oregon. *Ecological Applications*, **4**, 248-257.
- Hamlet, A. F., and Lettenmaier, D. P. (1999). Effects of climate change on hydrology and water resources in the Columbia basin. *Journal of the American Water Resources Association*, **35**, 1597-1623.
- Hamlet, A. F., Mote, P. W., Clark, M. P., and Lettenmaier, D. P. (2005). Effects of temperature and precipitation variability on snowpack trends in the western United States. *Journal of Climate*, **18**, 4545-4561.
- Hanson, C. L. (1982). Distribution and stochastic generation of annual and monthly precipitation on a mountainous watershed in southwest Idaho. *Water Resources Bulletin*, **18**, 875-883.

- Hay, L. E., Wilby, R. L., Leavesley, G. H. (2000). A comparison of delta change and downscaled GCM scenarios for three mountainous basins in the United States. *Journal of the American Water Resources Association*, **36**, 387-397.
- Hedstrom, N. R., and Pomeroy, J.W. (1998). Measurements and modelling of snow interception in the boreal forest. *Hydrological Processes*, **12**, 1611-1625.
- Henderson-Sellers, A. (1996). Soil moisture: A critical focus for global change studies. *Global and Planetary Change*, **13**, 3-9.
- Hood, E., Williams, M., and Cline, D. (1999). Sublimation from a seasonal snowpack at a continental, mid-latitude alpine site. *Hydrological Processes*, **13**, 1781-1797.
- Hungerford, R. D., Nemani, R. R., Running, S. W., and Coughlan, J. C. (1989). MTCLIM: mountain microclimate simulation model. Research Paper INT-414, US Department of Agriculture, Forest Service, Intermountain Research Station, Ogden, UT.
- Huntington, T. G. (2006). Evidence for intensification of the global water cycle: Review and synthesis. *Journal of Hydrology*, **319**, 83-95.
- Huntington, T.G., Hodgkins, G.A., Keim, B.D., and Dudley, R.W. (2004). Changes in the proportion of precipitation occurring as snow in New England (1949-2000). *Journal of climate*, **17**, 2626-2636.
- Hutchinson, M. F. (2004). ANUSPLIN version 4.3 user guide. Australian National University, Canberra. 71 pp.
- IPCC-TGCIA. (1999). *Guidelines on the Use of Scenario Data for Climate Impact and Adaptation Assessment*. Version 1. Prepared by Carter, T.R., M. Hulme, M. Lal, Intergovernmental Panel on Climate Change, Task Group on Scenarios for Climate Impact Assessment. 69 pp.
- IPCC. (2007). Climate Change 2007. Solomon, S., Qin, D., Manning, M., Chen, Z., Marquis, M., Averyt, K. B., Tignor, M., Miller, H. L. (Eds.), *The Physical Science Basis. Contribution of Working Group 1 to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change*. Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA. 996 pp.

- Karl, T. R., Groisman, P. Y., Knight, R. W., and Heim, R. R. (1993). Recent variations of snow cover and snowfall in North America and their relation to precipitation and temperature variations. *Journal of Climate*, **6**, 1327-1344.
- Karl, T. R., and Trenberth, K. E. (2003). Modern global climate change. *Science*, **302**, 1719-1723.
- Kienzle, S. W. (2008). A new temperature based method to separate rain and snow. *Hydrological Processes*, **22**, 5067-5085.
- Knowles, N., Dettinger, M. D., and Cayan, D. R. (2006). Trends in snowfall versus rainfall in the western United States. *Journal of Climate*, **19**, 4545-4559.
- Lapp, S., Byrne, J., Kienzle S. W., and Townshend, I. (2002). Linking Global Circulation model synoptics and precipitation for western North America. *International Journal of Climatology*, **22**, 1807-1817.
- Lapp, S., Byrne, J., Kienzle, S. W., and Townshend, I. (2005). Climate warming impacts on snowpack accumulation in an alpine watershed: A GIS based modeling approach. *International Journal of Climatology*, **25**, 521-526.
- Laprise, R., Caya, D., Frigon, A., and Paquin, D. (2003). Current and perturbed climate as simulated by the second-generation Canadian Regional Climate Model (CRCM-II) over northwestern North America. *Climate Dynamics*, **21**, 405-421.
- Larson, R., 2008: Modelling climate change impacts on mountain snow hydrology, Montana-Alberta. M.Sc. Unpublished Master's Thesis. University of Lethbridge, Alberta, Canada.
- Larson, R., Byrne, J. M., Johnson, D. L., and Letts, M. G. (2008). Refinement and Validation of a Spring Runoff Model for the Rocky Mountains, Northern Montana. *Submitted to the Journal of Hydrology*.
- Laternser, M., and Schneebeli, M. (2003). Long-term snow climate trends of the Swiss Alps (1931-99). *International Journal of Climatology*, **23**, 733-750.
- Leavesley, G.H. (1994). Modelling the effects of climate change on water resources-A review. *Climatic Change*, **28**, 159-179.

