SPATIAL AND TEMPORAL TRENDS IN NORTH AMERICAN SNOW DEPTH AND RELATIONSHIPS WITH STREAMFLOW AND ABLATION

By

JAMIE L. DYER

(Under the Direction of Thomas Mote)

ABSTRACT

This dissertation examines secular trends in North American snow depth from 1960-2000. Results show substantial decreases in snow depth across much of the continent during the late winter and early spring, with the most pronounced negative trends in central Canada since the mid 1980s. These findings support previous research documenting decreased snow cover extent and duration using satellite observations since 1966. Decreased March snow depth is shown to be related to the increased frequency and intensity of snow ablation, suggesting an earlier onset of spring melt. A significant positive trend in the frequency of warmer air masses over central Canada during March is related to the increased frequency of ablation events. The change in frequency of warmer air masses is also associated with an increase in sensible heat advection over this region. As snowmelt runoff resulting from snow ablation is the major source of water for many major North American watersheds, the relationship between snow volume (snow depth over a unit area) and hydrology for five major North American watersheds is examined. Results indicate that snow volume provides important information on spring streamflow in the Yukon, Mackenzie and Saskatchewan Rivers.

INDEX WORDS: North America, snow depth, climate change, snow ablation, snow hydrology.

SPATIAL AND TEMPORAL TRENDS IN NORTH AMERICAN SNOW DEPTH AND RELATIONSHIPS WITH STREAMFLOW AND ABLATION

by

JAMIE L. DYER

B.S., University of Georgia, 1999

M.S., University of Georgia, 2001

A Dissertation Submitted to the Graduate Faculty of The University of Georgia in Partial Fulfillment of the Requirements for the Degree

DOCTOR OF PHILOSOPHY

ATHENS, GEORGIA

2005

SPATIAL AND TEMPORAL TRENDS IN NORTH AMERICAN SNOW DEPTH AND

RELATIONSHIPS WITH STREAMFLOW AND ABLATION

by

JAMIE L. DYER

Major Professor: Thomas L. Mote

Committee: Andrew Grundstein
John Knox
Vernon Meentemeyer
Todd Rasmussen

Electronic Version Approved:

Maureen Grasso
Dean of the Graduate School
The University of Georgia
May 2005

DEDICATION

I dedicate this work to a wonderful mother, wife, sister, and friend, in memory of all that she has done for this world during her short time here. Although we all miss her terribly, it is a comfort to know that she is at peace, and that she is and will always be patiently looking down on us with unending love and affection. I know that we will see her again when it is our time to go Home.

May you rest in peace in God's merciful arms.

In memory of Michalina Dębicka,

Born to this world – September 29, 1953

Called to God – January 30, 2005

ACKNOWLEDGEMENTS

First and foremost, I would like to thank God, family, and friends for sticking with me while I have been writing this dissertation. My wife, Ania, who I love more than anything in this world, I owe more than I can ever repay. You have been patient and understanding, no matter what, and that has helped to keep me calm and confident when I thought I would never be able to make it. My family, both in the US and in Poland, I can't even begin to describe how much you all mean to me. I am forever thankful to have such love and support.

I would also like to thank everyone in the Department of Geography for helping me with the completion and submission of this dissertation. Everyone in the front office who has helped with the random questions that no one else could answer, and to those in the Climate Lab who kept me laughing and motivated at the same time. I would especially like to thank those on my committee for their help and support, and also for their guidance in writing and research. Lastly, I would like to thank my major professor, Tom Mote, for his patience and hard work in helping me to make this dissertation a reality.

This research was partially supported by grants from the National Aeronautics and Space Administration (NAG511592) and National Oceanographic and Atmospheric Administration (NA04OAR4310169) to the University of Georgia Research Foundation, Inc., Thomas L. Mote, PI.

TABLE OF CONTENTS

LIST OF TABLES

LIST OF FIGURES

CHAPTER 1

INTRODUCTION AND LITERATURE REVIEW

Snow cover is a vital component of the Earth's climate system due to its sensitivity to changes in surface energy and moisture fluxes, especially at high latitudes (Groisman et al., 1994), as well as its response to changes in temperature (Karl et al., 1993; Brown and Goodison, 1996). As a result, snow cover is considered to be a useful indicator of climate change. Moreover, seasonal snow cover provides an important climate feedback by dramatically altering the Earth's surface albedo (Kukla, 1979). Therefore, studying the historical variability of snow cover can help explain past climate trends and better understand potential climate scenarios.

Over the Northern Hemisphere, satellite snow cover records have indicated that snow cover has been considerably less extensive since the mid 1980s (Robinson et al., 1993; Robinson, 1999; Robinson and Frei, 1997; Groisman et al., 1994; Karl et al., 1993). The greatest decreases in snow cover extent (SCE) occurred in the spring and early summer, with no corresponding change in winter snow extent. SCE over North America has shown similar changes over the past century to those of the Northern Hemisphere, based on reconstructed snow cover records obtained from weekly satellite observations (Frei and Robinson, 1999; Frei et al., 1999; Brown, 2000).

While changes in SCE are concentrated along the periphery of the snow cover, snow depth can be used to examine trends across the entirety of the seasonal snow cover. Additionally, snow depth may be used to infer snow mass; therefore, analysis of snow depth over North America should provide considerable information regarding water resources and hydrology. Several studies using conventional surface observations have examined snow depth over Canada and the Great Plains of the US (Brown and Goodison, 1996; Hughes and Robinson, 1996). However, to date no study has examined long-term trends and variability in snow depth across all snow covered areas of the North American continent.

The less extensive spring snow documented from the satellite record may be due to decreased snow accumulation during the winter or more rapid ablation of snow cover in the spring. There is evidence of increasing snowfall over the northern US and southern Canada (Mekis and Hogg, 1999; Hughes and Robinson, 1996; Groisman and Easterling, 1994; Leathers et al., 1993), which corresponds with an increase in 20th Century SCE over the mid-latitudes (below 55°N) during the winter months (Hughes et al., 1996). The increase in *winter* SCE and increased snowfall would argue against decreased snow accumulation as an explanation for the spring "snow drought". Moreover, Brown (2000) compared changes in SCE over the winter season with changes in monthly temperature trends (Jones, 1994) and found a correlation with warmer temperatures in April and cooler temperatures in November. Together these studies suggest that decreased spring snow extent is likely a result of earlier and/or more intense ablation. Yet, no studies have examined changes in ablation frequency and/or intensity across North America.

The effects of an increase in frequency and intensity of snow ablation should influence hydrology because snow ablation intensity is related to flooding potential due to the rapid influx of water during snowmelt. In high-latitude rivers, winter snow storage and spring snowmelt are strongly related to the timing and magnitude of discharge (Rango, 1997; Zhang et al., 2001; Cao et al., 2002; Yang et al., 2003). For many regions of North America, snowmelt is the dominant hydrologic event during the year, providing valuable water resources and the potential for spring flooding (e.g., Kane, 1997; Leathers et al., 1998; Dyer, 2001). As a result, changes in snow cover may lead to additional variability in the hydrologic regime within the mid- and high-latitudes of North America (Miller and Russell, 2000; Serreze et al., 2000). Also, snowmelt runoff is especially important in maintaining freshwater volumes and driving circulation in large water bodies such as the Arctic Ocean due to the large volume of water that is released during spring ablation (Aagaard and Carmack, 1989; Miller and Russell, 2000; Peterson et al., 2002). The importance of snowmelt on hydrology extends to the mid-latitudes, although the association between snow and streamflow is reduced with decreasing latitude due to a more ephemeral snow cover. Nevertheless, Roads et al. (1994) showed that river streamflow values in the northern regions of the U.S., calculated as the net stream outflow per unit hydrologic surface area, were significantly dependent on winter snow accumulation. Therefore, changes in snow depth and spring snow ablation should also be associated with changing streamflow in major North American watersheds; yet, to date, no studies have examined this relationship for large watersheds across the continent.

The purpose of this dissertation is to examine variability and trends of snow depth over North America from 1960-2000, analyze mechanisms responsible for the documented changes in snow depth, and quantify the relationship between continental-scale snow cover and hydrology of major watersheds. This dissertation is presented in three studies. Each of the three studies draws on a new dataset of daily 1° latitude x 1° longitude gridded snow depth, snowfall, temperature and precipitation for North America since 1960, created explicitly for this research. The first section (Chapter 2) examines historical patterns of snow depth and snow cover extent over North America. The primary objective is to document the spatial and temporal trends of snow depth variability across North America. Principal components analysis (PCA) is performed using monthly gridded snow depths in order to ascertain where North American snow depth varies coherently over space and time. Results provide a more comprehensive understanding of the historical trends and patterns of North American snow cover.

The second section (Chapter 3) examines ablation frequency and intensity over North America from 1960-2000 to ascertain what role changes in ablation frequency and intensity play in the snow depth trends documented in Chapter 2. In addition, areas shown to have significant trends in ablation frequency are examined to determine the responsible forcing mechanisms. These machanisms are identified by an analysis of the surface energy budget components from the NCAR/NCEP Reanalysis data (Kalnay et al., 1996) and air mass characteristics from the Spatial Synoptic Climatology 2 dataset (Sheridan, 2002).

A change in the timing and rapidity of snowmelt can have a considerable effect on hydrology and energy transfer due to shallower snow depths and more frequent or rapid snowmelt runoff, especially in the higher latitudes. The third section (Chapter 4) demonstrates the strength and form of the relationship between variations in snow volume (snow depth per area) and streamflow for five large watersheds in North America. Defining the relationship between snow volume and river discharge should lead to improved knowledge of continental-scale hydrologic processes, and should aid in determining how long-term variations in snow cover effect discharge. Research objectives include the development of basin-specific statistical models utilizing information gained from analysis of the relationships between mean and peak annual snow volume and streamflow. The inclusion of snow volume in the analysis provides information regarding available water within the snowpack, which improves the predictive capability of the statistical models.

Results of the research presented in the three sections of this dissertation help to identify those regions experiencing substantial changes in snow depth, which enhances our knowledge of the state of the cryospheric system in North America. Analysis of the frequency and intensity of ablation over North America provides information on why snow depths are changing, as well as defining the factors most important in initiating and sustaining ablation in regions experiencing significant trends in snow depth. Finally, by comparing snow volume with discharge in major North American watersheds, relationships between the cryospheric and hydrologic systems can be quantitatively identified. Statistical models of snow depth and streamflow specific for each watershed may aid in the estimation of water resources and freshwater influx into the Arctic Ocean.

CHAPTER 2

SPATIAL VARIABILITY AND TRENDS IN SNOW DEPTH OVER NORTH AMERICA[1]

---

[1] Dyer, J.D. and T.L. Mote. To be submitted to *Geophysical Research Letters*.

ABSTRACT

This study uses a gridded dataset of daily U.S. and Canadian surface observations from 1960-2000 to study historical spatial and temporal variability and trends in snow depth across North America. Analysis shows decreasing snow depths beginning in mid-winter (-0.5 cm yr^{-1}; $p < 0.01$) with little change in North American snow depths throughout the fall and early winter. This springtime decreasing trend amplifies before reaching a peak in early- to mid-April (-1.0 cm yr^{-1}; $p < 0.01$). The region showing the greatest decreases in snow depth is central Canada, with slight increases in snow depth occurring during the fall and winter in central and western Alaska. A decrease in shallow snow cover (2-10 cm) during the late spring and early fall (May and October, respectively) is evident, while the extent of deeper snowpacks (>40 cm) is shown to decrease most rapidly during mid- to late winter (March-April). A principal component analysis of monthly snow depth shows coherent regions of interannual fluctuations in snow cover along the snow-free line during early winter and early spring, as well as along the leeward side of the Rocky Mountains into northern Alberta throughout the winter. During the fall and late spring, coherent regions exist in northern Canada.

2.1 Introduction

Snow cover is a vital component of the Earth's climate system due to its interactions with energy and moisture budgets from the local to the global scale. Most notably, it displays important linkages to the radiative balance at higher latitudes (Groisman et al., 1994), and is sensitive to variations in air temperature (Karl et al., 1993; Brown and Goodison, 1996). As a result, snow cover is considered to be a useful indicator of climate change.

Over the Northern Hemisphere, satellite data have indicated that mean snow cover extent (SCE) has been considerably less extensive after the mid-1980s (Robinson et al., 1993; Robinson, 1999; Robinson and Frei, 1997; Groisman et al., 1994; Karl et al., 1993). Robinson et al. (1993) noted that the greatest negative anomalies after 1987 occurred in the spring and early summer, while during the winter no significant change was noted. Based on satellite snow cover data over North America, the variability of SCE is similar to the hemispheric records, with decreasing SCE in the 1980s (Frei and Robinson, 1999; Frei et al., 1999; Brown, 2000). Using surface observations of snow cover, several investigators have shown an increase in North American winter snow cover over most of the 20th Century (Brown and Goodison, 1996; Hughes and Robinson, 1996). Although this seems counter to results obtained using satellite information, the difference is most likely attributable to the longer time series of the surface measurements.

Snow depth provides an additional dimension for snow cover studies than snow cover extent. Trends in snow depth have been analyzed for Canada (Brown and Braaten, 1998) and Eurasia (Ye et al., 1998), and several studies have looked at snow depth in relation to climate variability (Aguado et al., 1992; Changnon et al., 1993; Cayan, 1996). Brown et al. (2003) examined snow depth over North America from 1979 to 1996. However, no studies of snow depth variations over North America over a period at least as long as the satellite record (starting in 1966) have been published.

The primary objective of this project is to analyze the spatial and temporal trends of North American snow depth. This is performed using daily snow depth data from a combined U.S. and Canadian surface observation network. Selected areas shown to have significant annual trends in snow depth are examined in more detail to provide a more thorough understanding of regional snow depth patterns. Additionally, a principal components analysis (PCA) is performed on monthly snow depths to ascertain where continental snow cover varies coherently over space and time. This project provides a comprehensive understanding of the historical trends and patterns of snow depth over North America.

2.2 Snow Depth Data and Gridding Methodology

The snow depth data used in this project include daily surface observations from the United States (U.S. Department of Commerce, 2003) and Meteorological Service of Canada (Braaten, 1996; Brown, pers. comm.). Records are available at some U.S. and Canadian sites since the late 1800s; however, a majority of the US data is available beginning in 1948 due to the modernization of the cooperative observer network and digitization of historical data (Figure 2.1). In Canada, the Cryospheric System to Monitor Global Change in Canada (CRYSYS) program (Goodison and Brown, 1997), initiated in 1995, added approximately 400 stations with varying lengths of record from 1900-1990 to the existing Digital Archive of Canadian Climate Data (Brown and Braaten, 1998; Figure 2.1).

Figure 2.2 shows that the increase in the number of US stations after 1950 is concentrated along the eastern and western seaboards of the US and the Great Plains region of the Midwest, while the increase in Canadian stations is concentrated along the US-Canada border and the west coast of British Columbia. By 1960 the Canadian daily snow depth observation sites are more evenly distributed, including stations in the far northern reaches of the country; therefore, the study period for this project begins in 1960 and extends through 2000.

Snow depth values represent an average of a series of measurements taken from a given area, with care taken to include all snow and ice layers and to avoid snow drifts; however, some local variability does exist. In addition, snow depth observations are normally taken in clearings; therefore, observations are unlikely to be fully representative of surrounding terrain and vegetation cover. The most apparent limitation of snow depth observations at the continental scale is that station locations frequently follow population patterns, resulting in observations that are often biased to lower elevations, and in the case of the Canadian network, to lower latitudes. Nevertheless, it is possible for individual snow depth observations to demonstrate regional-scale consistency, especially in regions characterized by relatively homogeneous surface characteristics, i.e., the southern plains of Canada and the Great Plains of the U.S.

a. Quality Control

The reasonableness and internal consistency of the U.S. and Canadian snow depth data were tested by subjecting all observations to a quality control routine following Robinson (1989). This process involved comparing a time series of daily snow depth measurements for a given station with associated daily snow fall, maximum and minimum temperature, and precipitation data. Any snow depth data that did not meet the quality control criteria were flagged as inconsistent and not included in the analysis. Overall, 3.3% of all potential snow depth measurements were recorded as missing in the original datasets, with less than 1.2% of the remaining non-missing data flagged by the QC routines and removed.

b. Grid Generation

Snow depth grids were created using the U.S. and Canadian daily observations to make analysis of the North American snow depth data spatially consistent. The spatial extent of the gridded area was chosen to include the U.S. and the major land area of Canada. Based on the limited spatial distribution of snow depth observing stations in northern Canada, an inverse-distance interpolation algorithm was employed that interpolated the grids on a spherical surface before projecting them onto a two-dimensional Cartesian plane (Willmott et al., 1984). Initially, 0.25°x0.25° grids were produced for each day of the 1960-2000 study period, with the bounded grid region including those areas between 53°W - 168°W longitude and 20°N-71°N latitude. The 0.25°x0.25° snow depth grids were then spatially averaged to create the final 1°x1° to minimize the effect of inconsistent snow depth observations on interpolated grid values. Although a portion of the North American study area is characterized by mountainous terrain, topography was not considered in the interpolation.

Once the daily snow depth grids were complete, they were used to create five-day mean, or pentad, grid values. This was done to minimize high frequency temporal variations in the data resulting from the interpolation procedures while still maintaining a reasonably detailed temporal resolution. The pentads were calculated relative to the first day of each year such that the first pentad includes data for January 1-5, with the 12th pentad (February 25-March 1) including February 29 on leap years. This time resolution was used instead of monthly or weekly averages because of the superior temporal resolution relative to monthly data, and because using weekly values (7-day means, e.g. NOAA snow charts) makes multi-annual analysis difficult due to the offset time periods between years.

2.3 North American Snow Depth Analysis

a. Snow Depth Climatology

Figure 2.3 illustrates the progression of the growth and decline of winter snow depths during the snow season. From pentad 60 (October 23-27) through pentad 70 (December 12-16), snow cover extent (SCE) increases quickly from northern Canada and Alaska to the central United States, accompanied by a minimal increase in snow depth. From pentad 1 through 10 (January 1-5 through February 15-19, respectively), however, the reverse is true in that SCE changes little while snow depths consistently increase throughout central Canada and Alaska. This pattern is a result of the relatively rapid migration of the mean 0°C line southward during the fall into the central U.S., followed by a moderate rate of snow depth increase over a large area of the continent. In the higher latitudes, there is no substantial melt during the winter, allowing snow depths to continually increase.

b. Snow Depth Trends

A linear regression analysis was performed on each individual 1°x1° grid cell in the study region to identify trends in snow depth over North America during 1960-2000. Results generally show little change in snow depth from pentad 60 (October 23-27) through pentad 70 (December 12-16), except for smaller areas showing locally significant decreases in northeastern Canada ($>$ 0.25 cm yr^{-1}; $p < 0.10$; Figure 2.4). As the winter progresses through the turn of the year, regions with significant negative trends greater than 1 cm yr^{-1} ($p < 0.05$) appear along the eastern and western margins of Canada. This pattern continues through March, with the areas of significant negative trends in snow depth extending to include a large area of the central Canada. From pentad 15 (March 12-16) through pentad 25 (May 1-5), the band of significant negative snow depth trends through central Canada becomes more pronounced as it slowly shifts northwards during the progression of spring melt. Within the U.S. the trends in snow depth are minimal from pentad 60 (October 23-27) through pentad 10 (February 15-19; < 1cm yr^{-1}); however, during pentad 15 (March 12-16), areas of significant negative trends in snow depth (0.5-1 cm yr^{-1}; $p < 0.05$) appear in the central Great Plains, Rocky Mountains, and northern New England regions (Figure 2.4).

c. *Seasonal Snow Cover Extent (SCE)*

To give a more complete examination of the seasonal patterns of snow depth over North America, the analysis described in the previous section is augmented by an analysis of snow cover extent (SCE). SCE is examined using various snow depth threshold levels, defined as the areal extent of snow cover deeper than a given value. For example, a 10 cm threshold level would include that area in North America overlain by a snowpack with a depth of at least 10 cm. By including this type of analysis over a range of threshold levels from 2 cm to 100 cm, information regarding the overall change in depth of the snow cover, as well as the change in SCE, can be obtained.

Results of the SCE analysis, including the mean and trends at various threshold snow depth levels, are shown in Figure 2.5. Mean SCE decreases and the timing of the peaks are later as the snow depth threshold value increases (Figure 2.5a). Although the peak mean SCE decreases roughly logarithmically from a 2 cm threshold (15,679 km^2) to a 100 cm threshold

(500 km^2), the timing of the maximum mean SCE increases from pentad 3 to pentad 14, respectively (Table 2.1). This shows that a substantial area of North America is covered by a relatively homogeneous snowpack of less than 10 cm early in the snow season, and that an overall deepening of continental snow cover occurs after the initial snowpack has reached its maximum extent.

Analysis of the mean pentad SCE over the 1960-2000 study period shows the greatest trend at the 40 cm snow depth level (-67.5 km^2 yr^{-1}, $p < 0.01$; Table 2.1; Figure 2.5b), with a second peak at 2 cm (-60.6 km^2 yr^{-1}; $p < 0.01$). These results show that North American snow cover is decreasing in depth as well as extent, with the highest negative trends in snow depth occurring in areas with deep winter snowpacks, including central Canada.

2.4 Principal Component Analysis (PCA)

A principal components analysis (PCA) was performed on the gridded snow depth data to find regions within North America characterized as having spatially coherent interannual fluctuations in snow depth over time. The PCA was completed using a modified version of the gridded snow depth data set used in the previous section. The modifications consisted of removing every other 1°x1° grid point such that the North American continent consisted of 716 grid points instead of 2891, as well as using monthly mean values of snow depth instead of daily or pentad values. This was performed in order to decrease the size of the correlation matrix and increase the efficiency of the computational analysis. Although the modification to the gridded snow depth data alters the temporal and spatial resolution of the original data set, the regional patterns of continental snow depth remain intact subsequent to the reduction of grid resolution.

The snow depth data set was separated into monthly S-mode arrays, wherein columns represented individual grid cells and rows represented annual snow depth time series for each grid point. By using an S-mode design, the PC loadings yield information on the spatial structure of the snow depth field, while the PC scores provide information on the temporal variability of snow depth within the defined coherent regions. The final step in preparing the snow depth data for PCA involved calculating the correlation matrix from the associated S-mode array, which standardized the snow depth data by removing means and standard deviations from individual grid time series.