- Lettenmaier, D.P., Wood, A.W., Palmer, R.N., Wood, E.F., and Stakhiv, E.Z. (1999). Water resources implications of global warming: A U.S. regional perspective. *Climatic Change*, **43**, 537-579.
- Leung, L. R., and Wigmosta, M.S. (1999). Potential climate change impacts on mountain watersheds in the Pacific northwest. *Journal of the American Water Resources Association*, **35**, 1463-147.
- Leung, L. R., Qian, Y., Bian, X., Washington, W. M., Han, J., and Roads, J. O. (2004). Mid-century ensemble regional climate change scenarios for the western United States. *Climatic Change*, **62**, 75-113.
- Loukas, A., Vasiliades, L., Dalezios, N.R. (2004). Climate change implications on flood response of a mountainous watershed. *Journal of Geophysical Research*, **112**, D11124, doi:10.1029/2006JD007561.
- Lundquist, J. D., and Cayan, D. R. (2007). Surface temperature patterns in complex terrain: Daily variations and long-term change in the central Sierra Nevada, California. *Water, Air, and Soil Pollution: Focus*, **3**, 331-347.
- MacDonald, R. J., Byrne, J. M., Kienzle, S. W. (2008). A physically based daily hydrometeorological model for complex mountain terrain. *Manuscript submitted for publication*.
- Marquinez, J., Lastra, J., and Garcia, P. (2003). Estimation models for precipitation in mountainous regions: the use of GIS and multivariate analysis. *Journal of Hydrology*, **270**, 1-11.
- Martin, P. (1993). Vegetation responses and feedbacks to climate: a review of models and processes. *Climate Dynamics*, **8**, 201-210.
- Mather, J. R., and Yoshioka, G. A. (1968). The role of climate in the distribution of vegetation. *Annals of the Association of American Geographers*, **58**, 29-41.
- McFarlane, N. A., Boer, G. J., Blanchet, J. P., and Lazare, M. (1992). The Canadian climate centre second-generation General Circulation Model and its equilibrium climate. *Journal of Climate*, **5**, 1013-1040.
- McKenzie, D., Peterson, D. W., Peterson, D. L., and Thornton, P. E. (2003). Climatic and biophysical controls on conifer species distributions in mountain forests of Washington State, USA. *Journal of Biogeography*, **30**, 1093-1108.

- Mearns, L.O., Easterling, W., Hays, C., and Marx, D. (2001). Comparison of agricultural impacts of climate change calculated from high and low resolution climate change scenarios: Part 1. The uncertainty due to spatial scale. *Climatic Change*, **51**, 131-172.
- Merritt, W. S., Alila, Y., Barton, M., Taylor, B., Cohen, S., and Neilsen, D. (2006). Hydrologic response to scenarios of climate change in sub watersheds of the Okanagan basin, British Columbia. *Journal of Hydrology*, **326**, 79-108.
- Milewska, E. J., Hopkinson, R. F., and Niitsoo, A. (2005). Evaluation of geo-referenced grids of 1961-1990 Canadian temperature and precipitation normals. *Atmosphere-Ocean*, **43**, 49-75.
- Morrison, J., Quick, M. C., and foreman, M.G.G. (2002). Climate change in the Fraser River watershed: flow and temperature projections. *Journal of Hydrology*, **263**, 230-244.
- Mote, P. W. (2003). Trends in snow water equivalent in the Pacific Northwest and their climatic causes. *Geophysical Research Letters*, **30**, 1-4.
- Mote, P. W., Hamlet, A. F., Clark, M. P., and Lettenmaier, D. P. (2005). Declining mountain snowpack in western North America. *American Meteorological Society*, **86**, 39-49.
- Mote, P. W. (2006). Climate driven variability and trends in mountain snowpack in western North America, *Journal of Climate.*, **19**, 6209-6220.
- Nakicenovic, N., Alcamo, J., Davis, G., de Vries, B., Fenhann, J., Gaffin, S. (2000). *Special Report on Emissions Scenarios: A Special Report of Working Group III of the Intergovernmental Panel on Climate Change*, Cambridge University Press, Cambridge,U.K., 599 pp. Available online at: <http://www.grida.no/climate/ipcc/emission/index.htm>.
- NOAA/NCDC. (2006). Web Climate Service. [www.ncdc.noaa.gov/oa/climate/stationlocator.html](http://www.ncdc.noaa.gov/oa/climate/stationlocator.html). National Ocean and Atmospheric Administration and National Climatic Data Center, Accessed Jun 15, 2006.
- Nolin, A., and Daly, C. (2006). Mapping “at risk” snow in the Pacific Northwest. *Journal of Hydrometeorology*, **7**, 1164-1171.
- NRCS. (2007). SNOTEL Data & Products. <http://www.wcc.nrcs.usda.gov/snow/>. Natural Resources Conservation Service, Accessed Sep 2006.