Only those PCs explaining greater than 5% of the variance were retained, which represents a practical lower limit for identifying coherent snow cover response regions (Brown, 1995; Brown and Braaten, 1998). Normally the number of PCs to retain is based on the eigenvalues or the shape of the associated scree plot; however, in this case too many PCs would be retained. A list of the PCs retained for analysis for each month is shown in Table 2.2. Upon obtaining the PC loadings for snow depth for each month, the retained PCs were subjected to a Varimax rotation to avoid difficulties with unrotated loadings, including domain shape dependence and inaccurate portrayal of the physical relationships embedded within the input dataset (Richman, 1986). A Varimax rotation was used because it is generally accepted as the most accurate analytic algebraic orthogonal rotation when applied to "known" data sets (Richman, 1986), and because of its wide use in continental and regional analysis of SCE and snow depth (Hughes and Robinson, 1996; Leathers and Luff, 1997; Brown and Braaten, 1998; Frei and Robinson, 1999).

a. *Analysis of Coherent Regions*

The location of the regions where monthly snow depths vary coherently within North America is shown in Figure 2.6. For October and November, the rotated PCs that explain the most variance tend to be located in areas where the snow line is progressing most rapidly (northern and central Canada; Figure 2.3), indicating that the variability of snow depth in this region dominates the variability throughout North America. During December, the rotated PCs become positioned longitudinally through the central prairie regions of North America due to the rapid increase of snow depth in the northern prairies and the ephemeral effect of the leeward Chinook winds in the southern prairies of Canada and the U.S. Great Plains. These patterns are similar to those of Brown (1995), who found that the dominant region of North American winter snow cover variability was centered over the Canadian Prairies and Great Plains region of the U.S.

In April, when seasonal temperature increases dominate the continent, and the area south of 55°N is snow free except for some high elevation regions in the mountain west, the location of the rotated PCs remains along the retreating snow line through central Canada and into southern Alaska. Due to local climatic influences resulting from varying topographic and surface characteristics, these areas remain relatively close in proximity but differ in overall significance and magnitude of variability.

b. *Analysis of Score Time Series*

Based on the location of the PCs, nine regions were recognized that best grouped the individual rotated PCs found over the seven months used in this study. To best illustrate the trends of snow depth over the study period within each defined area, a regression analysis was performed on the score time series of the associated rotated PCs (Table 2.3). The largest trends in snow depth occurred during the fall and spring, and showed a decrease in snow depth across the North American continent. During the fall, snow depth in Quebec and Northern Canada showed negative trends over the study period, with the largest trends occurring in November (-0.55 and -1.12 cm yr^{-1}, respectively, $p < 0.05$). In March the negative trends continued in central Canada (-1.12 cm yr^{-1}), and extended into northern Canada during April (-1.44 cm yr^{-1} and -1.25 cm yr^{-1}, respectively).

Select snow depth score time series for regions characterized as having unique and important patterns in snow depth are shown in Figure 2.7. During March, component scores within the Quebec region are higher during the 1960s than at any other time during the study period, dropping steadily from the late 1960s to the early 1970s (Figure 2.7a). As occurs in the fall, though, variability decreases during the late 1990s and into the turn of the century. Within the central Canada region, snow depth scores increase during February throughout the 1960s and early 1970s, before decreasing through the mid 1980s (Figure 2.7b). This snow depth trend is similar to that in January ($R^2 = 0.60$) and March ($R^2 = 0.61$), indicating that this pattern extends to early winter and spring. In northern Canada during October, snow depth scores tend to remain relatively consistent through the late 1980s, before decreasing rapidly through the 1990s (Figure 2.7c). In April, snow depth scores increase from the beginning of the study period through the late 1970s before dropping rapidly into the mid 1980s. In the Alaska/Yukon region, October and December snow depth patterns show a steady increase from 1960 into the early 1970s, after which a slight decrease occurs into the early 1980s (Figure 2.7d).

PCA of North American snow depth from October through April shows significant negative trends ($p < 0.05$) to occur through central Canada during spring. These trends are not as large earlier in the winter and fall, although significant decreases are still evident in northern Canada. Examination of score time series associated with PC regions in central Canada indicates that the greatest decreases in snow depth scores occur during the 1980s during February and March. These results indicate that snow depth is decreasing over North America during the 1960-2000 study period, most notably during the 1980s, with the greatest negative trends located in central Canada during winter and early spring.

2.5 Conclusions

This project identifies the spatial and temporal trends in snow depth over North America, and quantifies the magnitude and frequency of the defined interannual variations. This is performed using a gridded dataset composed of daily snow depth measurements from United States cooperative observer sites (U.S. Department of Commerce, 2003) and Meteorological Service of Canada observation sites (Braaten, 1996) over the period 1960-2000. In addition, a principal components analysis (PCA) was performed using the snow depth dataset, which allowed for regions showing coherent, or temporally correlated, interannual fluctuations in snow depth to be defined.

Regression analysis of North American snow depth over the study period shows little change from November through January, with the exception of smaller areas of local decreases in central Quebec and the northern Mackenzie River basin. Through February these areas expand to include much of central Canada. This large area of significant decreases in snow depth (> 1 cm yr^{-1}, $p < 0.05$) reaches a maximum extent in March, at which point U.S. snow depths show a decreasing trend coinciding with the initiation of the melt season, implying an earlier onset of spring thaw. The same is true of central Canada, although the maximum negative snow depth trends occur later in April in this area because of the lag in spring temperature increases. Overall, the large areas of negative trends in snow depth over North American show that continental snow cover is decreasing consistently and substantially, with the most notable changes occurring in late winter and early spring.

Seasonal SCE analysis shows the 2 cm snow depth shield reaches its maximum extent in early January, while the 100 cm snow depth shield does not reach maximum extent until early March. This is expected since the deepest snowpacks should exist just before the onset of the spring melt period; however, trend analysis shows that SCE at all depths is decreasing most rapidly during the early to mid-spring, with the greatest change occurring in the 40 cm snow depth shield in late March and early April. This supports the argument of an early onset of spring melt, and that snow depth is decreasing substantially over North America as well as SCE.

Analysis of snow depths by region shows large coherent areas to exist in the fall and winter corresponding to the location of the snow free line and up the leeward side of the Rocky Mountains into western Canada. During early winter accumulation and spring ablation, however, the coherent regions exist in a longitudinal pattern, roughly centered on central Canada from Alberta through western Manitoba. The patterns in the associated score time series show that snow depth is decreasing most rapidly in central Canada and Quebec during the 1960-2000 study period (> 1 cm yr^{-1}; $p < 0.05$), with the greatest negative trends in snow depth occurring in central Canada in February. The most substantial decreases in snow depth occur during the mid-1970s through the 1980s, with both the Alaska/Yukon and central Canadian regions showing this trend in the winter and spring.

The results of this study shed light on the spatial patterns and trends of North American snow depth, and provide evidence that continental changes in snow depth are occurring. These changes may have repercussions in regional hydrologic systems due to a change in the availability and release of snowmelt runoff, effecting water resources and freshwater flux into the Arctic Ocean. Additionally, the radiative balance may be affected at higher latitudes by reduced continental snow depths due to variations in net radiation resulting from shallower snowpacks. The fact that the greatest changes in snow depth occurred in central Canada, where winter snow cover is a dominate factor in both climatic and hydrologic systems, is particularly relevant. Future research should be performed on the climatological causes and effects of changes in North American snow depth to better understand the associated mechanisms responsible for the documented trends. This includes analysis of temperature and precipitation fields, synoptic storm tracks, and global teleconnection patterns. In addition, because of the greater availability of SCE data but the greater detail of snow depth data, correlations between the two at a regional scale should be carried out such that changes in SCE can be used to infer associated changes in snow depth.

REFERENCES

Aguado, E., D. Cayan, L. Riddle, and M. Roos, 1992: Climatic fluctuations and the timing of West Coast streamflow. *J. Climate*, **5**, 1468-1483.

Braaten, R. O., 1996: An updated Canadian snow depth database. Contract report prepared for Atmospheric Environment Service, September, 1996, 24 pp.

Brown, R. D., 1995: Spatial and temporal variability of North American snow cover, 1971-1992. *Proc. of the Eastern Snow Conference, 52nd Annual Meeting*, June 6-8, 1995, Toronto, Ontario, pp. 69-78.

Brown, R. D., 2000: Northern Hemisphere snow cover variability and change, 1915-97. *J. Climate*, **13**, 2339-2355.

Brown, R.D., 2003: Canadian snow depth data. Personal communication.

Brown, R.D. and R.O. Braaten, 1998. Spatial and temporal variability of Canadian monthly snow depths, 1946-1995. *Atmos.-Ocean*, **36**, 37-45.

Brown, R.D. and B.E. Goodison, 1996. Interannual variability in reconstructed Canadian snow cover, 1915-1992. *J. Climate*, **9**, 1299-1318.

Brown, R.D., B. Brasnett, and D. Robinson, 2003: Gridded North American monthly snow depth and snow water equivalent for GCM evaluation. *Atmos.-Ocean*, **41**, 1-14.

Cayan, D. R., 1996: Interannual climate variability and snowpack in the western United States. *J. Climate*, **9**, 928-948.

Changnon, D, T. B. McKee, and N. J. Doesken, 1993: Annual snowpack patterns across the Rockies: long-term trends and associated 500-mb synoptic patterns. *Mon. Wea. Rev.*, **121**, 633-647.

Frei, A. and D. A. Robinson, 1999: Northern Hemisphere snow extent: regional variability 1972-1994. *Int. J. Climatol.*, **19**, 1535-1560.

Frei, A., D. A. Robinson, and M. G. Hughes, 1999: North American snow extent: 1900-1994. *Int. J. Climatol.*, **19**, 1517-1534.

Goodison, B. E. and R. D. Brown, 1997: CRYSYS – Use of the Cryospheric System to Monitor Global Change in Canada. *In*: Proc. GER'97 (Int. Symp.: Geomatics in the Era of Radarsat), 24-30 May 1997, Ottawa, Ontario. Canadian Geomatics CD-ROM Vol. 1, Natural Resources Canada, Paper 259, 13 pp.

Groisman, P. Y., T. R. Karl, and R. W. Knight, 1994: Changes of snow cover, temperature, and radiative heat balance over the Northern Hemisphere. *J. Climate*, 7, 1633-1656.

Hughes, M.G. and D.A. Robinson, 1996: Historical snow cover variability in the Great Plains region of the USA: 1910 through to 1993. *Int. J. Climatol.*, **16**, 1005-1018.

Karl, T. R., P. Y. Groisman, R. W. Knight, and R. Heim Jr., 1993: Recent variations of snow cover and snowfall in North America and their relation to precipitation and temperature variations. *J. Climatol.*, **6**, 1327-1344.

Leathers, D. J. and B. L. Luff, 1997: Characteristics of snow cover duration across the northeast United States of America. *Int. J. Climatol.*, **17**, 1535-1547.

Richman, M. B., 1986: Rotation of principal components. *J. Climatol.*, **6**, 293-335.

Robinson, D. A., K. F. Dewey, and R. R. Heim, 1993: Global snow cover monitoring: an update. *Bull. Amer. Meteor. Soc.*, **74**, 1689-1696.

Robinson, D. A., 1989: Construction of a United States historical snow data base. *Proc. of the 45th Eastern Snow Conference*, 50-59.

Robinson, D. A., 1999: Northern Hemisphere snow extent during the satellite era. *Preprints: Fifth Conf. on Polar Meteorology and Oceanography*, Dallas, TX, American Meteorological Society, pp. 255-258.

Robinson, D. A. and A. Frei, 1997: A Northern Hemisphere snow cover climatology using satellite information. *Proc. of the 10th Conf. on Applied Climatology*, Reno, NV, American Meteorological Society, pp. 24-28.

U.S. Department of Commerce, 2003: TD-3200 Surface Summary of the Day, National Climatic Data Center, NOAA. [Available online: http://www.ncdc.noaa.gov].

Willmott, C.J., C.M. Rowe, and W.D. Philpot. 1984: Spheremap. Center for Climatic Research, Department of Geography, University of Delaware, Newark DE.

Ye, H., H. R. Cho, and P. E. Gustafson, 1998: Changes in Russian winter snow accumulation during 1936-83 and its spatial patterns. *J. Climate*, **11**, 856-863.

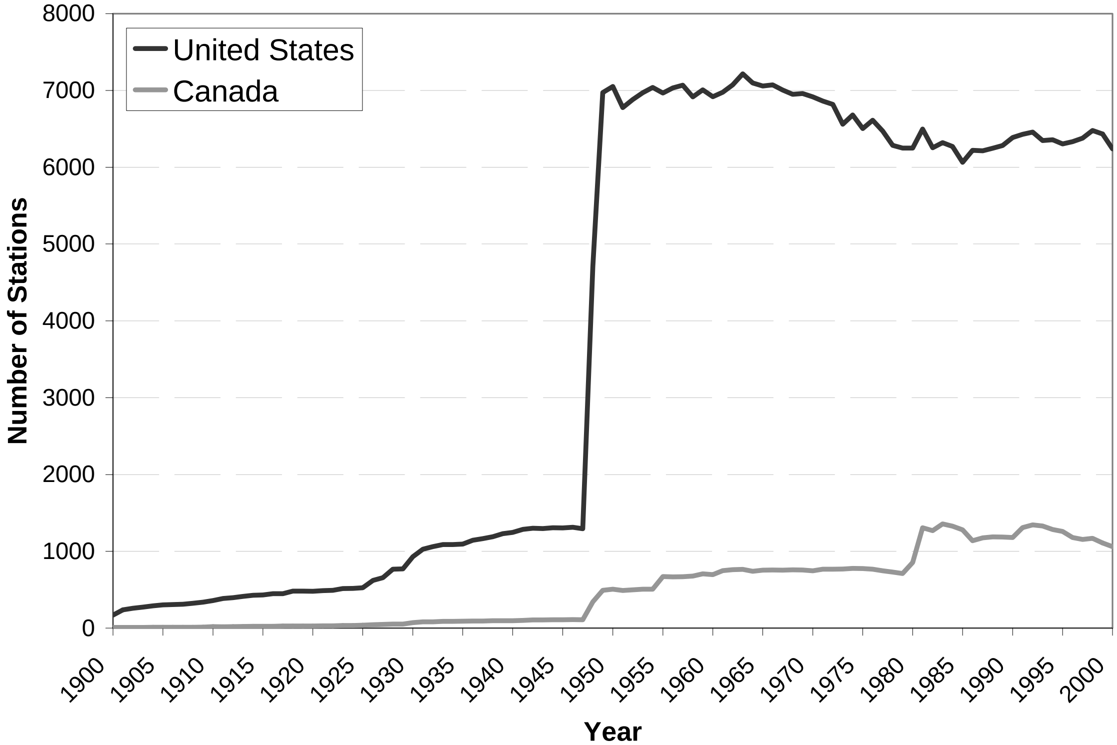


Figure 2.1. Annual average number of U.S. and Canadian daily snow depth observation sites in this study's dataset.

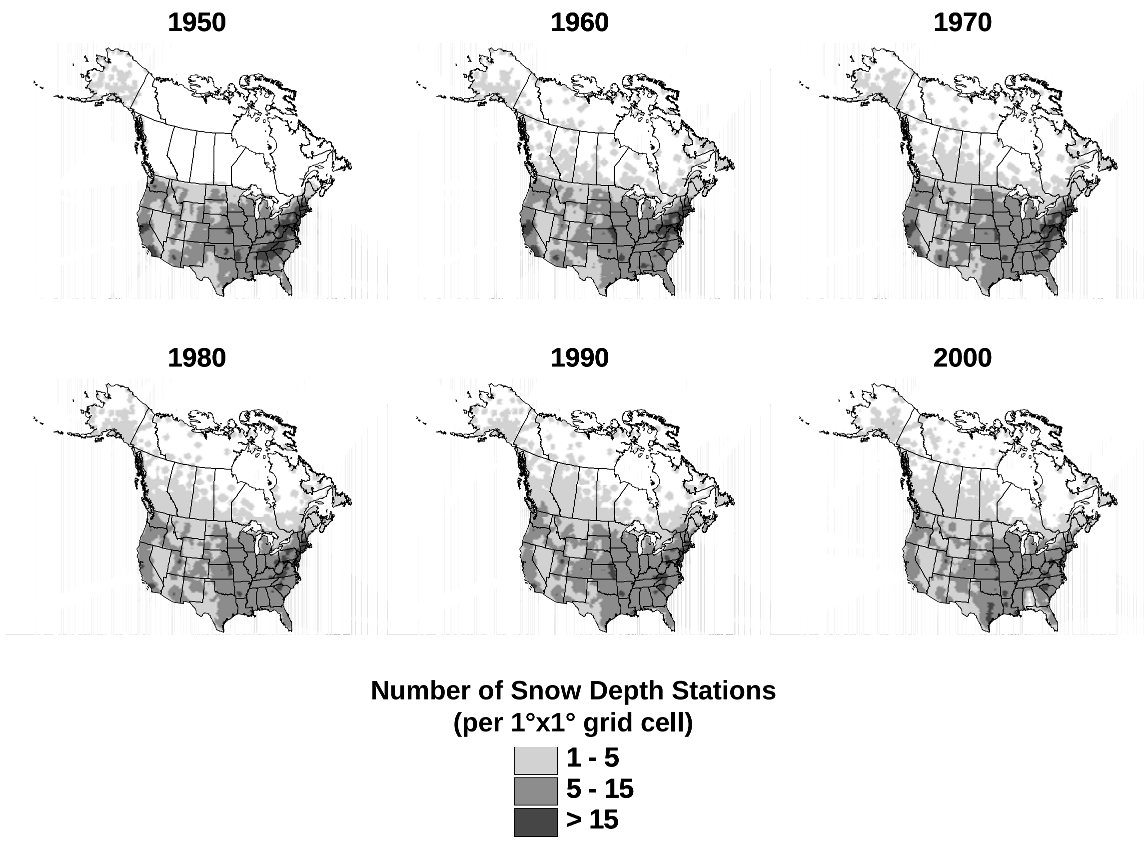


Figure 2.2. Density (number per 1°x1° area) of daily snow depth observation sites over North America for select years during the 1960-2000.

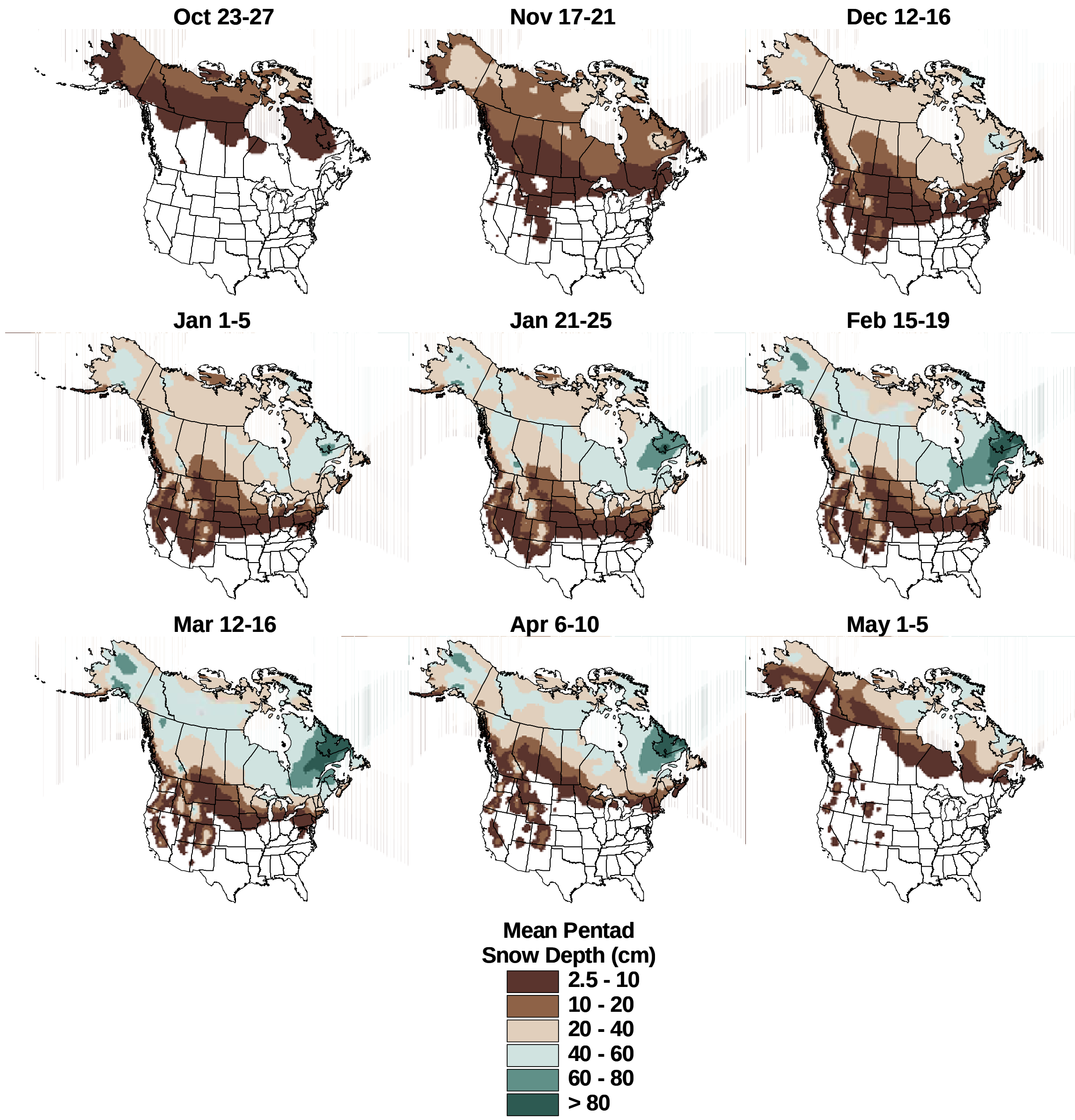


Figure 2.3. Mean North American snow depth for select pentads.

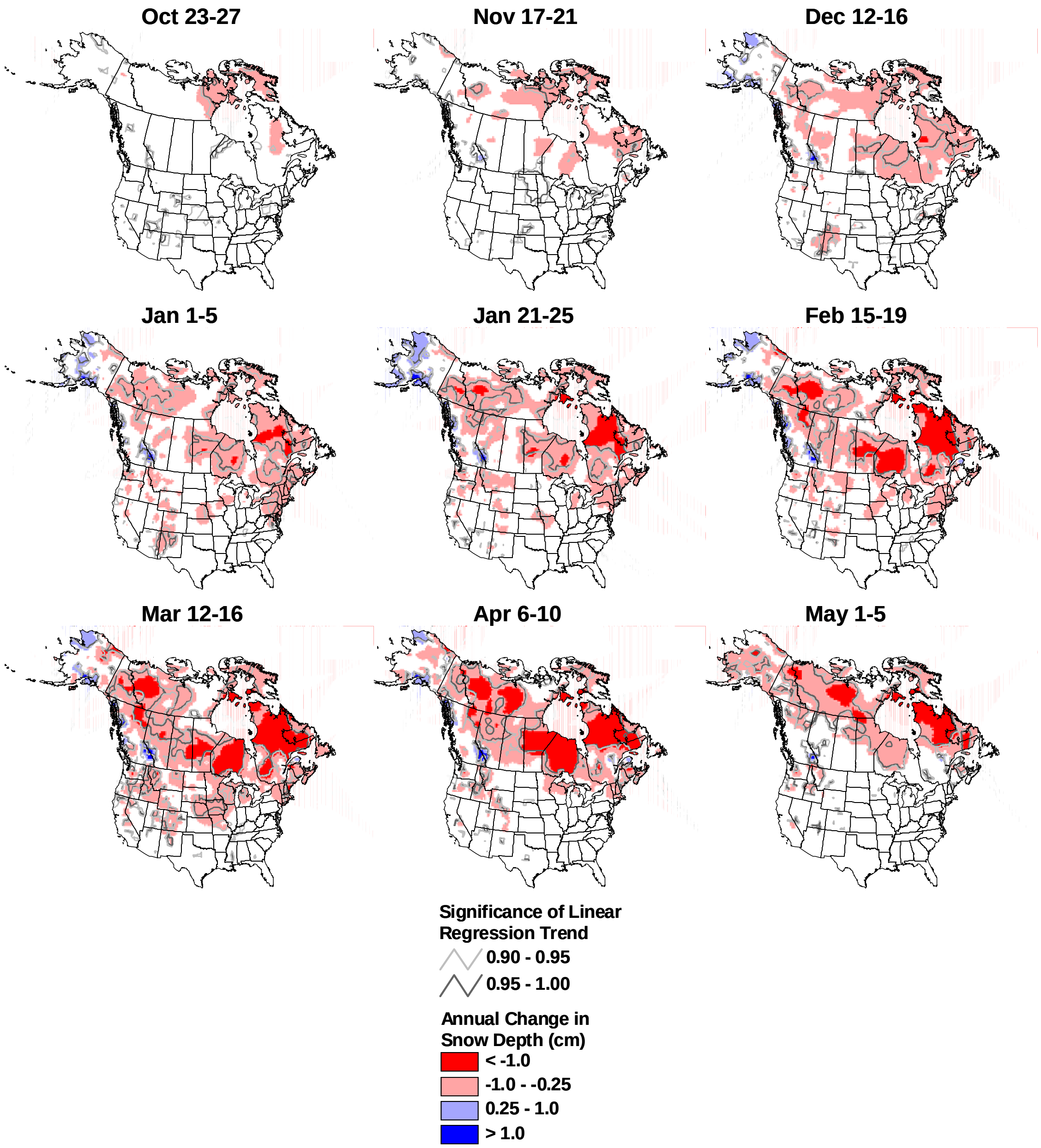


Figure 2.4. Trend in snow depth for select pentads (shaded areas), calculated using linear regression. Outlines show areas of significant trends.

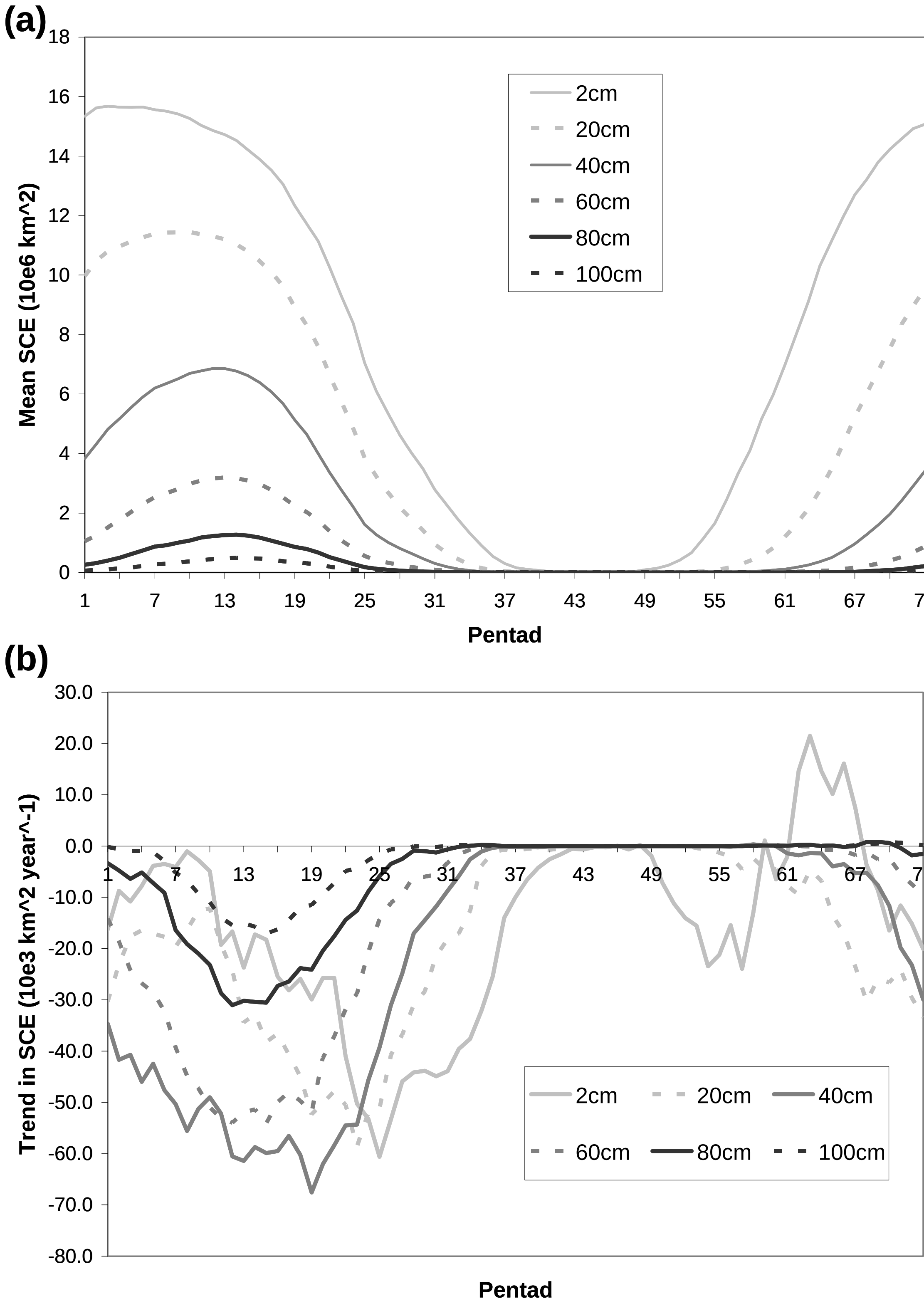


Figure 2.5. Snow cover extent (SCE) values over North America at select snow depth levels: (a) pentad means, and (b) pentad linear regressions.

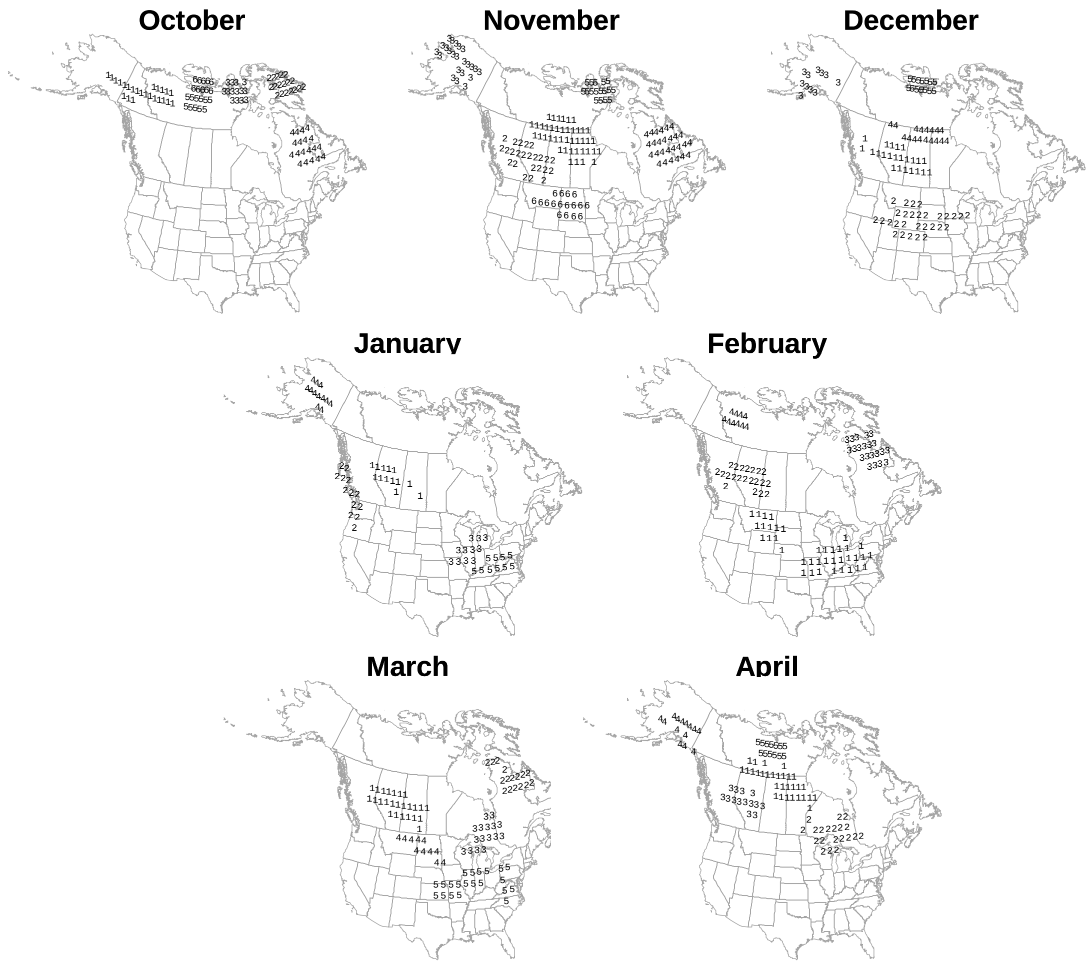


Figure 2.6. Locations of rotated PCs showing greater than 5% of the variance in monthly North American snow depth over the 1960-2000 study period. Numbers show the order of the rotated PCs and approximate areas with rotated PC loadings greater than 0.7.

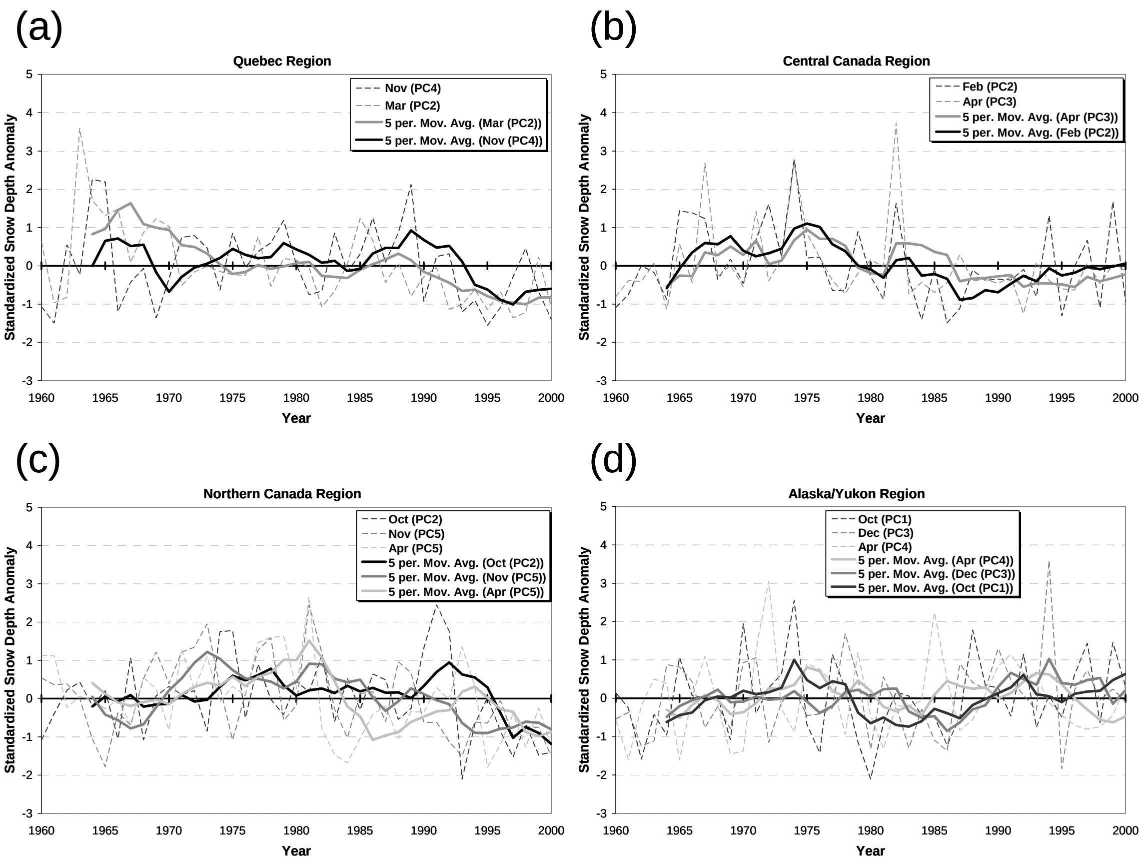


Figure 2.7. Score time series of rotated PCs from selected regions within North America, including (a) Quebec, (b) central Canada, (c) northern Canada, and (d) Alaska/Yukon regions.

Table 2.1. Timing and values of maximum snow cover extent (SCE) at select snow depth threshold levels.

Snow Depth (cm)	Max Mean (10x6 km^2)	Timing of Max Mean (pentad)	Max Change (10x6 km^2)	Timing of Max Change (pentad)	Max Change (% of Mean)
2	15679.4	3	-60.6	25	-0.39
10	13568.8	5	-54.7	25	-0.40
20	11446.4	10	-58.6	23	-0.51
30	9162.9	12	-65.0	19	-0.71
40	6858.4	13	-67.5	19	-0.98
50	4828.5	13	-62.0	19	-1.28
60	3200.8	13	-54.1	15	-1.69
70	2039.4	14	-41.9	15	-2.05
80	1273.5	14	-31.0	15	-2.44
90	788.1	14	-23.6	15	-2.99
100	500.2	14	-17.1	15	-3.42

Table 2.2. Percent explained variance for PCs. These values correspond to those PC loadings retained for Varimax rotation and subsequent analysis.

	Percent Explained Variance						
	Oct	Nov	Dec	Jan	Feb	Mar	Apr
PC1	24.9	18.4	12.3	11.9	14.6	16.3	18.5
PC2	13.1	11.3	10.5	10.5	11.5	10.2	9.5
PC3	7.3	7.8	8.6	9.7	10.5	7.5	8.3
PC4	6.1	7.3	6.8	7.1	6.4	6.3	6.6
PC5	5.3	6.2	5.9	5.3	-	5.5	5.2
PC6	5.2	5.0	-	-	-	-	-

Table 2.3. Snow depth regression coefficients for defined regions based on rotated PC loadings during 1960-2000.

	Oct	Nov	Dec	Jan	Feb	Mar	Apr
Quebec	-0.36	-0.55	-	-	-1.61	-1.92	-
Alaska/Yukon	0.50	0.47	0.63	0.98	-	-	0.18
N. Canada	-0.50	-1.12	-0.48	-	-	-	-1.25
Great Plains	-	0.95	0.07	-	-0.23	-0.22	-
Central Canada	-	0.26	-0.19	-0.36	-0.54	-1.12	-0.51
Ohio Valley	-	-	-	-0.19	-0.23	0.12	-
West Coast	-	-	-	-0.05	-	-	-
Great Lakes	-	-	-	-	-	-0.17	-0.54
Central Canada	-	-0.23	-0.04	-	-	-	-1.44

CHAPTER 3

TRENDS IN SNOW ABLATION OVER NORTH AMERICA[2]

[2] Dyer, J.L. and T.L. Mote. To be submitted to *International Journal of Climatology*.

ABSTRACT

A substantial decrease in snow cover extent (SCE) and snow depth over North America has been observed over the last 40 years. One possible explanation for the changes in North American snow cover is a change in the frequency and/or intensity of snow ablation. This study uses a gridded dataset of US and Canadian surface observations from 1960-2000 to examine patterns of snow ablation over North America.. Results show a positive trend in the frequency of ablation events during March and a significant negative trend in May, indicating an earlier onset of ablation. This pattern is consistent for ablation of varying intensity. Surface energy budget components and air mass frequencies are examined in relation to the observed trends in snow ablation. Changes in March ablation frequency are shown to be dominated by increases in the sensible heat flux. A higher frequency of dry moderate instead of moist polar air masses during high ablation years may explain the increase in sensible heat flux and ablation over the study period.

3.1 Introduction

High-latitude environments within North America have been shown to have high seasonal and interannual variability in snow cover during the winter season (Frei and Robinson, 1999; Frei et al., 1999; Hughes and Robinson, 1996; Groisman et al., 1994). Brown (2000) demonstrated that over North America, there were statistically significant changes in snow cover extent (SCE) for the months of November and April. Frei et al. (1999) also showed a general pattern of increased SCE in the fall and decreased SCE in the spring. Chapter 2 documented decreasing snow depths across much of North America in March and April, particularly in the interior of Canada. The earlier end to the snow season is a central conclusion of the current and previous research because of the importance of the timing and magnitude of snowmelt runoff to regional hydrologic systems.

One possible explanation for the decrease in the spring snow depth and snow extent is lower snow accumulation and subsequent snow depths during the winter. However, Hughes et al. (1996) found that there was an increase in 20th Century SCE over the mid-latitudes (below 55°N), especially during the winter months, which was also consistent with observed evidence of increasing snowfall over the northern US and southern Canada (Hughes and Robinson, 1996; Groisman and Easterling, 1994; Mekis and Hogg, 1999).

Results from Chapter 2 showed snow depth to have a decreasing trend throughout central and northern Canada. The increase in SCE in the mid-latitudes and decrease in snow depth in central Canada indicate a shallowing of the snowpacks in the winter, which could lead to a more rapid decrease in snow cover in the spring. These findings suggest that the decrease in spring snow cover may be due to an increase in the frequency and/or intensity of snow ablation (decreases in snow mass through melt and/or sublimation). A change in snowmelt frequency or intensity would also suggest a change in the meteorological conditions related to snowmelt. Brown (2000) compared the changes in SCE over the winter season with changes in monthly temperature trends (Jones, 1994) and found higher temperatures in April and lower temperatures in November.

This research examines the patterns and trends in ablation frequency and intensity over North America from 1960-2000 to ascertain if ablation is occurring more frequently. In addition, areas shown to have significant ($p < 0.05$) trends in ablation frequency are examined in more detail to determine the mechanisms responsible for changing patterns of snow ablation. This is done through analysis of surface energy budget components and air mass characteristics. A change in the timing and rapidity of snowmelt can have a considerable effect on hydrology due to more frequent or rapid snowmelt runoff. Therefore, knowledge of the patterns and mechanisms related to ablation over North America is vital in understanding the patterns and trends of snow extent and depth.

3.2. Data and Methodology

a. Snow Depth Data

Data used for this project come from stations within the United States cooperative observer network (U.S. Department of Commerce, 2003) and Meteorological Service of Canada (Braaten, 1996; Brown, pers. comm..), and include daily snow depth, maximum and minimum temperature, snowfall, and precipitation observations. Records are available at select U.S. and Canadian sites since the late 1800s; however, a vast majority of the U.S. stations have periods of record beginning in 1948 subsequent to the nationwide modernization of the cooperative observer network. For the Canadian stations, the Cryospheric System to Monitor Global Change in Canada (CRYSYS) program (Goodison and Brown, 1997), initiated in 1995, added approximately 400 stations with varying lengths of record between 1900-1990 to the existing Digital Archive of Canadian Climate Data, allowing for the creation of a separate Canadian snow depth database (Brown and Braaten, 1998).

All daily snow depth observations were subjected to a quality control routine following Robinson (1989) to omit unreasonable values and test the internal consistency of the US and Canadian snow cover data. The quality control process involved comparing a time series of daily snow depth measurements for a given station with associated daily snow fall, maximum and minimum temperature, and precipitation data from the same station for the same day. The quality control check was based upon the existence of snow depth data before and after the current day. Snow depth observations that did not meet the quality control criteria were flagged as inconsistent and not included in the analysis.

The most apparent limitation of daily snow cover or meteorological observations at the continental scale is that station locations frequently follow population patterns; therefore, observation networks are often biased to lower elevations, and in the case of the Canadian network, to lower latitudes. The snow depth data were gridded at a resolution of 1°x1° to make the North American snow depth and climatological data more spatially consistent. The spatial extent of the gridded area was chosen to include the U.S. and the major land area of Canada (south of 71°N), with areas south of latitude 20°N omitted because of the lack of substantial snow cover at any time during the year. The grids were calculated using an inverse-distance algorithm that interpolated the grids on a two-dimensional Cartesian plane before projecting them to a spherical surface (Willmott et al., 1984).

b. Defining Snow Ablation

An ablation event is defined as a decrease in snow depth between two successive daily snow depth observations as a result of snowmelt. However, it is possible for snow depth to decrease due to other processes besides melt, including snow compaction and sublimation. Sublimation can often be substantial over large scales. For example, over northeastern Canada sublimation removes an average 29 mm yr^{-1} of water from the snowpack, which is equivalent to 7% of the local annual precipitation (Déry and Yau, 2002). However, sublimation is difficult to distinguish from melt using daily surface observations; therefore, no measures are taken to remove bias in ablation frequency and/or intensity due to sublimation.

Compaction is a process that can substantially decrease snow depth without directly altering the mass of the snowpack, and is related to small-scale ice crystal interactions. When snow first reaches the surface, snow crystals rapidly metamorphose into more rounded forms through mechanical stress (wind) or thermodynamic processes related to vapor pressure differences along the periphery of ice crystals (Colbeck, 1983). As a result, the snowpack compresses through settling of individual snow particles, leading to a denser and shallower snowpack. With additional snow accumulation, the rate of compaction increases due to overburden forces, which also act to further metamorphose ice particles into shapes that are more efficiently packed (Colbeck, 1973). Gunn (1965) found the greatest compaction rates occur immediately after snow has fallen, with density increasing at an average rate of 1% per hour. This rate may increase with strong winds or more intense snowfall (Mellor, 1977).

The criteria defining an ablation event for this project is adjusted to include only those instances where two-day snow depth decreases and the maximum daily temperature on the second day of the associated ablation event is above 0°C. This is done to remove the effect of compaction as effectively as possible. It is assumed that the snowpack is relatively isothermal and mature, minimizing the rate of compaction included in the recorded snow depth change. Many of the defined ablation events are associated with snowfall that must be incorporated into calculation. This is done by adding the depth of snowfall on the second day of a given ablation event to the calculated change in snow depth.