- NRCS. (2007). SSURGO Data & Products. <http://soildatamart.nrcs.usda.gov/>. Natural Resources Conservation Service, Accessed Jan 2008.
- Oglesby, R. J., and Erickson, D. J. (1989). Soil moisture and the persistence of North American drought. *Journal of Climate*, **2**, 1362-1380.
- Pigeon, K., and Jiskoot, H. (in press). Meteorological controls on snowpack formation and dynamics in the southern Canadian Rocky Mountains. *Arctic, Antarctic and Alpine Research*.
- Pomeroy, J. W., and Gray, D. M. (1995). Snowcover accumulation, relocation and management. *National Hydrology Research Institute Science Report No.7*. NHRI, Environment Canada, Saskatoon. 144pp.
- Pomeroy, J. W., Gray, D. M., Shook, K. R., Toth, B., Essery, R. L. H., Pietroniro, A., and Hedstrom, N. (1998a). An evaluation of snow accumulation and ablation processes for land surface modelling. *Hydrological Processes*, **12**, 2339-2367.
- Pomeroy, J. W., Parviainen, J. P., Hedstrom, N., and Gray, D. M. (1998b). Coupled modelling of forest snow interception and sublimation. *Hydrological Processes*, **12**, 2317-2337.
- Pomeroy, J. W., Hedstrom, N., Parviainen, J. (1999). The snow mass balance of Wolf Creek, Yukon: Effects of snow sublimation and redistribution. *Wolf Creek Research Basin – Hydrology, Ecology, Environment*. Proceeding of Wolf Creek Workshop, Whitehorse, March 5-7, 1998. NWRI publication, J.W. Pomeroy., and R.J. Granger, (Eds.), Saskatoon., 15-30.
- Quick, M. C., and Pipes, A. (1977). UBC Watershed Model. *Hydrological Sciences Bulletin*, **306**, 215-233.
- Raupach, M. R., and Finnigan, J. J. (1997). The influence of topography on meteorological variables and surface-atmosphere interactions. *Journal of Hydrology*, **190**, 182-213.
- Refsgaard, J. C., Knudsen, J. (1996). Operational validation and intercomparison of different types of hydrological models. *Water Resources Research*, **32**, 2189-2202.
- Regonda, S. K., Rajagopalan, B., Clark, M., and Pitlick, J. (2005). Seasonal cycle shifts in hydroclimatology over the western United States. *Journal of Climate*, **18**, 372-384.

- Robock, A., Vinnikov, K. Y., Srinivasan, G., Entin, J. K., Hollinger, S. E., Speranskaya, N. A., Liu, S., and Namkhai, A. (2000). The global soil moisture data bank. *Bulletin of the American Meteorological Society*, **81**, 1281-1299.
- Rood, S. B., Mahoney, J. M., Reid, D. E., and Zilm, L. (1995). Instream flows and the decline of riparian cottonwoods along the St. Mary River, Alberta. *Canadian Journal of Botany*, **73**, 1250-1260.
- Rood, S. B., Samuelson, G. M., Weber, J. K., and Wywrot, K. A. (2005). Twentieth-century decline in streamflows from the hydrographic apex of North America. *Journal of Hydrology*, **81**, 1281-1299.
- Sauchyn, D. J., Stroich, J., and Beriault, A. (2003). A paleoclimatic context for the drought of 1999-2001 in the northern Great Plains of North America. *The Geographical Journal*, **169**, 158-167.
- Sauchyn, D. J., Pietroniro, A., and Demuth, M. (2006). Upland watershed management and global change - Canada's Rocky Mountains and the Western Plains. *Fifth Biennial Rosenberg Forum on Water Policy*, Banff, Alberta.
- Schindler, D. W. (2001). The cumulative effects of climate warming and other human stresses on Canadian freshwaters in the new millennium. *Canadian Journal of Fisheries and Aquatic Sciences*, **58**, 18-29.
- Schindler, D. W., and Donahue, W. F. (2006). An impending water crisis in Canada's western prairie provinces. *PNAS*, **103**, 7210-7216.
- Schmidt, R. A. (1991). The sublimation of snow intercepted by an artificial conifer. *Agricultural and Forest Meteorology*, **54**, 1-27.
- Schmidt, R. A., Troendle, C. A., and Meiman, J. R. (1998). Sublimation of snowpacks in subalpine conifer forests. *Canadian Journal of Forest Research*, **28**, 501-513.
- Schmidt, R. A. (1982). Vertical profiles of wind speed, snow concentrations, and humidity in blowing snow. *Boundary-Layer Meteorology*, **23**, 223-246.
- Selkowitz, D. J., Fagre, D. B., and Reardon, B. A. (2002). Interannual variations in snowpack in the Crown of the Continent Ecosystem. *Hydrological Processes*, **54**, 3651-3665.