Once an ablation event is defined, it is standardized by the area of the grid cell in which it was calculated. This is necessary because of the change in area of 1°x1° grid cells with latitude, and the bias encountered in the number of ablation events at high latitudes where grid cells are smaller compared to lower latitudes. Ablation "events" are defined as an interdiurnal snow depth change exceeding a critical value over an area of 10,000 km^2.

c. Surface Energy Balance and Air Mass Data

The surface energy balance is shown to play an important role in snow ablation in mid- and high latitude prairie snowpacks (e.g., Dyer and Mote, 2002; Grundstein and Leathers, 1999; Kuusisto, 1986). At below freezing temperatures, the net shortwave radiation is found to be the dominant source of energy available for melt, while at above freezing temperatures the sensible and latent heat fluxes dominate. Energy balance components related to snow ablation, given by the equation below, govern the frequency and intensity of ablation events over a given region and time of year.

$$\Delta E = SW_{NET} + LW_{NET} + H + LE + G + F + \mathrm{Pr} \qquad \textbf{(3.1)}$$

where ΔE is the change in energy available at the surface, SW_{net} and LW_{net} are net short and longwave radiation, respectively, H is the sensible heat flux, LE is the latent heat flux, G is the ground heat flux, F is the advective heat flux, and Pr is the energy available from liquid precipitation. Monthly mean values (calculated from six-hour observations at 00, 06, 12, and 18 UTC) for each of the components included in the equation except for Pr are obtained from the National Center for Environmental Prediction (NCEP) reanalysis data, with 700 hPa temperature and 1000-500 hPa heights used to estimate the advective heat flux (NOAA-CIRES, 2005; Kalnay et al., 1996). The NCEP upper-air data are gridded at a 2.5°spatial resolution and the surface fluxes at a nominal 1.875° resolution Data from cells overlying the regions included in the analyses were used and resampled to the same 1° x1° grids used for the snow depth data. The number of events in which a minimum of 1 cm of liquid precipitation was added to the snowpack comprised a minimal percentage of the total number of ablation events (< 1%). Therefore, despite the efficiency of rain in melting snow, Pr was not included in analysis of the surface energy budget and the influences on snow ablation. Future studies should include an analysis of the influence of rain on snow.

Surface energy flux components, especially the sensible and latent heat fluxes, are related to variations in temperature and humidity resulting from influences of air masses and air mass modification, as well as differences in wind speed near the surface (Leathers et al., 2004). Data from the Spatial Synoptic Classification 2 (SSC2) climatology were used to identify air mass type and frequency over selected regions for analysis (Sheridan, 2002). The SSC2 utilizes discriminate analysis on hourly standard meteorological observations to identify which of seven distinct air masses exist over a particular location for a given day. Air mass type is identified using existing meteorological conditions over an area and not the location from which the air mass originated. The same air mass type can have different characteristics at different locations and at different times during the year. The air mass types included in the SSC2 are: dry polar (DP), dry moderate (DM), dry tropical (DT), moist polar (MP), moist moderate (MM), and moist tropical (MT). There is also a transition (Tr) category that represents a day in which an air mass transition occurred, such as during a frontal passage. For this study, air mass types determined from observations at La Ronge, Saskatchewan, Canada (55°N latitude, 105°W longitude), were used because this station is closest to the region used in the analyses (approximately 150km distant). The large spatial extent of air masses and relatively slow modification through central Canada permit the use of air mass information from this nearby station.

3.3. Patterns and Trends in Snow Ablation

A distinct seasonal pattern of snow ablation frequency over North America during 1960-2000 is evident, characterized by a maximum frequency of events during April (total of 16,756 events; Figure 3.1). This peak in the frequency of ablation drops rapidly through June (approximately 6000 events $month^{-1}$), before rising from September through mid-winter (approximately 1600 events $month^{-1}$). The increase in the number of ablation events from October through March is attributable to snowmelt along the periphery of the snowpack as it advances and retreats in the lower latitudes of the U.S. During April and May, rising seasonal temperatures provide abundant energy with which to initiate ablation and melt the snowpack in the higher latitudes of Canada and Alaska.

April does not have a significant trend in the frequency of ablation despite having the greatest frequency of events. March has the largest positive trend in ablation frequency (56 events yr^{-1}; $p < 0.05$), and a significant negative trend was evident in May (-65 events yr^{-1}; p <

0.05; Figure 3.1). Positive trends in ablation in March and negative trends in ablation in May and June indicate a substantial change in the pattern of ablation events over North America during 1960-2000. With the existence of such a pattern, it is clear that the frequency of ablation events is shifting so that the melt season occurs earlier in the year. This trend in ablation frequency matches well with studies of North American snow cover extent (Frei and Robinson, 1999; Frei et al., 1999; Brown, 2000) and snow depth (Chapter 2), leading to the conclusion that changes in the frequency of ablation may have a considerable influence on patterns of continental snow depth and snow cover extent.

In addition to a change in the frequency of ablation over North America from 1960-2000, a change in the intensity of ablation may also play a large role in explaining trends in continental snow extent and depth. To analyze the trends in intensity of ablation over the study region, four threshold levels were defined based on 10 cm changes in snow depth. The first level of ablation is referred to hereafter as minor ablation, defined as having a one-day change in snow depth between 0-10 cm, followed by moderate, major, and extreme ablation, with one-day changes in snow depth between 10-20 cm, 20-30 cm, and 30-40 cm, respectively.

Analysis of the frequency of ablation at the defined intensity threshold levels indicates a relatively greater fraction of minor ablation events in fall and a relatively smaller fraction in spring (Figure 3.2a). The standardized monthly means (frequency of events standardized by long-term mean) suggest that the frequencies of events at higher ablation intensities remain relatively constant. The number of ablation events within successively higher threshold levels decreases substantially, due to fewer days with snow depth exceeding that threshold (Table 3.1). Figure 3.2a also shows a maximum in the frequency of ablation events in April at all intensities (2.4 to 3.5 units), with a minimum in the summer and a steady increase from October through

March. This pattern illustrates the consistency inherent in the frequency of ablation at all intensity levels, meaning that both frequency and intensity are related and vary together.

Above the minor ablation threshold level, the patterns in the trends of ablation show a uniform progression, with positive trends occurring in March (5.5 to 7.0 units) and negative trends occurring in April and May (-3.5 to -1.8 units; Figure 3.2b). However, the trends in minor snow ablation for each month show a maximum in ablation frequency during March (16.8 units), with a secondary peak in November (12.3 units). These minor events in Figure 3.2b match Figure 3.1, which is expected because roughly 95% of all defined ablation events have a change in snow depth of less than 10 cm. The trends in ablation (March maximum) at all intensity threshold levels indicates that both the number and intensity of ablation events is changing uniformly over the 1960-2000 study period, with a uniform shift from May and June towards March.

3.4. Snow Ablation Processes

One regions in North America shown to have unique patterns of ablation and snow depth variability during 1960-2000 were chosen for detailed analysis. This regions was selected during March, corresponding to the month with the greatest positive trends in the number of ablation events (Figure 3.1). The area was defined using the location of the first principal component (PC) from principal components analysis of monthly snow depths across North America (see Chapter 2) (Figure 3.3). The region chosen for further analysis in March extends from central Saskatchewan through central Alberta and into the eastern margin of British Columbia. Snow depth in this region is shown to have the large negative trends from 1960-2000 (Chapter 2).

The component score time series and ablation frequency for March is shown in Figure 3.4, along with the associated snow depth time series for reference. The frequency of ablation rose significantly ($p < 0.01$), while component scores (representative of a standardized, weighted snow depth for this region) decreased significantly from 1960 through the early 1990s (Figure 3.4). It is expected that an increase in the frequency of ablation will lead to a decrease in snow depth, which is what is shown to occur ($R^2 = 0.54$ between component score and ablation frequency).

Within the region described above, five-year high and low ablation periods were defined based on the time series of ablation frequency. The high ablation period is defined as 1990-1995, while the low ablation period is defined as 1965-1970 (Figure 3.5). Using these five-year periods, it is possible to compare the climate during years with high and low ablation frequencies. Additionally, by using five-year periods near the beginning and end of the study period, forcing mechanisms responsible for the positive trend in ablation over the study period can be inferred.

The surface energy balance components related to snow ablation in March give a mean energy input into the surface (sum of latent, sensible, ground, and radiative fluxes) of 10.6 W m^{-2} per 6-hour time period during the high ablation time period (1990-1995), while during the low ablation period (1965-1970) only 5.6 W m^{-2} of energy is added to the surface (Table 3.2). The difference in energy input of 5 W m^{-2} is dominated by a substantial difference in the sensible heat flux, where the mean during the high ablation period is 16.3 W m^{-2} higher than in the low ablation period ($t = -2.87$; $p < 0.05$). Since all other surface energy and radiative balance components show no significant difference in means between the high and low ablation periods, it suggests that the sensible heat flux is the dominant process related to the increased ablation through the 1960-2000 study period. The energy available for the increased sensible heat flux is likely attributable to an increase in advected sensible heat, which is shown by a 2.8°C increase in 700 hPa temperatures during the high ablation period relative to the low ablation period ($t = 1.620$; $p < 0.15$; Table 3.2).

The change in 700 hPa temperatures also is associated with a change in air mass type and frequency between high and low ablation periods. Between the high and low ablation periods there is a higher frequency of DM air masses (320 and 188, respectively) and a lower frequency of MP air masses (328 and 409, respectively; Figure 3.6). Additionally, over the study period there is a significant positive trend in days with DM air masses during March over the region (0.22 days yr^{-1}; $p < 0.05$). DM air masses are normally associated with zonal flow aloft, leading to adiabatic drying and warming of air as it traverses the Rocky Mountains, while MP air masses are cool and humid due to advected moisture from the North Pacific. The higher frequency of the warmer DM air masses (mean 1500 LST temperature of 6°C), would lead to increased ablation due to more energy available for melt (higher temperatures) and enhanced evaporation and sublimation (less moisture) (Table 3.3). It is possible that an increase in precipitation and/or wind speeds would further increase ablation by increasing the number of rain-on-snow events. However, such a change might logically be expected to be associated with transitional (Tr) days. Our analysis shows that there is no significant difference in the number of Tr days between the low and high ablation periods, despite a significant negative trend of -0.12 days yr^{-1} ($p < 0.05$) of Tr days during March from 1960-2000.

Analysis of ablation and snow depth over North America shows that trends in snow depth are more sensitive to changes in the frequency of ablation during the spring melt season than during the fall accumulation season. This is true despite the fact that the frequency of ablation is increasing at both times over the 1960-2000 period of record. Additionally, the significant positive trend in the frequency of ablation during March is a result of a higher flux of sensible heat into the surface, caused by an increase in the frequency of warmer air masses over the region and higher temperatures in the mid troposphere.

3.5. Conclusions

It has been documented that winter snow cover over North America is undergoing a change such that fall accumulation and spring ablation are occurring earlier (Chapter 2; Brown, 2000; Frei et al., 1999). There are several possible reasons for the trend in snow cover over North America. This chapter examines the relationship between snow depth and ablation over North America in order to ascertain if snow depth trends can be explained by changes in the frequency and/or intensity of snow ablation. Additionally, surface energy budget components as well as air mass type are analyzed in order to ascertain which meteorological processes are most important with respect to snow depth trends over a region North America with dramatically thinning spring snow cover.

The frequency of snow ablation over North America is shown to be highest during April, with a rapid decline in the number of events during May and June. The associated trends in the frequency of ablation events show the highest positive trend to occur in March, with a significant negative trend in May ($p < 0.05$). This indicates a shift in the frequency of ablation events over North America toward earlier in the season, which supports findings of a spring decrease in continental snow cover. The same frequency and trend patterns extend to ablation intensity, in that ablation at all intensities maximizes in April, but has a maximum positive and negative trend in March and May, respectively.

Analysis of ablation and snow depth patterns shows a close relationship to exist during March, when snow depth has a significant negative trend and ablation frequency has a significant positive trend ($p < 0.01$). An examination of energy budget components and air mass type and frequency for high and low ablation periods shows that the sensible heat flux is the dominant mechanism responsible for explaining the increasing trend in March snow ablation. The increase in sensible heat flux and subsequent increase in ablation frequency is most likely caused by an increase in the frequency of dry moderate (DM) air masses over central Canada and a decrease in moist Polar (MP) air masses, leading to a change in available energy at the surface.

The results of this project aid in determining when and where the frequency and intensity of snow ablation is changing over North America from 1960-2000, and help in determining the causes for the trends in ablation frequency through central Canada in March when the number of events is increasing most rapidly. Additionally, these results provide important information regarding the mechanisms responsible for the patterns and variability of snow depth over North America from 1960-2000, and help to define regions most susceptible to changes in ablation properties. This information can be used in future research focused on relationships between snow ablation, snow cover, and other climatological and hydrological systems related to surface processes.

REFERENCES

Braaten, R. O., 1996: an updated Canadian snow depth database. Contract report prepared for Atmospheric Environment Service, September, 1996, 24 pp.

Brown, R.D. and R.O. Braaten, 1998. Spatial and temporal variability of Canadian monthly snow depths, 1946-1995. *Atmos.-Ocean*, **36**, 37-45.

Brown, R. D., 2000: Northern Hemisphere snow cover variability and change, 1915-97. *J. Climate*, **13**, 2339-2355.

Brown, R.D., 2003: Canadian snow depth data. Personal communication.

Colbeck, S.C., 1973: Theory of metamorphism of wet snow. USA Cold Regions Research Engineering Laboratory, Research Report 313.

Colbeck, S.C., 1983: Theory of metamorphism of dry snow. *J. Geophys. Res.*, **88**, 5475-5482.

Déry, Stephen J. and M.K. Yau, 2002: Large-scale mass balance effects of blowing snow and surface sublimation. *J. Geophys. Res.*, **107**, 4679, doi: 10.1029/2001JD001251.

Dyer, J.L. and T.L. Mote, 2002: Role of energy budget components on snow ablation from a mid-latitude prairie snowpack. *Polar Geog.*, **26**, 87-115.

Frei, A. and D. A. Robinson, 1999: Northern Hemisphere snow extent: regional variability 1972-1994. *Int. J. Climatol.*, **19**, 1535-1560.

Frei, A., D. A. Robinson, and M. G. Hughes, 1999: North American snow extent: 1900-1994. *Int. J. Climatol.*, **19**, 1517-1534.

Goodison, B. E. and R. D. Brown, 1997: CRYSYS – Use of the Cryospheric System to Monitor Global Change in Canada. *In*: Proc. GER'97 (Int. Symp.: Geomatics in the Era of Radarsat), 24-30 May 1997, Ottawa, Ontario. Canadian Geomatics CD-ROM Vol. 1, Natural Resources Canada, Paper 259, 13 pp.

Groisman, P. Ya. and D. Easterling, 1994. Variability and trends of total precipitation and snowfall over the United States and Canada. *J. Climate*, **7**, 184-205.

Groisman, P. Y., T. R. Karl, and R. W. Knight, 1994: Changes of snow cover, temperature, and radiative heat balance over the Northern Hemisphere. *J. Climate*, 7, 1633-1656.

Grundstein, A. J. and D. J. Leathers, 1999: A spatial analysis of snow-surface energy exchanges over the northern Great Plains of the United States in relation to synoptic scale forcing mechanisms. *Int. J. Climatol.*, **19**, 489-511.

Gunn, K.L.S., 1965: Measurements on new-fall snow. McGill University, Montreal, Stormy Weather Group Scientific Report MW-44.

Hughes, M.G. and D.A. Robinson, 1996: Historical snow cover variability in the Great Plains region of the USA: 1910 through to 1993. *Int. J. Climatol.*, **16**, 1005-1018.

Hughes, M.G., A. Frei., and D.A. Robinson, 1996. Historical analysis of North American snow cover extent: merging satellite and station derived snow cover observations. *Proc. of the 1996 Eastern Snow Conference*, Williamsburg, VA, 21-32.

Jones, P.D., 1994. Hemispheric surface air temperature variations: A reanalysis and an update to 1993. *J. Climate*, **7**, 1794-1802.

Kalnay, E., M. Kanamitsu, R. Kistler, W. Collins, D. Deaven, L. Gandin, M. Iredell, S. Saha, G. White, J. Woollen, Y. Zhu, A. Leetmaa, and B.Reynolds 1996: The NCEP/NCAR 40-year reanalysis project, *Bull. Amer. Meteor. Soc*., 77, 437–471.

Kuusisto, E., 1986: The energy balance of a melting snow cover in different environments. *Proc. Of the Budapest Symp*., July 1986. IAHS Publ. No. 155, 37-45.

Leathers, D.J., Leathers, D.J., D. Graybeal, T.L. Mote, A.J. Grundstein and D.A. Robinson, 2004: The role of air mass types and surface energy fluxes in snow cover ablation in the central Appalachians, *J. Appl. Meteor.*, **43**, 1887-1898.

Mekis, É. and W.D. Hogg, 1999: Rehabilitation and analysis of Canadian daily precipitation time series. *Atmos.-Ocean*, **37**, 53-85.

Mellor, M., 1977: Engineering properties of snow. *J. Glaciol.*, **19**, 15-66.

NOAA-CIRES Climate Diagnostics Center, 2005: NCEP Reanalysis Data, Boulder, CO, [Available online http://www.cdc.noaa.gov/].

Robinson, D. A., 1989: Construction of a United States historical snow data base. *Proc. of the 45th Eastern Snow Conference*, 50-59.

Sheridan, S.C., 2002: The redevelopment of a weather-type classification scheme for North America. *Int. J. Climatol*., 22, 51-68.

Sheridan, S.C., 2005: Spatial Synoptic Classification homepage, Department of Geography, Kent State University. [Available online: http://sheridan.geog.kent.edu/ssc.html].

U.S. Department of Commerce, 2003: TD-3200 Surface Summary of the Day, National Climatic Data Center, NOAA. [Available online: http://www.ncdc.noaa.gov].

Willmott, C.J., C.M. Rowe, and W.D. Philpot. 1984: Spheremap. Center for Climatic Research, Department of Geography, University of Delaware, Newark DE.

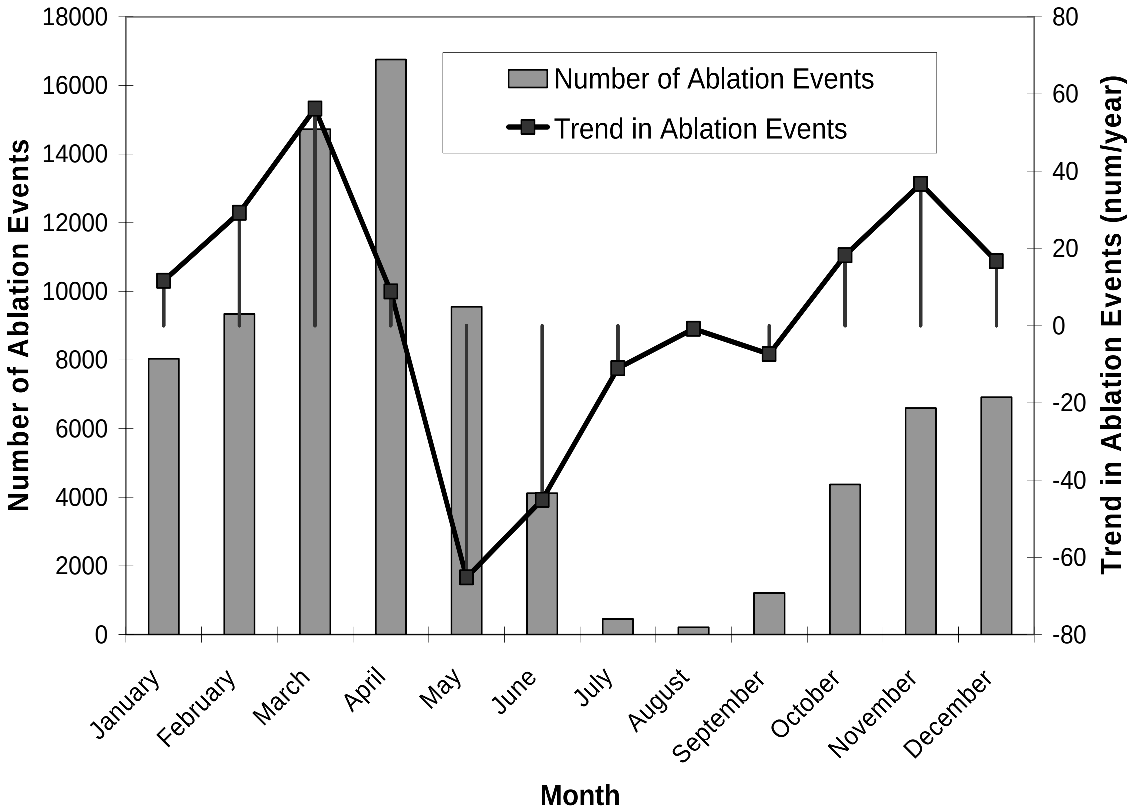
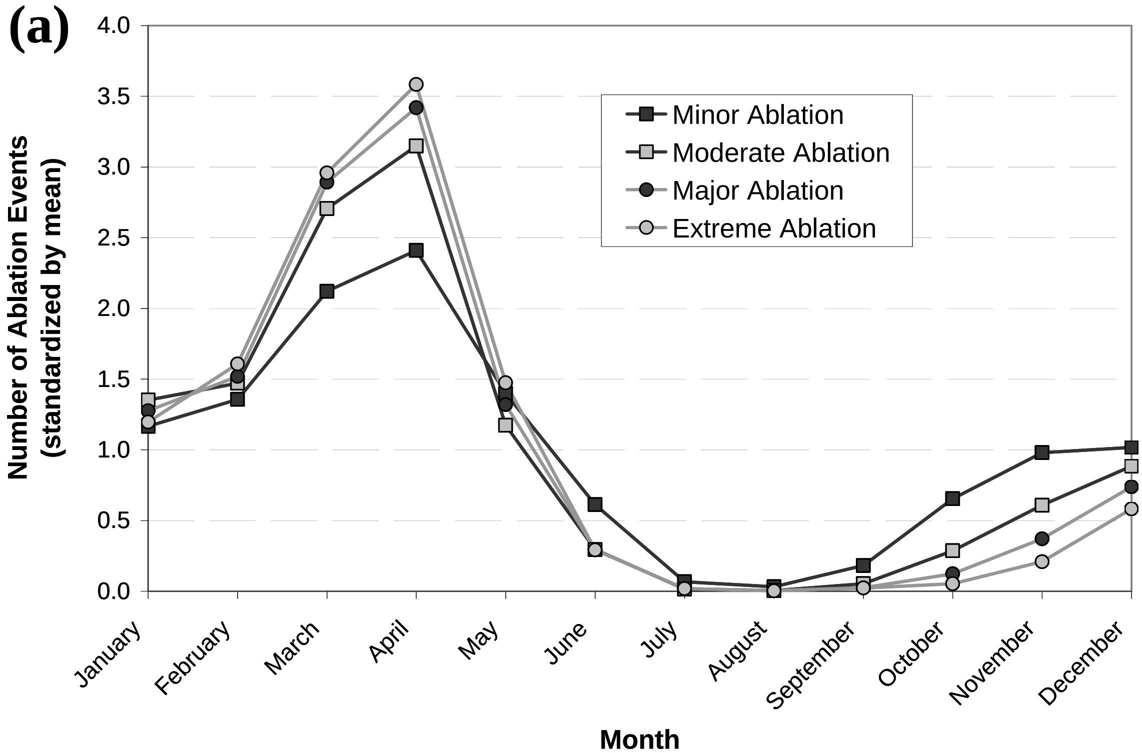


Figure 3.1. Monthly means and trends in number of ablation events over North America from 1960-2000. Vertical drop lines indicate the difference in trend from 0.



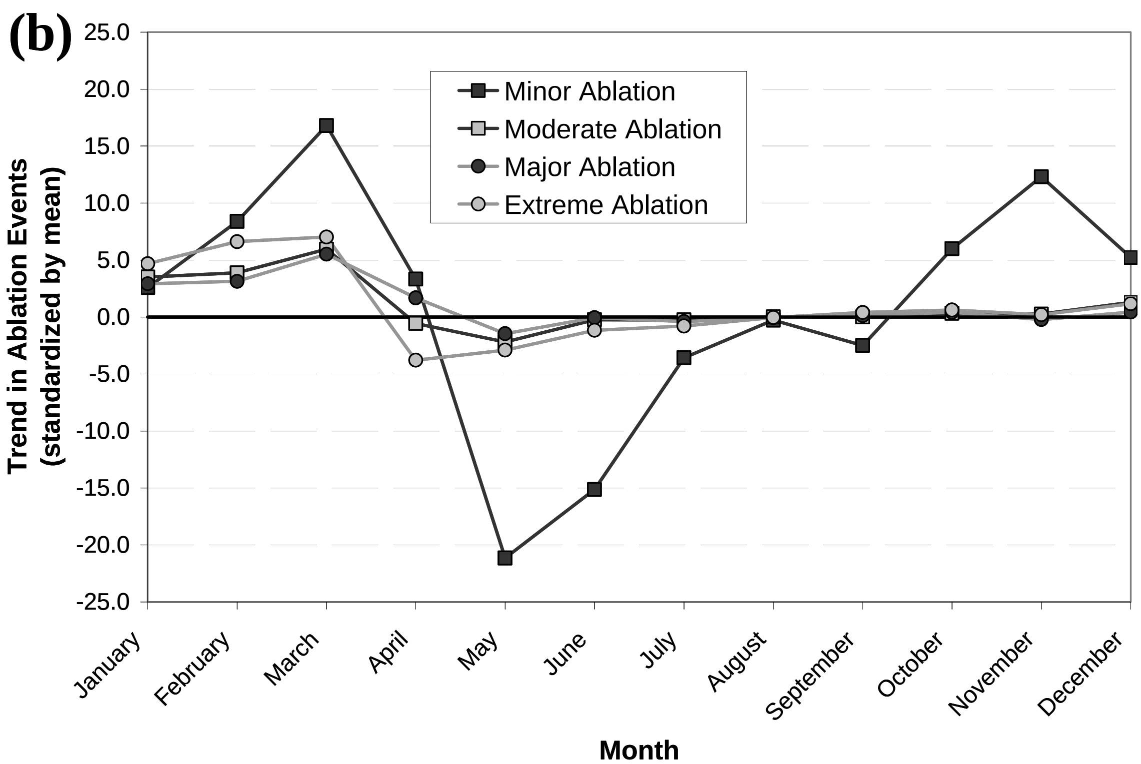


Figure 3.2. Standardized (a) monthly mean number of ablation events at defined threshold levels and (b) monthly trend in number of ablation events at defined threshold levels.

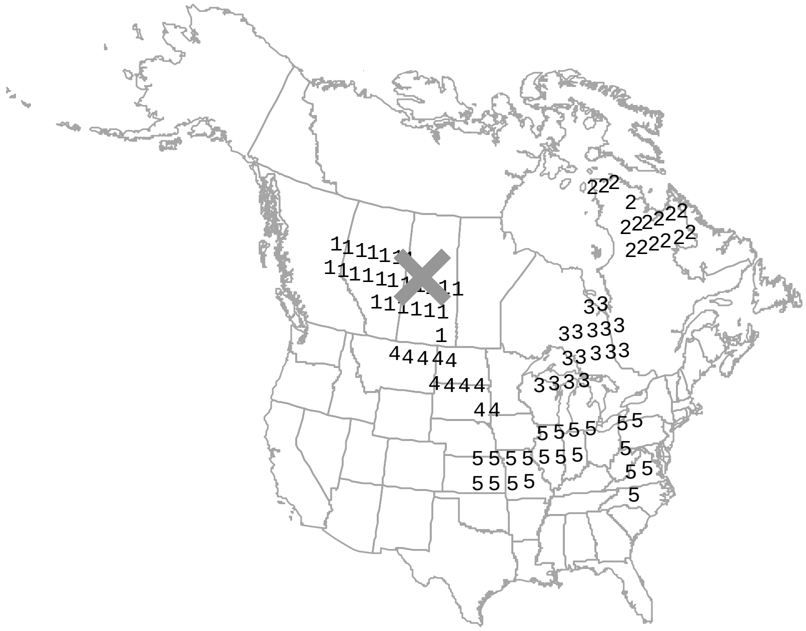


Figure 3.3. Location of March PC region from PCA of North American snow depth (from Chapter 2). “X” marks the position of the highest loading within the region defined by the first rotated PC.

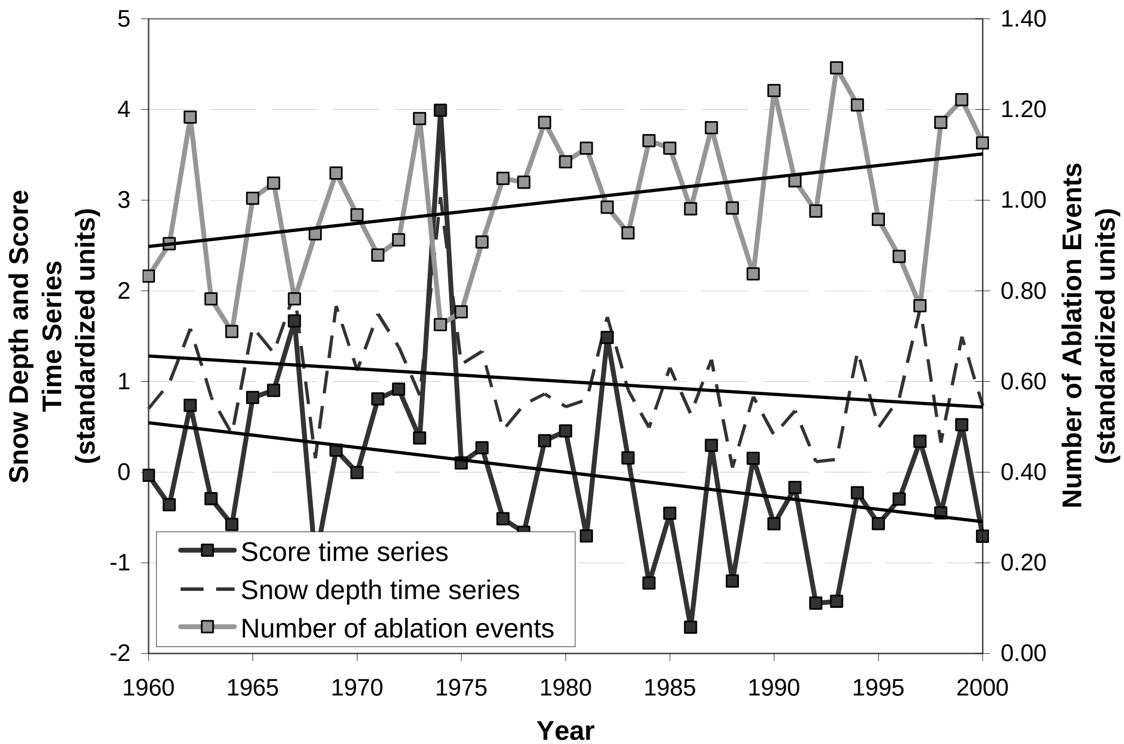


Figure 3.4. Snow depth and score time series and number of ablation events (standardized by mean) for March. Linear trend lines are included to give indications of the direction and strength of trends.

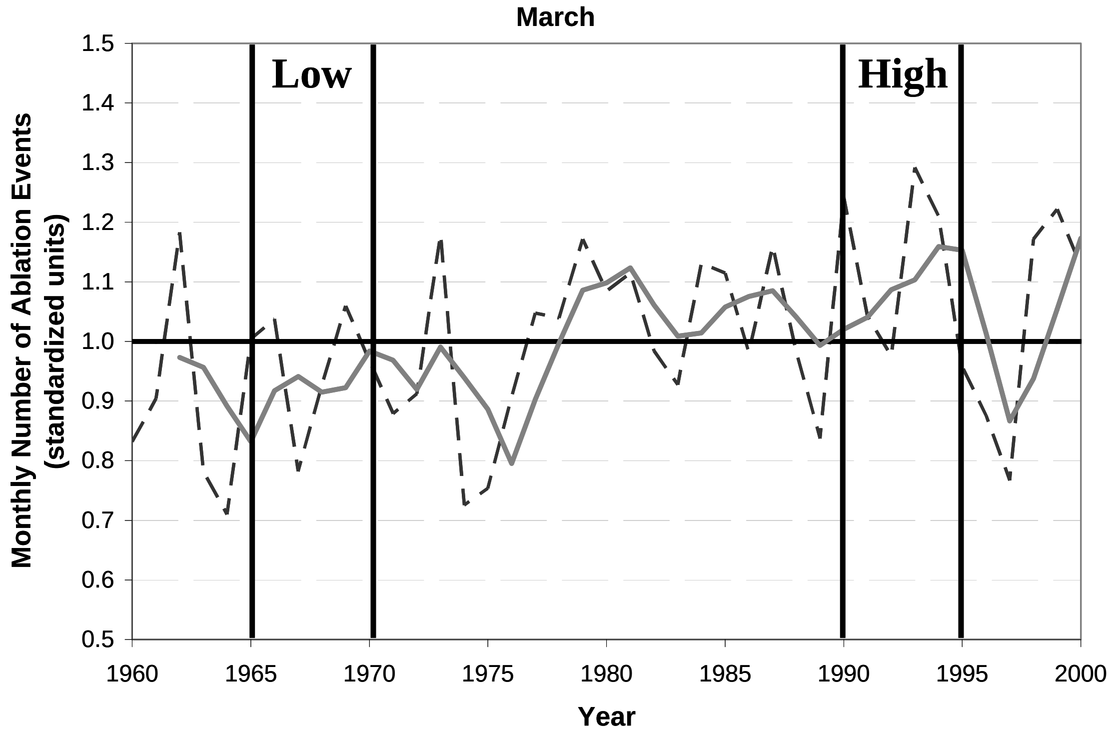


Figure 3.5. Standardized number of ablation events for March. Dotted lines denote the standardized monthly mean number of ablation events, while the solid lines are the three-year moving average of the mean number of events. High and low five-year periods are outlined by solid vertical lines.

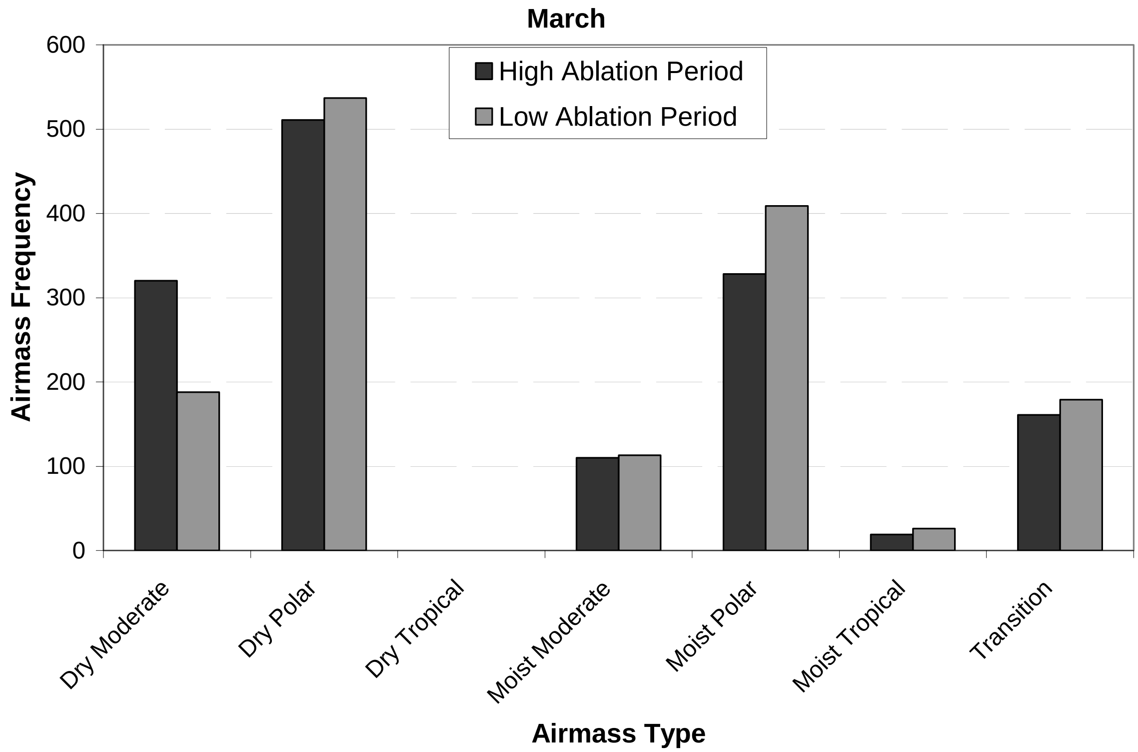


Figure 3.6. Frequency of SSC2 airmass type for high and low ablation periods in March.

Table 3.1. Monthly mean, standard deviation, trend, and trend significance for all months showing significant ($p < 0.01$) trends in the number of ablation events at the defined threshold levels.

Month	Mean	Std Dev	Trend	Significance
All Ablation Events				
February	9344.2	1361.0	29.2	0.90
March	14722.9	2040.1	56.2	0.97
May	9551.3	1733.6	-65.2	1.00
June	4116.9	1035.1	-45.2	1.00
November	6591.6	1021.4	36.7	1.00
Minor Ablation Events (0 cm < Change in Snow Depth < 10 cm)				
March	13979.2	1981.1	49.8	0.94
May	9219.3	1684.1	-62.7	1.00
June	4039.6	1005.3	-44.8	1.00
July	446.7	446.9	-10.6	0.93
November	6458.2	1012.2	36.5	1.00
Moderate Ablation Events (10 cm < Change in Snow Depth < 20 cm)				
January	238.3	55.5	2.5	1.00
February	259.2	78.9	2.8	0.99
March	477.0	107.2	4.2	1.00
May	206.9	59.8	-1.6	0.95
Major Ablation Events (20 cm < Change in Snow Depth < 30 cm)				
January	60.2	17.7	0.7	1.00
February	71.7	21.8	0.7	0.99
March	136.6	33.0	1.3	1.00
Extreme Ablation Events (30 cm < Change in Snow Depth < 40 cm)				
January	24.3	9.3	0.3	0.98
February	32.7	12.1	0.4	0.99
March	60.1	16.8	0.4	0.94

Table 3.2. Monthly mean values for the surface energy and radiative fluxes for both the high and low ablation periods defined for March. Additional information includes the difference in means, and the results of a paired t-test describing the sign significance of the difference in means. All heat flux and radiation values are in W m^{-2}, temperature is in °C, and heights are in meters. Negative fluxes indicate energy directed into the surface, positive fluxes indicate energy directed away from the surface.

	Latent Heat Flux	*Sensible Heat Flux*	*Ground Heat Flux*	*Net Longwave*	*Net Shortwave*	*700 hPa Temperature*	*1000-500 hPa Height*
High Ablation	13.9	-25.1	0.9	62.3	-62.6	-12.4	2625
Low Ablation	11.7	-8.8	2.4	58.0	-68.8	-15.2	2615
Difference in means	2.2	-16.3	-1.5	4.4	6.3	2.8	10.3
Significance	0.73	0.97	0.65	0.87	0.50	0.83	0.36

Table 3.3. Description of monthly mean frequency, temperature, dew point, and cloud cover of SSC2 air mass types during March at La Ronge, Saskatchewan, Canada. Data are from Sheridan (2005).

	Frequency of Days (%)	Temperature (°C)		Dew Point (°C)		Cloud Cover (tenths)
		0300 LST	1500 LST	0300 LST	1500 LST	
Dry Moderate	16.8	-4	6	-8	-5	6
Dry Polar	45.4	-17	-8	-21	-16	5
Dry Tropical	0.0	0	0	0	0	0
Moist Moderate	5.5	-2	2	-4	-2	9
Moist Polar	19.7	-8	-5	-11	-9	9
Moist Tropical	0.8	-1	9	-3	1	6
Transition	11.7	-16	-5	-20	-12	8

CHAPTER 4

SNOW DEPTH AND STREAMFLOW RELATIONSHIPS IN LARGE NORTH AMERICAN WATERSHEDS[3]

[3] Dyer, J.L. To be submitted to *Water Resources Research*.

ABSTRACT

Snowmelt runoff in the spring is an important component in regional hydrologic systems in the northern US and Canada, having a vital influence on water resources. In northern latitude rivers, snowmelt runoff provides a considerable volume of freshwater to drive circulation in the Arctic Ocean. This project defines and analyzes patterns of snow volume and discharge in major North American watersheds, and determines the strength and form of the associated relationships. The results are used to develop statistical models applicable to each individual watershed. The watersheds included in the analysis are the Yukon and Mackenzie basin in northern Canada and Alaska, the Saskatchewan basin in southern Canada, and the Missouri and upper Mississippi watersheds in the northern US. It is shown that snow volume can predict winter and early spring discharge in all watersheds in the study region, with the best model performance in the higher latitude Yukon and Mackenzie basins during late fall and winter accumulation. In the lower latitude Missouri and upper Mississippi basins, despite additional influences of rain on discharge patterns, the statistical models based on snow volume were still able to estimate streamflow with percent relative error around 50%. To improve modeled discharge estimates during peak spring runoff, additional snow cover variables, including the value and timing of peak snow volume and the duration of snowmelt, were compared to peak annual discharge during periods of intense snowmelt. Significant results were found to occur in the Yukon and Saskatchewan basins due to the extreme sensitivity to snowmelt runoff and the fast river response time, respectively.

4.1. Introduction

Snowmelt is the dominant hydrologic event during the year for many regions of North America, providing valuable water resources but also increasing the potential for spring flooding (e.g., Kane, 1997; Leathers et al., 1998; Dyer, 2001). In addition, the large flux of freshwater that is released in the spring through snowmelt runoff is important in maintaining freshwater volumes and driving circulation in the Arctic Ocean (Aagaard and Carmack, 1989; Miller and Russell, 2000; Peterson et al., 2002). Due to this relationship between snow cover and hydrology, changes in the patterns of snow cover may accentuate variability in the hydrologic regime within the mid- and high-latitudes of North America (Miller and Russell, 2000; Serreze et al., 2000).

Previous research has shown that winter snow storage and subsequent melt are strongly related to the timing and magnitude of discharge in high-latitude rivers (Rango, 1997; Zhang et al., 2001; Cao et al., 2002; Yang et al., 2003). This relationship extends to the mid-latitudes, although the strength of the association weakens with decreasing latitude due to a more ephemeral snow cover. Nevertheless, Roads et al. (1994) showed that river streamflow values in the northern regions of the U.S. were dependent on winter snow accumulation, but added that the relative contribution of spring snowmelt to the total large-scale hydrologic budget was not well known.

Studies have examined snow cover and hydrology relationships in the U.S. to determine the importance of snowmelt in regional hydrologic systems (Groisman et al., 2001), but few studies have been published on snow and hydrology relationships over major watersheds in North America. Yang et al. (2003) conducted research involving large-scale snow cover and streamflow associations in Eurasia using satellite snow extent maps over three major watersheds.

Although it was found that snow cover extent could predict discharge patterns with significant accuracy, their findings were limited due to the two-dimensional nature of the snow extent data.

The overall objective of this study to find the strength and form of the relationship between variations in snow volume (area-averaged snow depth) and streamflow for five major watersheds in North America (Yukon, Mackenzie, Saskatchewan, Missouri, and upper Mississippi River basins). Specifically, statistical models are developed for each watershed based on the relationships between mean and peak annual snow volume and streamflow. The use of snow volume in the analyses rather than snow cover extent will improve the predictive capability of the statistical models. This will lead to improved knowledge of continental-scale hydrologic processes, and will aid in determining how long-term variations in snow cover effect discharge in large, high-latitude watersheds.

4.2. Data and Methodology

a. Study Area

To examine the relationship between snow volume and streamflow over North America, suitable large-scale watersheds were delineated based primarily on climate, topography, and drainage area. Climate was taken into account by not including areas within North America that have inconsistent winter snow cover, which excludes approximately the southern half of the US and some higher latitude coastal regions where the average annual number of days with snow cover (>= 2.5 cm) is minimal (< 50 days yr^{-1}; Figure 4.1). In addition, watersheds in regions characterized by complex terrain were not included in the study area unless they were part of a larger inland drainage network. The varying slope and aspect and sharp changes in elevation in these regions lead to relatively rapid and complex patterns of runoff that make it difficult to accurately predict long-term seasonal streamflow. Finally, watersheds in close proximity to the ocean, which normally have small spatial extent and tidal effects associated with streamflow at the basin outlet, were omitted from the analysis.

The North American study area is comprised of five major drainage basins: the Yukon, Mackenzie, Saskatchewan, Missouri, and upper Mississippi River basins. These basins encompass a large part of the US Great Plains and Canadian prairies, as well as the northern tundra and boreal regions of Canada and Alaska. Although a large portion of the North American land area was not included in the study area, it has been shown that the interior of North America accounts for much of the variability in continental-scale variations in snow cover (Frei and Robinson, 1995; Frei and Robinson, 1999; Brown and Goodison, 1996). Defining the relationship between streamflow and snow volume in these areas will provide information regarding the influence of snowmelt runoff on discharge over the North American continent.

The basins listed previously were created using watershed outline data from the United States Geological Survey (USGS; U.S. Department of the Interior, 1994) hydrologic unit codes (HUC; Seaber et al., 1987) and Natural Resources Canada (2003) drainage boundaries. Major basins were delineated using the USGS HUC hydrologic region boundaries and the Water Resources Canada primary watershed subdivision identifiers. Watersheds that crossed the U.S. – Canada border were merged to obtain continuous watershed outlines. These basins range in size from roughly 458,000 km^2 for the upper Mississippi River basin to around 1,699,000 km^2 for the Mackenzie River basin (Figure 4.2).

b. Streamflow and Snow Depth Data

Streamflow data for the U.S. and Canada were obtained from the USGS (U.S. Department of the Interior, 2003) and Environment Canada's (2003) National Water Data Archive, respectively. For each major drainage basin, a single streamflow measurement station was chosen that matched the following criteria: (1) the observed time series must include at least 25 years of daily data, (2) the observed streamflow cannot be influenced by tidal or hydraulic effects, such as backwater flow (3) the station must be located along the main stem of the river at a point where the river does not branch into more than one channel. Although it is known that streamflow measurements at the outlets of rivers are an imperfect measure of the total surface runoff due to unmeasured groundwater flow across watershed boundaries and regulatory and diversion structures and activities within the basin, using large-scale watersheds minimizes these effects (Zektser and Loaiciga, 1993; Waggoner and Schefter, 1990). The locations of the stream gauging stations are shown in Figure 4.2.