- Sheppard, A. (1996). Modelling Hydrometeorology in the Upper Oldman River Basin. M.Sc. Unpublished Master's Thesis. University of Lethbridge, Alberta, Canada.
- Stephenson, N. L. (1990). Climatic control on vegetation distribution: The role of the water balance. *The American Naturalist*, **135**, 649-670.
- Stewart, I. T., Cayan, D. R., and Dettinger, M. D. (2004). Changes in snowmelt runoff timing in western North America under a 'business as usual' climate change scenario. *Climatic Change*, **62**, 217-232.
- Stewart, I. T., Cayan, D. R., and Dettinger, M. D. (2005). Changes toward earlier streamflow timing across western North America. *Journal of Climate*, **18**, 1136-1155.
- Strasser, U., Bernhardt, M., Weber, M., Liston, G.E., and Mauser, W. (2007). Is snow sublimation important in the alpine water balance? *The Cryosphere Discussions*, **1**, 303-350.
- Thomson, A. M., Brown, R. A., Rosenberg, N. J., Srinivasan, R., and Izaurrealde, C. (2005). Climate change impacts for the conterminous USA: An integrated assessment. *Climatic Change*, **69**, 67-88.
- Thorpe, A. D., and Mason, B. J. (1966). The evaporation of ice spheres and ice crystals. *British Journal of Applied Physics*, **17**, 541-548.
- Todini, E. (1988). Rainfall-runoff modeling-past, present and future. *Journal of Hydrology*, **100**, 341-352.
- USGS. (2000). National Land Cover Data for Montana. <http://nris.mt.gov/nsdi/nris/nlcdgrid.html> United States Geological Survey, Accessed Feb 5, 2006.
- Valiantzas, J. D. (2006). Simplified versions for the Penman evaporation equation using routine weather data. *Journal of Hydrology*, **331**, 690-702.
- Von Hoyningen-huene, J. (1983). Die interzeption des Niederschlages in landwirtschaftlichen pflanzenbestanden. Duetscher Verband fur Wasserwirtschaft und Kulturbau, **57**, 1-66.
- Von Storch, H., Zorita, E., Cubasch, U. (1993). Downscaling of global climate change estimates to regions scales: an application to Iberian winter rainfall. *Journal of Climate*, **6**, 1161-1171.

- Wever, L. A., Flanagan, L. B., and Carlson, P. J. (2002). Seasonal and interannual variation in evapotranspiration, energy balance and surface conductance in a northern temperate grassland. *Agricultural and Forest Meteorology*, **112**, 31-49.
- Woo, M. K., and Marsh, P. (2005). Snow, frozen soils and permafrost hydrology in Canada, 1999-2002. *Hydrological Processes*, **19**, 215-229.
- Wood, A. W., Lettenmaier, D. P., and Palmer, R. N. (1997). Assessing climate change implications for water resources planning. *Climatic Change*, **37**, 203-228.
- Wyman, R. R. (1995). Modeling Snowpack Accumulation and Depletion. Guy, B. T., and Barnard, J. (Eds.), *Mountain hydrology: Peaks and valleys in research application*, Canadian Water Resources Association, 23-30.
- Xu, C. (1999). From GCMs to river flow: a review of downscaling methods and hydrologic modelling approaches. *Progress in Physical Geography*, **23**, 229-249.
- Zhang, X., Harvey, K. D., Hogg, W. D., Yuzyk, T. R. (2001). Trends in Canadian streamflow. *Water Resources Research*, **37**, 987-998.
- Zhang, X., Suzuki, K., Kadota, T., Ohata, T. (2004). Sublimation from snow surface in southern mountain taiga of eastern Siberia. *Journal of Geophysical Research*, **103**, doi:10.1029/2003JD003779.