The snow depth data used for this project comes from the National Weather Service cooperative observer network (U.S. data; U.S. Department of Commerce, 2003) and Meteorological Service of Canada (Canadian data; Braaten, 1996; Brown, pers. comm..) over the period 1960-2000. These data were quality controlled using a routine proposed by Robinson (1989), and daily 1° latitude x 1° longitude grids of each variable were created using an algorithm that interpolated the grids to a spherical surface before projecting them onto a two-dimensional Cartesian plane (Willmott et al., 1984). Although snow depth data are available over much of North America prior to 1960, the 1960-2000 study period was chosen due to the poor spatial distribution of surface observation stations in northern Canada prior to 1960. Using the mean daily snow depth from each grid cell, the associated daily snow volume was calculated by multiplying the depth by the surface area of the grid cell. For each drainage basin in the study area, all 1°x1° grid cells that had at least 50% of their area in a watershed boundary were assigned to that watershed, and basin-total snow volume was obtained by summing all daily snow volume values for each individual watershed. The average number of grid cells included in each basin varied from 51 for the upper Mississippi watershed to 281 for the Mackenzie watershed. The Saskatchewan, Missouri, and Yukon basins have 62, 144, and 149 grid cells, respectively.

Once the daily basin-total snow volume was calculated for all watersheds, five-day means (i.e. pentads) were produced for each individual basin over the 1960-2000 study period. This was done to reduce high frequency temporal variability inherent in the data. The pentads were produced such that the first pentad would always start on January 1, with the 12^{th} pentad (February 25 – March 1) including February 29 on leap years. This results in 73 pentads for each year. The same process was carried out for the streamflow data, although the time series for the Yukon and Mackenzie basins only extended to 1975-2000 and 1972-2000, respectively. For all other basins, the streamflow time series included data from 1960-2000.

The relationship between changes in snow volume and streamflow in each of the major watersheds is shown by comparing annual hydrographs developed using average pentad discharge from each stream gauging station over the respective periods of record with the associated basin-average snow volume. Because the patterns of snow accumulation and ablation within each watershed share distinct characteristics with respect to magnitude and rate of change of snow volume, which have differing influences on streamflow, the accumulation and ablation phase is analyzed separately. The form and strength of the relationship for each basin during accumulation and ablation are determined by regression of pentad snow volume against pentad streamflow. It is expected that the functional forms of the regression equations in each basin for accumulation and ablation will differ according to basin characteristics, runoff lag, and the influence of natural and man-made control structures. Weather will also be a control of variations from pentad to pentad.

It is hypothesized that using snow volume and by analyzing the accumulation and ablation phases independently, the predictive ability of the models will be improved. The most difficult parameter to estimate using these models is peak discharge, which is an important variable due to the close association with flood potential. As a result, in addition to looking at the strength and form of the relationship between mean pentad snow volume and streamflow during winter accumulation and ablation, annual peak discharge is analyzed using several variables related to snow cover. This is done using a multiple regression analysis comparing peak spring discharge with (a) annual maximum snow volume, (b) pentad when the maximum snow volume occurred, and (c) the length of the ablation period.

4.3. Snow Cover and Streamflow Characteristics

All of the basins selected have a distinct snow accumulation and ablation phase (Figure 4.3). For each basin, the timing of the autumn increase and spring decrease of snow cover, as well as the magnitude of the maximum snow extent, is different. The Yukon basin is the first to show a substantial increase (> 5% basin area) in SCE (pentad 55; Figure 4.3a; Table 4.1), and the Mississippi basin is the last (pentad 64), which corresponds with the first occurrence of snowfall and freezing maximum daily temperatures in the respective regions (Figure 4.4a-b). The Yukon basin has the highest average SCE, remaining near 100% from pentad 65 to pentad 21 (Figure 4.3a; Table 4.1). Peak snow volumes, however, are even more variable among the basins (Figure

4.3b). The Mackenzie basin has the highest mean basin snow volume, peaking at 790×10^9 m^3 on pentad 13, which is double the maximum snow volume of 389×10^9 m^3 in the Yukon basin that occurred on pentad 11. The lower latitude Saskatchewan, Missouri, and Mississippi basins show considerably lower snow volumes than the higher latitude Mackenzie and Yukon basins.

The timing of melt in the individual basins is influenced by the prolonged existence of high elevation snowpacks and unique basin characteristics (areal extent, soil type, terrain, vegetation), which can slow or delay the rate of snowmelt. As a result, there are three distinct periods during which the five large North American watersheds become snow free (< 5% snow covered; Figure 4.3a), despite the varying basin characteristics.

The river systems in each of the basins included in this study are subject to increased streamflow resulting from spring runoff. This is in the noticeable rise in streamflow during late winter and early spring (Figure 4.5). There is, however, a substantial difference between the river response to snowmelt in the Yukon and Mackenzie basins relative to the lower latitude basins, primarily due to the considerable difference in mean snow volume associated with each watershed (Figure 4.3). The response of river discharge to spring snowmelt is less clear along the lower latitude Saskatchewan, Missouri, and upper Mississippi Rivers, where basin average snow volume is lower (Figure 4.3b) and the depth of liquid mean precipitation is higher (Figure 4.4c); however, the release of snowmelt runoff in the spring does have a noticeable influence on river discharge. For the Saskatchewan and Missouri River basins there is a substantial rise in river discharge corresponding to a decrease in snow volume in the late winter and early spring.

4.4. Snow Cover and Streamflow Comparisons

The seasonal patterns of snow volume and streamflow within the five watersheds included in this study are shown in Figure 4.6. In the Yukon and Mackenzie basins, the influence of snowmelt on seasonal streamflow is most clearly defined, with the annual minimum discharge occurring just after snow volume has reached a peak, and peak discharge occurring as snow volume approaches zero (Figure 4.6a-b). In the Missouri and upper Mississippi Rivers, the addition of considerable depths of liquid precipitation throughout the year (Figure 4.4c) influences streamflow patterns during the summer, fall, and early winter (Figure 4.6c-e). The Saskatchewan River, because of its higher-latitude location and higher annual snowfall (Figure 4.4b), does not show the same variations; however, there are some patterns of discharge that do not correspond with variations in the associated snow volume. Most noticeable is a distinct secondary peak in streamflow during the summer that corresponds with a period of high mean precipitation (Figure 4.4c).

To determine the strength of the relationship between streamflow and snow cover within the basins included in this study, separate regression analyses were performed using mean pentad discharge and snow volume during the accumulation and ablation phases of the snow season (Figure 4.7a-e). The accumulation phase begins during the first pentad of snow volume, and continues to the pentad of maximum snow volume. The ablation phase extends from the pentad of maximum snow volume to the last pentad with recorded snow volume. Regression equations describing the relationship between snow volume and streamflow in all basins during both the accumulation and ablation phases are shown in Table 4.2.

Results of the statistical analysis show strong relationships to exist between snow volume and discharge during the accumulation phase in the Yukon and Mackenzie watersheds (R^2 = 0.97

and 0.92, respectively; $p < 0.05$; Figure 7a-b). In the ablation phase, power functions give the best fit for regression (R^2 = 0.92 and 0.89, respectively; $p < 0.05$) due to the rapid increase in discharge as snow volume decreases through the spring (Figure 4.6a-b). The same pattern of rapid discharge increase after the initiation of spring ablation exists in the Saskatchewan basin as in the higher latitude basins, although the accumulation phases are different as a result of smaller basin size and more rapid river response in the Saskatchewan watershed (Figure 4.6c). Nevertheless, the relationships between snow volume and streamflow are significant for both accumulation and ablation phases (R^2 = 0.95 and 0.93, respectively; $p < 0.05$; Figure 4.7c).

In the Missouri and upper Mississippi watersheds, seasonal changes in snow volume and discharge lead to a linear form of relationship during both the accumulation and ablation phases (Figure 4.7d-e). Despite the abundance of control structures within the Missouri basin and the lack of substantial snow volume in the upper Mississippi basin, snow volume and streamflow are highly related. This is especially true in the ablation phase in the upper Mississippi basin, which shows an R^2 of 0.98 ($p < 0.05$). During the accumulation phase in the upper Mississippi watershed, the slope of the linear regression line is low, indicating that river discharge remains fairly constant throughout the winter despite a steady increase in snow volume (Figure 4.6e). This is a result of continually deepening snowpacks in the northern reaches of the basin with rain occurring in the southern portions of the basin where seasonal temperatures remain at or near 0°C through the winter months. Rain leads to immediate runoff that keeps river levels consistent despite the storage of water in the basin snowpacks.

a. *Modeled Streamflow*

The predictive capability of the statistical models developed using basin-average snow volume and discharge is described through simulation of historical streamflow for specific years during the 1960-2000 period of record. For each basin, the years with the highest and lowest streamflow, as well as the year with streamflow closest to the long-term mean, are used to test the individual statistical models. Using years of mean and extreme discharge will test both the predictive range of the models, as well as their utility during years with average streamflow and snow volume conditions. Model fit is tested by comparing simulated and observed discharge within each of the defined years and calculating the percent relative error (PRE) at each pentad when the model was run.

The higher latitude Yukon and Mackenzie basins show low PRE during the accumulation phase, except for the medium streamflow year in the Yukon basin (Table 4.3). During this year (1983-1984), PRE decreases consistently during the accumulation season as snow volume increases (Figure 4.8a). The exponential function used to simulate discharge during the accumulation phase in the Yukon basin (Figure 4.7a) is least accurate at low snow volume levels, where patterns of discharge are most variable. During the high discharge year in the Mackenzie basin (1991-1992), a unique pattern existed in which the snow season actually began in early September (pentad 50) and extended into June (pentad 34), allowing for a long simulation period (Figure 4.8b).

Modeled streamflow in the Saskatchewan basin is considerably more accurate in the ablation phase than during the accumulation phase, which is a result of extrapolation beyond the polynomial regression curve for the accumulation phase during years with highs now volume (Table 4.3; Figure 4.7c). Despite the increased error in model predictions early in the snow season, simulated discharge is still able to follow observed streamflow patterns relatively well during the mean discharge year (1966-1967; Figure 4.8c). In the lower latitude Missouri and upper Mississippi basins, PRE between predicted and observed discharge during medium and high discharge years is moderately low (mean PRE of 43%), despite the importance of rain in the regional hydrologic systems (Table 4.3). This is best illustrated in the Missouri basin during the high discharge year (1992-1993), where the pattern of observed streamflow is well matched by the predicted streamflow (Figure 4.8d; mean PRE of 36%). However, the influence of rain on observed discharge is most apparent in the spring, seen by the peaks in streamflow during the ablation phase not indicated in the patterns of predicted discharge. Results show that even in mid-latitude watersheds, it is possible to model winter and spring discharge using snow volume, which helps in estimating regional water resources.

The predictive capability of the basin-specific statistical models developed through regression analysis of snow volume and discharge is good, especially in the accumulation phases in the higher latitude watersheds. The primary drawback with the statistical models is the inability to estimate peak spring river discharge, since snowpacks are often ablated by the time peak streamflow occurs and the statistical models, which are based on snow volume, are no longer applicable. However, since spring peak discharge is an important aspect of hydrologic systems due to the association with flooding potential, it is vital to address this issue using additional snow cover data.

b. Analysis of Peak Discharge

The greatest change in streamflow normally occurs during the spring subsequent to the release of spring snowmelt runoff. It is also at this time that the hydrologic systems are most sensitive to liquid precipitation since surface river channels, subsurface aquifers, and soil layers are usually saturated due to snowmelt runoff. In order to minimize the residuals of the regressions during peak spring discharge, it is necessary to understand how peak annual streamflow is related to additional snow cover variables related to spring ablation, such as peak snow volume, timing of occurrence of peak snow volume, and duration of spring snow melt. Clearly, rain and basin surface and subsurface characteristics play a large role in determining the magnitude of annual peak discharge; however, the estimation of spring discharge maximums using winter snow cover information is vital in understanding long-term hydrologic patterns.

Snow cover parameters used in this study to compare with peak spring discharge include the timing and value of maximum snow volume, as well as the duration of snow melt. The timing of maximum snow volume is defined as the pentad during which snow volume is greatest prior to the associated spring ablation season, with the maximum snow volume represented by the value associated with that pentad. Snow melt duration is defined as the duration (in pentads) between the defined pentad of maximum snow volume and the first pentad during which snow volume is less than 5% of the associated maximum. A threshold of 5% is used so that lingering, patchy snowpacks are not considered. Using each of these variables, a multiple regression analysis was performed comparing each of the snow cover variables to the peak annual discharge.

For the Yukon basin, results show that maximum snow volume explains 46% of the variance in the peak annual streamflow, with the timing of snow volume and snowmelt duration accounting for 9% and 10%, respectively. Overall, 56% of the variance of peak annual discharge is described by the defined snow cover variables (Table 4.4). For the Mackenzie basin, snow melt duration explained 21% of the variability, with all snow cover variables having a combined

R^2 of 0.29. The reason for the weaker relationship between snow cover variables and peak annual discharge is due to the complex nature of the hydrologic system within the basin. The Mackenzie basin is characterized as having several large lakes and tributaries, which act to slow the propagation of the spring snowmelt runoff that leads to the peak streamflow at the mouth of the watershed. The Saskatchewan basin, which has a simpler hydrologic structure, has 42% of the variance in peak annual discharge explained by maximum snow volume (Table 4.4). Each of the multiple regression analyses described above are shown to be significant ($p < 0.01$); therefore, peak annual discharge can be partially explained by incorporating the three defined snow cover variables.

4.5. Conclusions

This project uses basin-averaged snow volume, developed using gridded North American snow depth data, along with daily mean streamflow from select US and Canadian stations, to define and quantify the relationship between snow volume and discharge in major North American watersheds. Results indicate that snow volume plays a major role in hydrologic patterns within North America, and is able to predict winter and early spring discharge with considerable accuracy. This is especially true in the high latitude Yukon and Mackenzie watersheds in northern Canada and Alaska, where snow volume and streamflow show strong relationships during both the accumulation (R^2 = 0.97 and 0.92, respectively) and ablation (R^2 = 0.92 and 0.89, respectively) phases of the winter snow season.

In the lower latitude Saskatchewan basin, rapid river response times lead to higher errors in predicted streamflow during the accumulation phase of the snow season, despite having a strong relationship between snow volume and streamflow (R^2 = 0.95). This is due to the inability for the statistical model to accurately extrapolate beyond the second-power polynomial curve used for regression analysis; however, when snow volume is within the range of the model, discharge estimates improve considerably. For the Missouri and Mississippi basins, despite an increased sensitivity to rain, the statistical models using snow volume to predict discharge work fairly well during years with medium to high snow volume (mean PRE of 42%).

In order to define the influence of snow cover on peak annual discharge, a number of snow cover variables were included in a multiple regression analysis that described both the magnitude and timing of the peak snow volume and subsequent melt period. This was done because the peak annual discharge was found to be related not only to the volume of snowmelt runoff being introduced into the hydrologic system, but also the rate at which it was introduced. For the Yukon basin, 56% of the variance on peak annual discharge was explained by derived snow cover parameters, while only 29% was explained in the neighboring Mackenzie basin. The difference in explained variance is a result of basin characteristics such as vegetation and topography, which has a substantial influence on streamflow in the Mackenzie basin because of the large areal extent. For the Saskatchewan basin, which has a much smaller areal extent, 43% of the variance was explained by snow cover patterns.

This project is an important step in identifying and quantifying the relationship that exists between snow volume and hydrology within mid- and high-latitude North American watersheds. Because snowmelt runoff supplies a substantial volume of water to many regions within North America, affecting water resources and freshwater influx into the Arctic Ocean, it is important to define these associations for the purposes of seasonal hydrologic forecasting as well as long-term climate research. Using the statistical models developed in this study, it is possible to reconstruct streamflow records where there are gaps in the data record or too short of a time series is available to perform adequate analysis. This will allow for the development of longer streamflow time series that will aid in long-term hydrologic studies, which is especially important in high latitude rivers due to importance of freshwater discharge on Arctic Ocean circulation, and the resulting influence on high latitude climate patterns.

REFERENCES

Aagaard, K. and E.C. Carmack, 1989: The role of sea ice and other fresh water in the Arctic circulation. *J. Geophys. Res.*, **94**(C10), 14,485–14,498.

Braaten, R. O., 1996: an updated Canadian snow depth database. Contract report prepared for Atmospheric Environment Service, September, 1996, 24 pp.

Brown, R.D. and B.E. Goodison, 1996. Interannual variability in reconstructed Canadian snow cover, 1915-1992. *J. Climate*, **9**, 1299-1318.

Brown, R.D., 2003: Canadian snow depth data. Personal communication.

Cao, Z., M. Wang, B. Proctor, G. Strong, R. Stewart, H. Ritchie, and J. Burnford, 2002: On the physical processes associated with the water budget and discharge of the Mackenzie basin during the 1994/1995 water year. *Atmos.-Ocean*, **40(2)**, 125-143.

Dyer, J.D., 2001: *Simulating the 1997 Red River floods utilizing a coupled snowpack and hydrologic model*. M.S. thesis submitted to the Graduate School, The University of Georgia, Athens, Georgia, USA.

Environment Canada, 2003: HYDEX/HYDAT database, National Water Data Archive, Water Survey Branch.

Frei, A. and D. A. Robinson, 1995: Regional signals in Northern Hemisphere snow cover during autumn and spring. In *Proc. Fourth AMS Conf. on Polar Meteorology and Oceanography*, Dallas, Texas, 127-131.

Frei, A. and D. A. Robinson, 1999: Northern Hemisphere snow extent: regional variability 1972-1994. *Int. J. Climatol.*, **19**, 1535-1560.

Groisman, P. Y., R.W. Knight, and T.R. Karl, 2001: Heavy precipitation and high streamflow in the contiguous United States: Trends in the twentieth century. *Bull. Amer. Meteor. Soc.*, **82**, 219-246.

Kane, D. L., 1997: the impact of Arctic hydrologic perturbations on Arctic ecosystems induced by climate change, in *Global Change and Arctic Terrestrial Ecosystems*, *Ecol. Stud.*, vol. 124, pp. 63-81, Springer-Verlag, New York, 1997.

Leathers, D. J., D. R. Kluck, and S. Kroczynski, 1998: The severe flooding event of January 1996 across north-central Pennsylvania. *Bull. Amer. Meteor. Soc.*, **79** (5), 785-797.

Miller, J. R. and G. L. Russell, 2000: Projected impact of climate change on the freshwater and salt budgets of the Arctic Ocean by a global climate model, *Geophys. Res. Lett.*, **27** (8), 1183–1186.

Natural Resources Canada, 2003: Geomatics Canada, Earth Sciences Sector, Canada Center for Remote Sensing, GeoAccess Division [Available online: http://geogratis.cgdi.gc.ca/clf/en].

Peterson, B.J., R.M. Holmes, J.W. McClelland, C.J. Vorosmarty, R.B. Lammers, A.I. Shiklomanov, I.A. Shiklomanov, S. Rahmstorf, 2002: Increasing river discharge to the Arctic Ocean. *Science*, **298**, 2171-2173.

Rango, A., 1997: Response of areal snow cover to climate change in a snowmelt-runoff model. *Ann. Glaciol.*, **25**, 232-236.

Roads, J.O., S.-C. Chen, A.K. Guetter, and K.P. Georgakakos, 1994: Large-scale aspects of the United States hydrologic cycle. *Bull. Amer. Meteor. Soc.*, **75**, 1589-1610.

Robinson, D. A., 1989: Construction of a United States historical snow data base. *Proc. of the 45th Eastern Snow Conf.*, 50-59.

Seaber, P.R., F.P. Kapinos, and G.L. Knapp, 1987: Hydrologic Unit Maps. U.S. Geological Survey Water-Supply Paper 2294, 63 p.

Serreze, M. C., J. E. Walsh, E. C. Chapin, T. Osterkamp, M. Dyugerov, V. Romanovsky, W. C. Oechel, J. Morison, T. Zhang, and R. G. Barry, 2000: Observation evidence of recent change in the northern high-latitude environment, *Climatic Change*, **46**, 159-207.

U.S. Department of Commerce, 2003: TD-3200 Surface Summary of the Day, National Climatic Data Center, NOAA. [Available online: http://www.ncdc.noaa.gov].

U.S. Department of the Interior, 1994: Hydrologic units maps of the Conterminous United States, U.S. Geological Survey, [Available online: http://water.usgs.gov/GIS/metadata/usgswrd/huc250.html].

U.S. Department of the Interior, 2003: NWISWeb water data. U.S. Geological Survey. [Available online: http://waterdata.usgs.gov/nwis].

Waggoner, P. E. and J. Schefter, 1990: Future water use in the present climate. *Climate Change and United States Water Resources,* P. E. Waggoner, Ed. John Wily and Sons, 19-39.

Willmott, C.J., C.M. Rowe, and W.D. Philpot. 1984: Spheremap. Center for Climatic Research, Department of Geography, University of Delaware, Newark DE.

Yang, D., D. Robinson, Y. Zhao, T. Estilow, and B. Ye, 2003: Streamflow response to seasonal snow cover extent changes in large Siberian watersheds. *J. Geophys. Res.*, **108**, 1-14.

Zektser, I. S. and H. A. Loaiciga, 1993: Groundwater fluxes in the global hydrologic cycle: Past, present and future. *J. Hydrol.*, **144**, 405-427.

Zhang, X., K. D. Harvey, W. D. Hogg, and T. R. Yuzyk, 2001: Trends in Canadian streamflow, *Water Resour. Res.*, **37**, 987-998.

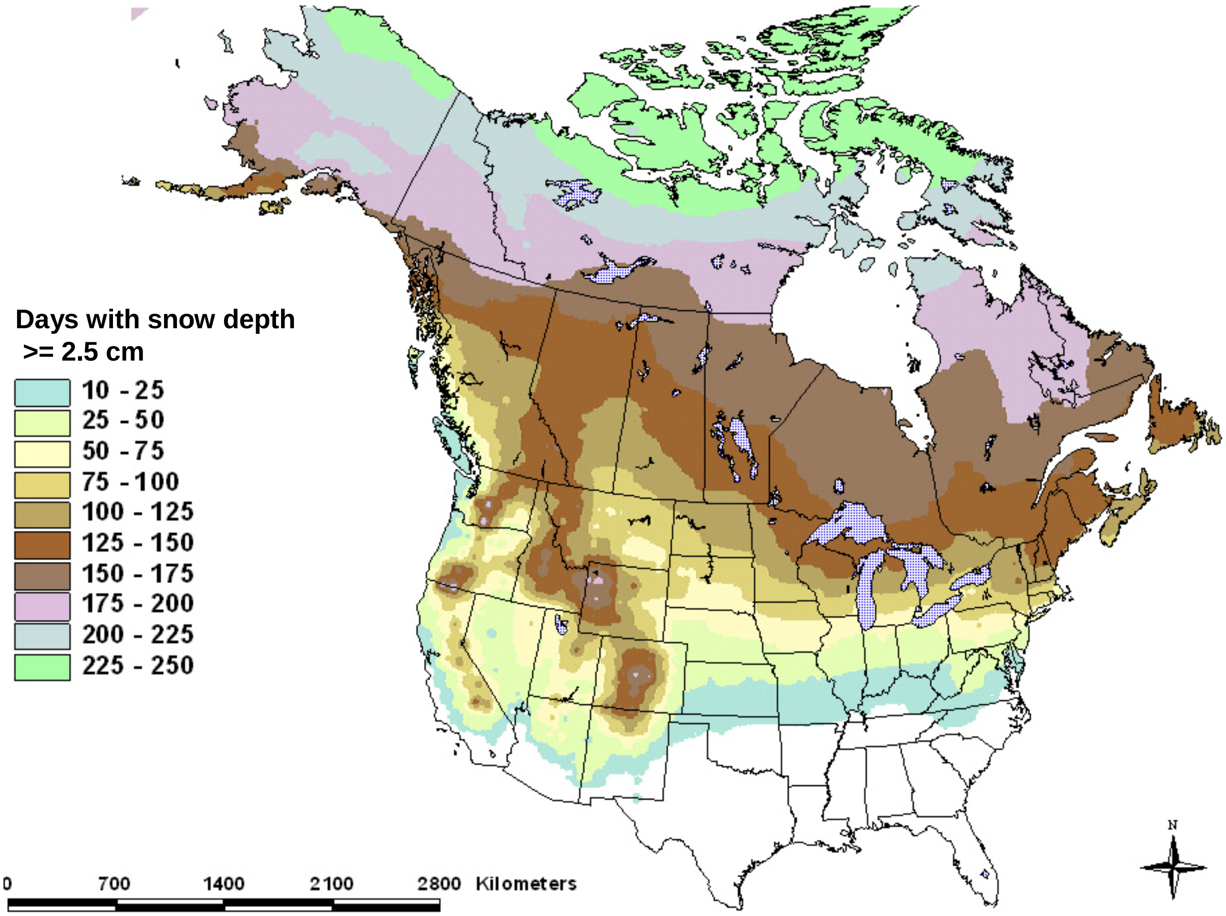


Figure 4.1. Number of days with snow depth greater than 2.5 cm over North America.

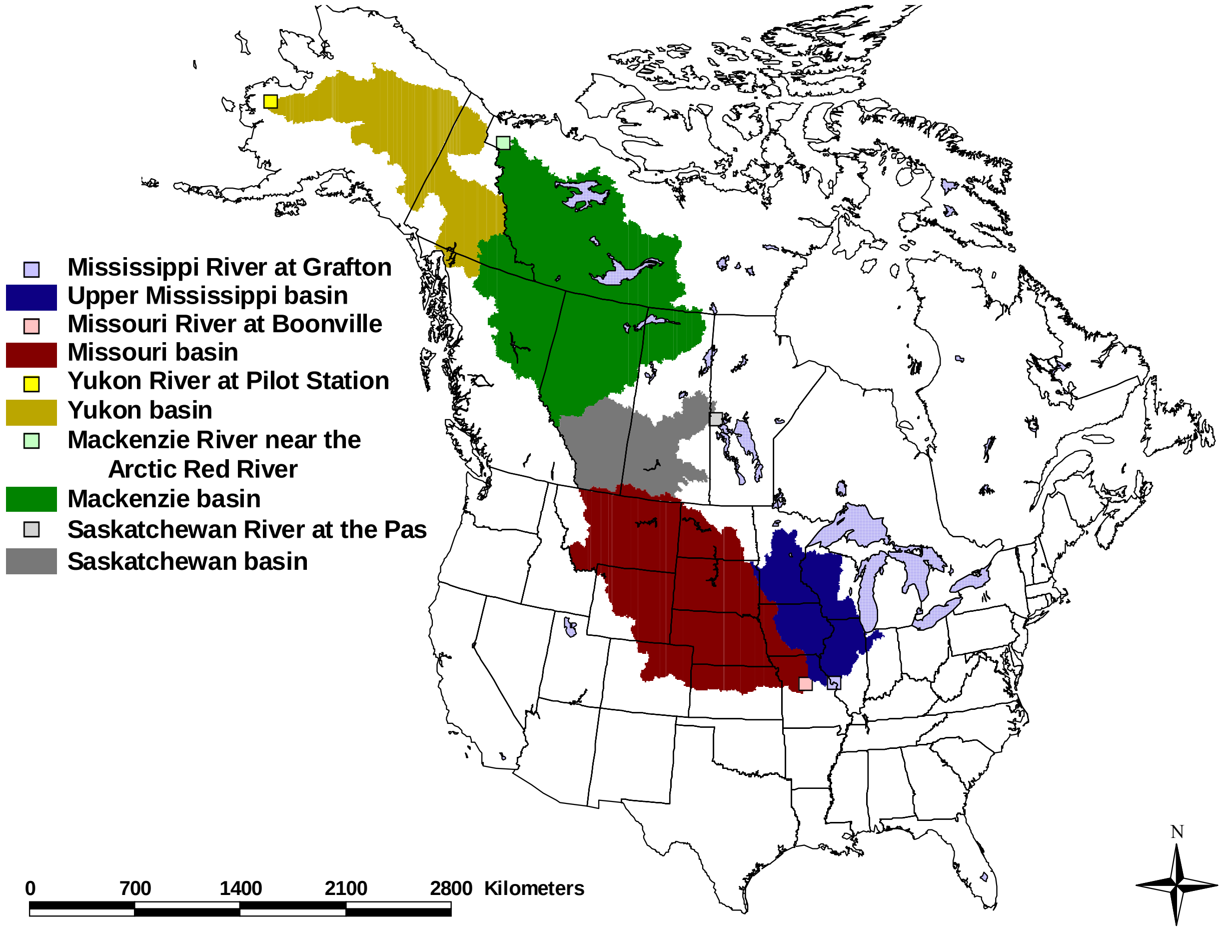


Figure 4.2. Outlines of major North American watersheds, along with symbols showing the associated river gages.

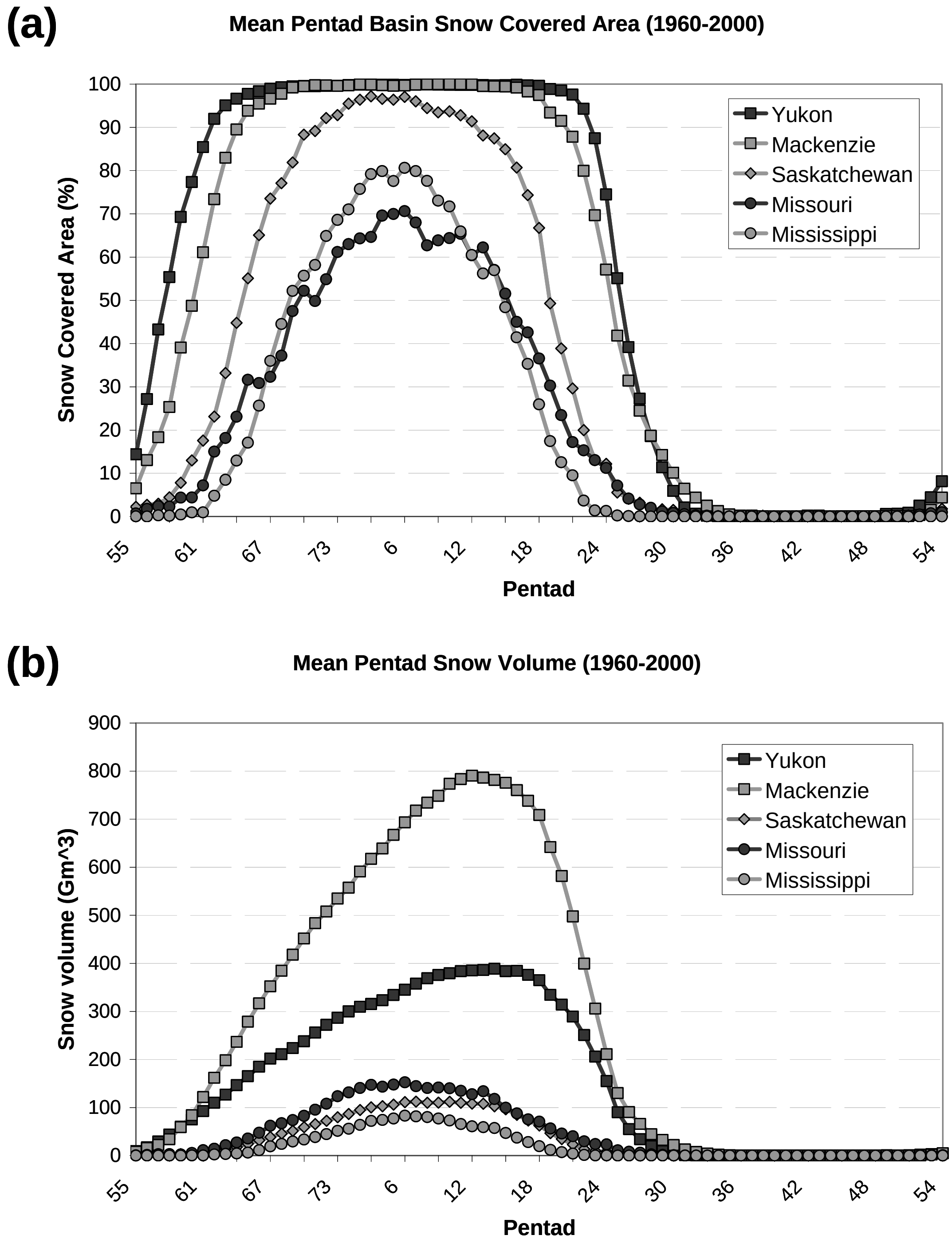


Figure 4.3. (a) Mean pentad snow covered area and (b) mean pentad snow volume for each major North American basin from 1960-2000.

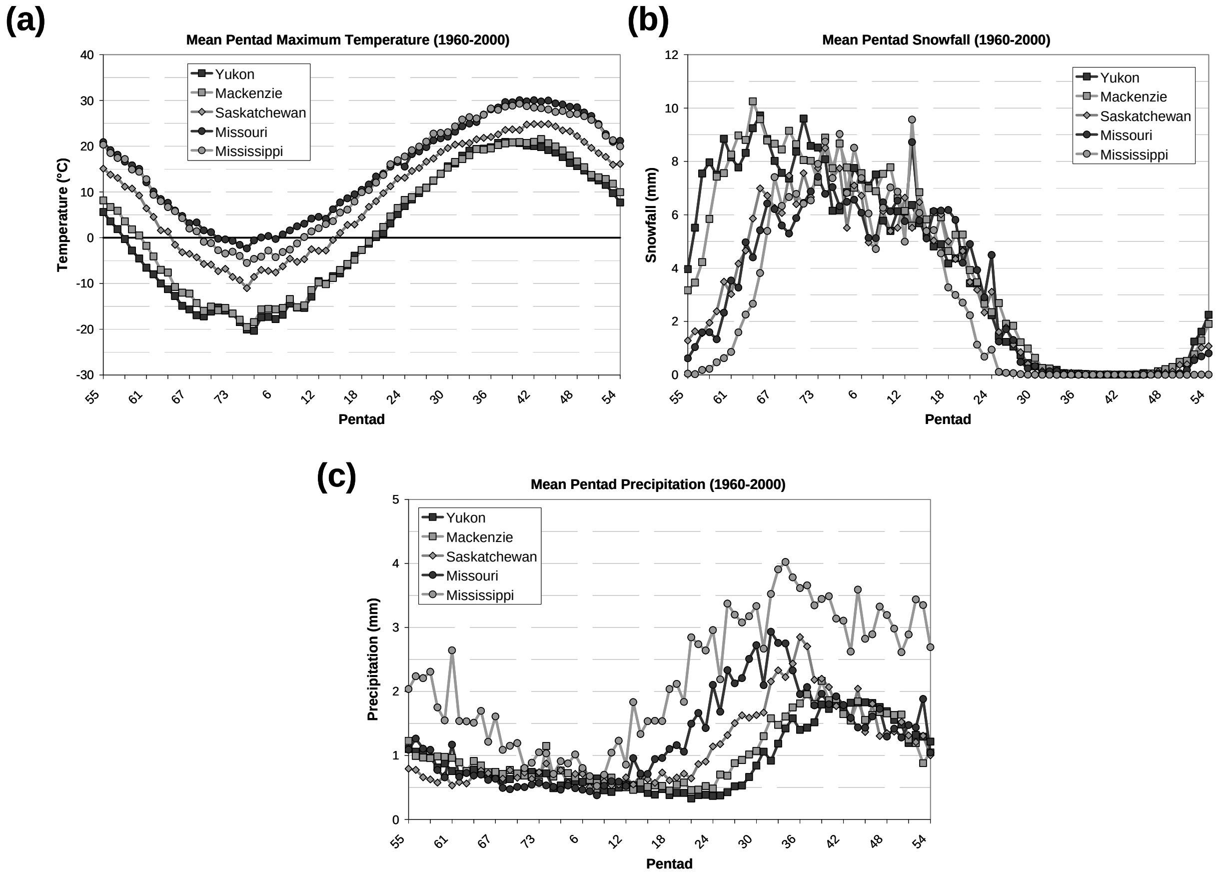


Figure 4.4. Mean pentad (a) minimum daily temperature, (b) daily snowfall, and (c) liquid precipitation for each major watershed over North America from 1960-2000.

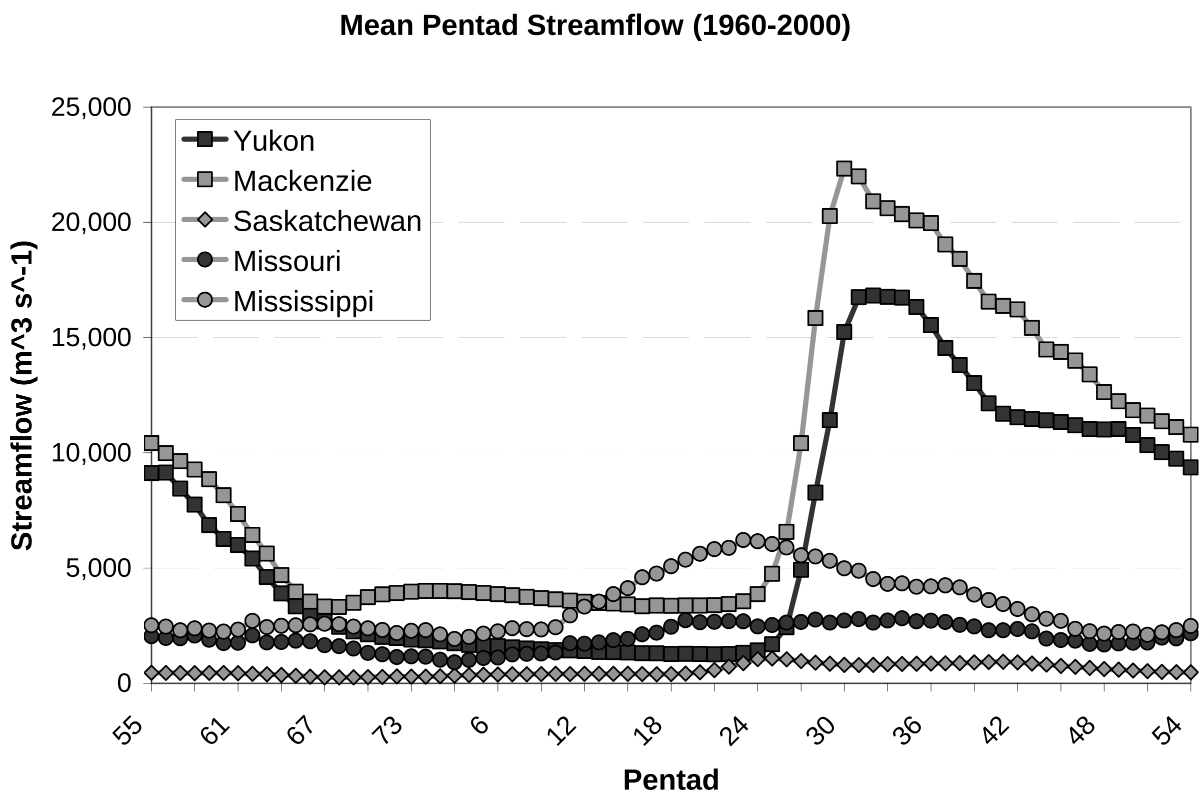


Figure 4.5. Mean pentad streamflow measured at the river gaging station for the associated major North American watershed.

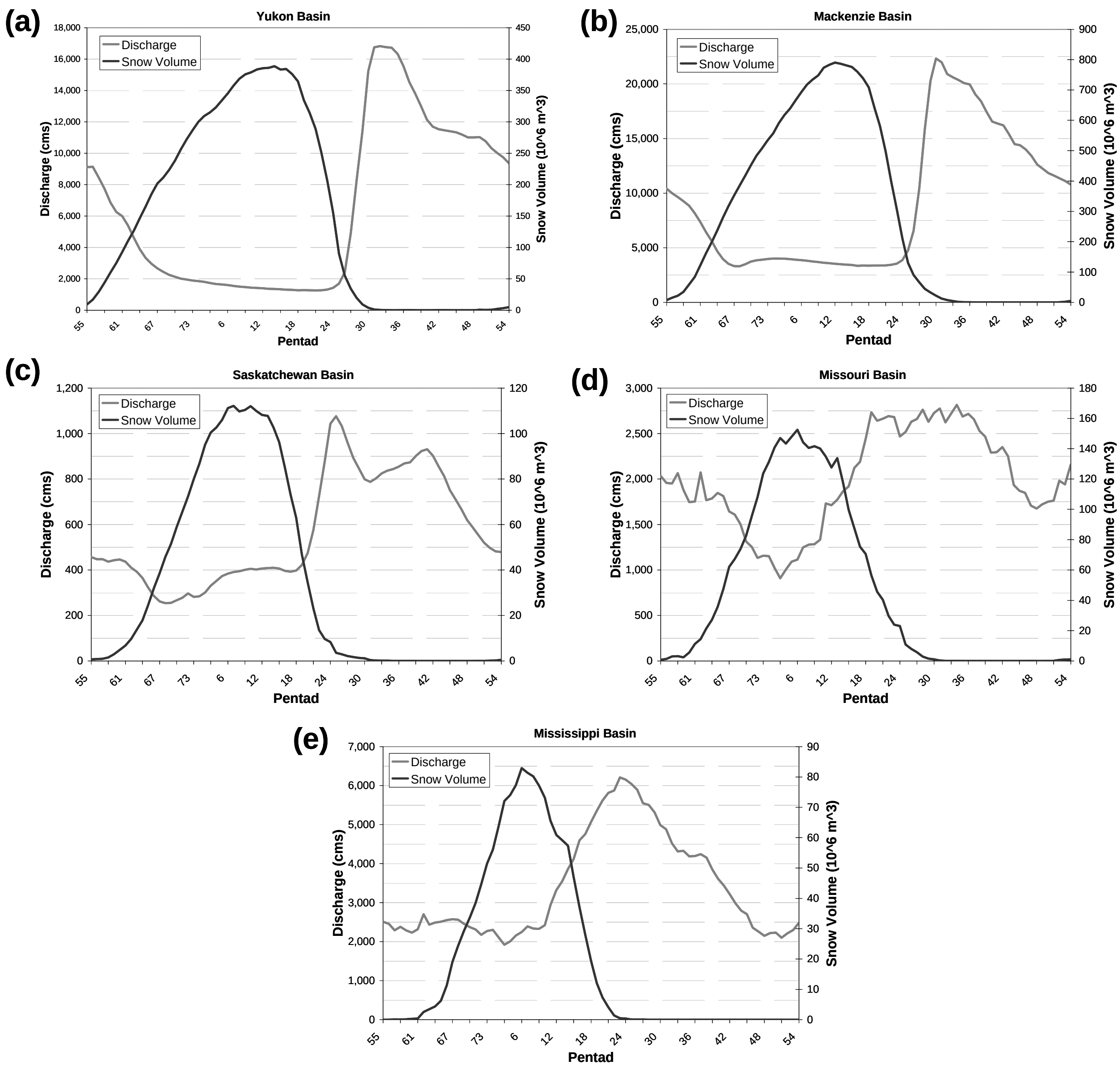


Figure 4.6. Seasonal snow volume and streamflow for the (a) Yukon, (b) Mackenzie, (c) Saskatchewan, (d) Missouri, and (e) Mississippi watersheds.

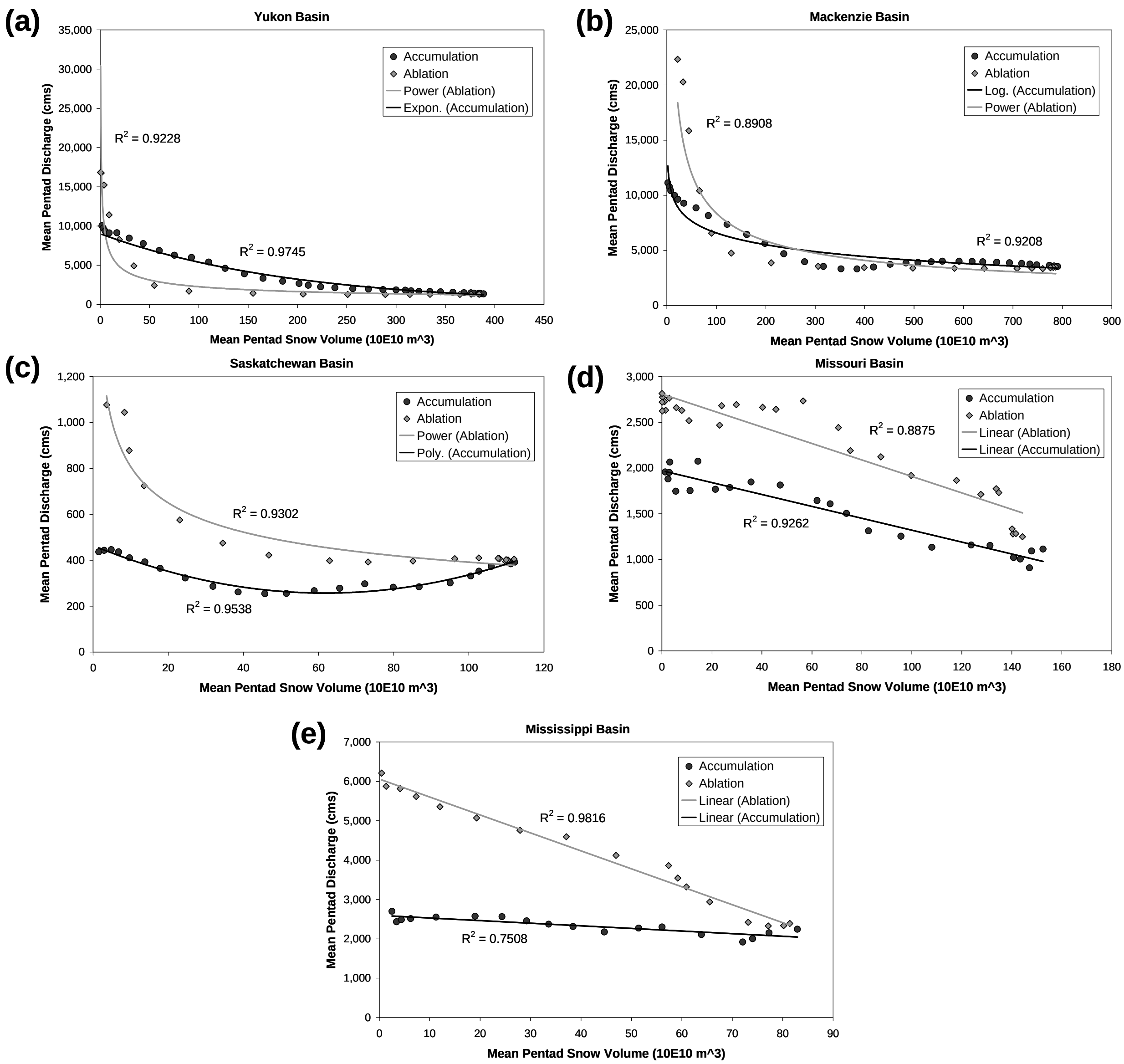


Figure 4.7. Scatter diagrams showing the strength and form of the relationship between snow volume and streamflow in the (a) Yukon, (b) Mackenzie, (c) Saskatchewan, (d) Missouri, and (e) Mississippi watersheds during the autumn accumulation and spring melt phase.

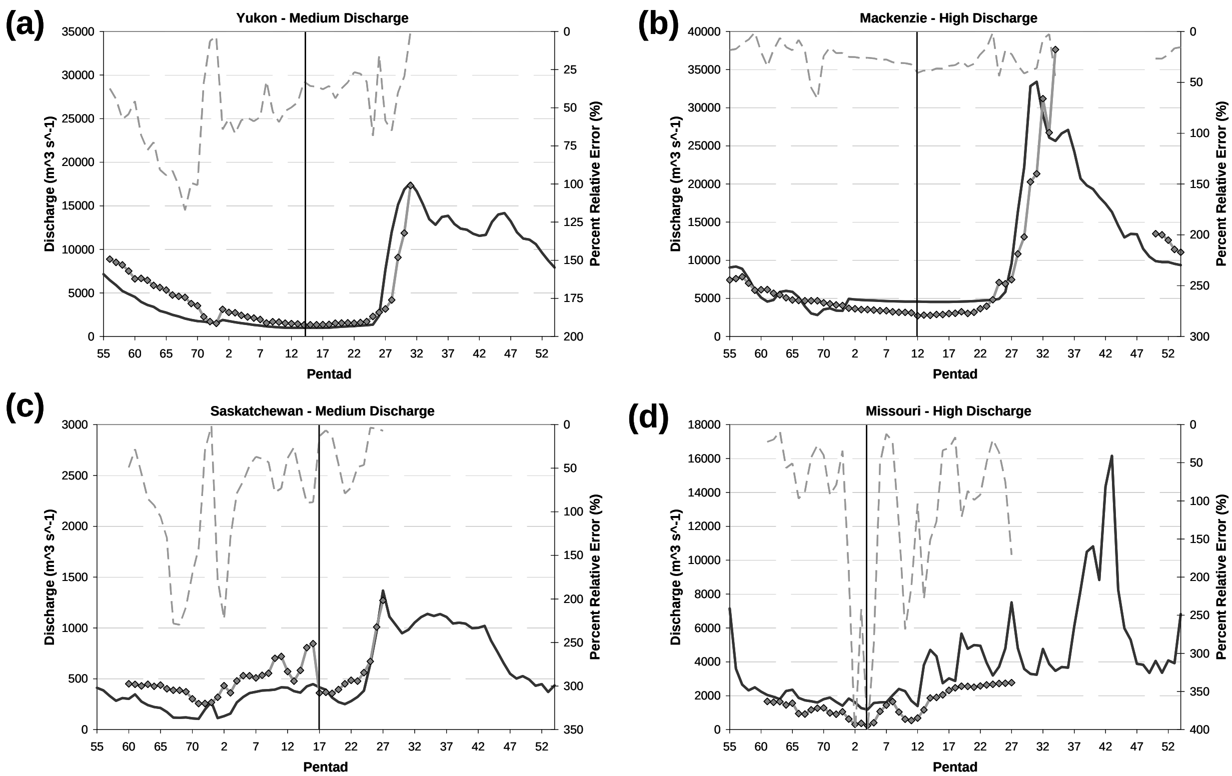


Figure 4.8. Predicted and observed discharge and associated percent relative error (dashed line) for (a) Medium discharge year in Yukon basin, (b) high discharge year in Mackenzie basin, (c) medium discharge year in Saskatchewan basin, and (d) high discharge year in Missouri basin. Note that scales within each figure vary.

Table 4.1. Variables describing snow cover and discharge patterns over major North American watersheds. The slope indicates the trend over time in the time series, shown as change per year (pentad yr^{-1}).

	Mean	Median	Slope of Regression	Standard Deviation
Initial pentad of snow cover	(first pentad when SCE > 5% consistently)			
Yukon	55.4	55.0	0.00	2.22
Mackenzie	55.8	56.0	-0.02	1.55
Saskatchewan	60.4	60.0	-0.07	2.35
Missouri	61.4	62.0	-0.01	3.13
Mississippi	64.2	64.0	-0.04	2.09
Pentad of peak snow volume	(pentad when snow volume reaches maximum value)			
Yukon	11.0	12.0	-0.04	5.33
Mackenzie	13.3	14.0	-0.09	3.00
Saskatchewan	9.0	10.0	-0.06	6.08
Missouri	5.6	4.0	-0.06	5.77
Mississippi	6.1	6.0	-0.04	5.45
Final pentad of snow cover	(last pentad when SCE > 5%)			
Yukon	29.2	30.0	-0.05	1.39
Mackenzie	30.9	31.0	-0.10	2.56
Saskatchewan	25.2	24.0	0.14	2.85
Missouri	24.9	25.0	0.11	2.43
Mississippi	20.1	21.0	-0.03	1.96
Snow cover duration	(number of pentads between initial and final pentad of snow cover)			
Yukon	46.9	47.0	-0.04	2.79
Mackenzie	48.0	48.0	-0.08	3.19
Saskatchewan	37.8	37.0	0.21	4.18
Missouri	36.5	37.0	0.13	4.53
Mississippi	28.9	29.0	0.01	3.35
Snow melt duration	(number of pentads between peak snow volume and final pentad of snow cover)			
Yukon	18.3	17.0	0.00	5.43
Mackenzie	17.5	17.0	0.00	3.93
Saskatchewan	16.2	15.0	0.20	6.69
Missouri	19.3	20.0	0.17	6.35
Mississippi	14.1	14.0	0.01	5.49
Maximum Discharge	(pentad when discharge reaches maximum value)			
Yukon	32.5	32.5	0.08	2.61
Mackenzie	32.3	31.0	-0.03	3.63
Saskatchewan	29.2	25.0	-0.11	8.02
Missouri	33.5	29.0	-0.11	15.00
Mississippi	29.9	26.0	0.23	13.26

Table 4.2. Regression equations for accumulation and ablation periods developed from mean snow volume and discharge from large North American watersheds. The variable x indicates the snow volume in 10^9 m^3 and y indicates streamflow in m^3s^{-1}.

Basin	Accumulation	Ablation
Yukon	$y = 9018.3\ e^{-0.0052x}$	$y = 19321x^{-0.466}$
Mackenzie	$y = -1544.3 \ln(x) + 13691$	$y = 90192x^{-0.5163}$
Saskatchewan	$y = 0.05x^2 – 6.71x + 464$	$y = 1668x^{-0.3148}$
Missouri	$y = -6.50x + 1969$	$y = -9.0x + 2808$
Mississippi	$y = -6.64x + 2599$	$y = -45.57x + 6058$

Table 4.3. Mean percent relative error between simulated and observed discharge in each major North American watershed for both the accumulation and ablation regression models.

	High Discharge		Medium Discharge		Low Discharge	
Basin	Accumulation	Ablation	Accumulation	Ablation	Accumulation	Ablation
Yukon	26.4	37.4	61.5	36.5	21.4	46.1
Mackenzie	28.1	62.2	16.0	42.1	23.6	29.7
Saskatchewan	27.4	24.4	95.1	30.9	113.9	35.4
Missouri	37.4	35.0	62.9	57.5	94.0	118.0
Mississippi	54.1	31.7	22.9	43.7	71.7	53.2

Table 4.4. Multiple regression coefficients resulting from analysis of peak snow volume and snowmelt and river discharge for large North American watersheds.

	Coefficients	Standard Error	Independent R^2
Yukon			
Max Snow Volume	38.9	9.6	0.4550
Pentad of Max Snow Volume	289.9	246.3	0.0910
Snow Melt duration	-100.4	141.4	0.0981
	Overall R^2:	0.5563	
Mackenzie			
Max Snow Volume	9.7	6.5	0.0203
Pentad of Max Snow Volume	-247.1	271.9	0.0856
Snow Melt duration	441.7	197.2	0.2147
	Overall R^2:	0.2907	
Saskatchewan			
Max Snow Volume	5.8	1.1	0.4173
Pentad of Max Snow Volume	14.7	16.1	0.0301
Snow Melt duration	9.8	11.1	0.0066
	Overall R^2:	0.4315	
Missouri			
Max Snow Volume	8.5	4.7	0.0756
Pentad of Max Snow Volume	-77.9	105.4	0.0191
Snow Melt duration	8.6	83.9	0.0026
	Overall R^2:	0.1029	
Mississippi			
Max Snow Volume	16.5	9.0	0.0985
Pentad of Max Snow Volume	93.7	133.2	0.0340
Snow Melt duration	7.8	94.8	0.0084
	Overall R^2:	0.1145	

CHAPTER 5

CONCLUSIONS

This dissertation begins by identifying temporal trends in snow depth across North America and quantifies the magnitude of interannual variations, using a gridded dataset of daily snow depth measurements from United States cooperative observers (U.S. Department of Commerce, 2003) and Meteorological Service of Canada observations (Braaten, 1996) from 1960-2000. An S-mode principal components analysis (PCA) was performed on monthly snow depth data that extracted regions exhibiting spatially coherent changes in snow depth over time.

Results from Chapter 2 indicates large areas of negative trends in snow depth across North America, with the most notable changes occurring in February and march. Little change is evident from October through January. Much of central and northern Canada shows significant decreases in snow depth (> 1 cm yr^{-1}, $p < 0.05$), with the greatest negative trend in March. Decreasing snow depths in March suggest an earlier onset of spring thaw.

During winter and spring, when continental snow depth is changing most rapidly, the PC score time series show the greatest decreases in central Canada and Quebec during the 1960-2000 study period. The greatest negative trends in snow depth occurring in central Canada in February and April.

Variations and trends in North American snow depth are influenced by the frequency and magnitude of snow ablation. The results from Chapter 3 show that ablation frequency has increased in March, and has a significant negative trend in May and June. These trends indicate a shift in ablation to earlier in the season, which provides a plausible explanation for the spring decrease in snow depth and snow cover extent. Similar trends exist in ablation intensity. Ablation frequencies at all intensities maximizes in April, but have positive and negative trends in March and May, respectively. The combination of an increase in frequency and intensity of ablation in March can have a considerable influence on regional hydrologic systems due to a more rapid influx of snowmelt runoff, and can also increase net radiation and temperature through an earlier removal of snow cover.

Analysis of ablation and snow depth shows a close relationship during March, when snow depth and ablation frequency have significant negative and positive trends, respectively. Analysis of energy budget components and air mass type and frequency for high and low ablation periods show that the sensible heat flux is the dominant mechanism responsible for initiating snow ablation in March. The increase in ablation frequency is most likely caused by an increase in the frequency of dry moderate (DM) air masses over central Canada, which shows a positive trend over the study period (0.22 days yr^{-1}; $p < 0.05$), leading to an increase in available sensible energy at the surface over.

Changes in the intensity and frequency of ablation over North America and subsequent effects on snow depth can have a considerable impact on hydrology due to changes in the volume and rate of snowmelt runoff. Chapter 4 shows that snow cover, expressed using snow volume, is highly correlated to discharge in major North American watersheds. The strength and form of the snow and discharge relationship within the large watersheds is determined from regression analysis during the accumulation and ablation phases of the winter season, which are used to develop statistical models within each of five individual basins. Results show that snow volume plays a major role in hydrologic patterns within North America, and is able to predict winter and early spring discharge with considerable accuracy in the higher latitude Yukon, Mackenzie, and Saskatchewan watersheds.

The statistical models perform best in the high latitude Yukon and Mackenzie watersheds in northern Canada and Alaska, where snow volume and streamflow show strong relationships during both the accumulation (R^2 = 0.97 and 0.92, respectively) and ablation (R^2 = 0.92 and 0.89, respectively) phases of the snow season. Model estimates of discharge in these basins show good predictive capabilities (percent relative error from 20-30% during accumulation, 40-60% during ablation), despite an increased sensitivity to streamflow estimates early in the ablation phase due to low snow volumes as well as difficulty in accurately predicting peak spring discharge. In the lower latitude Saskatchewan basin, rapid river response times lead to higher errors in predicted streamflow during the accumulation phase of the snow season, despite a strong relationship (R^2 = 0.95).

To define the influence of snow cover on peak annual discharge, which the basin-specific statistical models are unable to predict, a number of snow cover variables were included that described both the magnitude and timing of the peak snow volume and subsequent melt period. The peak annual discharge was found to be related not only to the volume of snowmelt runoff being introduced into the hydrologic system, but also the rate at which it was introduced. For the Yukon basin, 56% of the variance on peak annual discharge was explained by derived snow cover parameters, while only 29% was explained in the neighboring Mackenzie basin. The difference in explained variance is a result of basin characteristics such as vegetation and topography, which has a substantial influence on peak runoff in the Mackenzie basin. For the Saskatchewan basin, which has a much smaller areal extent and more rapid river response to runoff, 43% of the variance was explained by snow cover patterns.

This dissertation is an important step in identifying and quantifying the trends in North American snow cover from 1960-2000. Changes in snow depth over the higher latitudes can have substantial effects not only on regional hydrology, but also on the surface energy budget. Study of the processes related to the decrease in snow depth over North America is necessary in order to identify the causal relationships important in driving the trends in snow depth documented in Chapter 2. One such process involves a change in the frequency and/or intensity of ablation. Finally, because snowmelt supplies a substantial volume of water to many regions within North America, affecting water resources and flood potential, it is important to define these associations for the purposes of seasonal hydrologic forecasting as well as long-term climate research. Overall, this project is an important step in understanding the patterns and processes in the cryospheric system of North America, which is a vital component of the global climate system.

Results of this dissertation suggest a number of possible future research topics. Although there is a considerable increase in the amount of information gained by using snow depth instead of SCE, additional work should be completed using continental snow water equivalent (SWE) data. SWE provides direct information on snow mass, and examination of trends in SWE could provide useful results regarding trends in North American snow cover by including analysis of snow mass balance. Additionally, mechanisms responsible for snow ablation should be studied in more detail over more regions over North America to understand the role of the surface energy budget in snow ablation.

REFERENCES

Aagaard, K. and E.C. Carmack, 1989: The role of sea ice and other fresh water in the Arctic circulation. *J.Geophys. Res.*, **94**(C10), 14,485–14,498.

Aguado, E., D. Cayan, L. Riddle, and M. Roos, 1992: Climatic fluctuations and the timing of West Coast streamflow. *J.Climate*, **5**, 1468-1483.

Braaten, R. O., 1996: an updated Canadian snow depth database. Contract report prepared for Atmospheric Environment Service, September, 1996, 24 pp.

Brown, R. D., 1995: Spatial and temporal variability of North American snow cover, 1971-1992. *Proc. of the Eastern Snow Conf., 52nd Annual Meeting*, June 6-8, 1995, Toronto, Ontario, pp. 69-78.

Brown, R. D., 2000: Northern Hemisphere snow cover variability and change, 1915-97. *J.Climate*, **13**, 2339-2355.

Brown, R.D., 2003: Canadian snow depth data. Personal communication.

Brown, R.D. and R.O. Braaten, 1998. Spatial and temporal variability of Canadian monthly snow depths, 1946-1995. *Atmos.-Ocean*, **36**, 37-45.

Brown, R.D. and B.E. Goodison, 1996. Interannual variability in reconstructed Canadian snow cover, 1915-1992. *J.Climate*, **9**, 1299-1318.

Brown, R.D., B. Brasnett, and D. Robinson, 2003: Gridded North American monthly snow depth and snow water equivalent for GCM evaluation. *Atmos.-Ocean*, **41**, 1-14.

Cao, Z., M. Wang, B. Proctor, G. Strong, R. Stewart, H. Ritchie, and J. Burnford, 2002: On the physical processes associated with the water budget and discharge of the Mackenzie basin during the 1994/1995 water year. *Atmos.-Ocean*, **40(2)**, 125-143.

Cayan, D. R., 1996: Interannual climate variability and snowpack in the western United States. *J.Climate*, **9**, 928-948.

Changnon, D, T. B. McKee, and N. J. Doesken, 1993: Annual snowpack patterns across the Rockies: long-term trends and associated 500-mb synoptic patterns. *Mon. Wea. Rev.*, **121**, 633-647.

Colbeck, S.C., 1973: Theory of metamorphism of wet snow. USA Cold Regions Research Engineering Laboratory, Research Report 313.

Colbeck, S.C., 1983: Theory of metamorphism of dry snow. *J.Geophys. Res.*, **88**, 5475-5482.

Déry, S. J. and M.K. Yau, 2002: Large-scale mass balance effects of blowing snow and surface sublimation. *J.Geophys. Res.*, **107**, 4679, doi: 10.1029/2001JD001251.

Dyer, J.D., 2001: *Simulating the 1997 Red River floods utilizing a coupled snowpack and hydrologic model*. M.S. thesis submitted to the Graduate School, The University of Georgia, Athens, Georgia, USA.

Dyer, J.L. and T.L. Mote, 2002: Role of energy budget components on snow ablation from a mid-latitude prairie snowpack. *Polar Geog.*, **26**, 87-115.

Environment Canada, 2003: HYDEX/HYDAT database, National Water Data Archive, Water Survey Branch.

Frei, A. and D. A. Robinson, 1995: Regional signals in Northern Hemisphere snow cover during autumn and spring. In *Proc. Fourth AMS Conf. on Polar Meteorology and Oceanography*, Dallas, Texas, 127-131.

Frei, A. and D. A. Robinson, 1999: Northern Hemisphere snow extent: regional variability 1972-1994. *Int. J. Climatol.*, **19**, 1535-1560.

Frei, A., D. A. Robinson, and M. G. Hughes, 1999: North American snow extent: 1900-1994. *Int. J. Climatol.*, **19**, 1517-1534.

Goodison, B. E. and R. D. Brown, 1997: CRYSYS – Use of the Cryospheric System to Monitor Global Change in Canada. *In*: Proc. GER'97 (Int. Symp.: Geomatics in the Era of Radarsat), 24-30 May 1997, Ottawa, Ontario. Canadian Geomatics CD-ROM Vol. 1, Natural Resources Canada, Paper 259, 13 pp.

Groisman, P. Ya. and D. Easterling, 1994. Variability and trends of total precipitation and snowfall over the United States and Canada. *J. Climate*, **7**, 184-205.

Groisman, P. Y., T. R. Karl, and R. W. Knight, 1994: Changes of snow cover, temperature, and radiative heat balance over the Northern Hemisphere. *J.Climate*, 7, 1633-1656.

Groisman, PY; Knight, RW; Karl, TR. 2001: Heavy precipitation and high streamflow in the contiguous United States: Trends in the twentieth century. *Bull. Amer. Meteor. Soc.*, **82**, 219-246.

Grundstein, A. J. and D. J. Leathers, 1999: A spatial analysis of snow-surface energy exchanges over the northern Great Plains of the United States in relation to synoptic scale forcing mechanisms. *Int. J. Climatol.*, **19**, 489-511.

Gunn, K.L.S., 1965: Measurements on new-fall snow. McGill University, Montreal, Stormy Weather Group Scientific Report MW-44.

Hughes, M.G. and D.A. Robinson, 1996: Historical snow cover variability in the Great Plains region of the USA: 1910 through to 1993. *Int. J. Climatol.*, **16**, 1005-1018.

Hughes, M.G., A. Frei., and D.A. Robinson, 1996. Historical analysis of North American snow cover extent: merging satellite and station derived snow cover observations. *Proc. of the 1996 Eastern Snow Conf.*, Williamsburg, VA, 21-32.

Jones, P.D., 1994. Hemispheric surface air temperature variations: A reanalysis and an update to 1993. *J. Climate*, **7**, 1794-1802.

Kane, D. L., 1997: the impact of Arctic hydrologic perturbations on Arctic ecosystems induced by climate change, in *Global Change and Arctic Terrestrial Ecosystems*, *Ecol. Stud.*, vol. 124, pp. 63-81, Springer-Verlag, New York, 1997.

Kalnay, E., M. Kanamitsu, R. Kistler, W. Collins, D. Deaven, L. Gandin, M. Iredell, S. Saha, G. White, J. Woollen, Y. Zhu, A. Leetmaa, and B.Reynolds 1996: The NCEP/NCAR 40-year reanalysis project, *Bull. Amer. Meteor. Soc*., 77, 437–471.

Karl, T. R., P. Ya. Groisman, R. W. Knight, and R. Heim Jr., 1993: Recent variations of snow cover and snowfall in North America and their relation to precipitation and temperature variations. *J.Climatol.*, **6**, 1327-1344.

Kukla, G.J., 1979. Climatic role of snow covers. In: *Sea Level, Ice and Climatic Change*. IAHS Publ. 131, 79-107.

Kuusisto, E., 1986: The energy balance of a melting snow cover in different environments. *Proc. Of the Budapest Symp*., July 1986. **IAHS Publ. No. 155**, 37-45.

Leathers, D.J., T.L. Mote, K.C. Kuivinen, S. McFeeters, and D. R. Kluck, 1993: Temporal Characteristics of USA Snowfall 1945-1946 Through to 1984-1985. *Intl. J. Climatol.*, **13**, 65-76.

Leathers, D. J., D. R. Kluck, and S. Kroczynski, 1998: The severe flooding event of January 1996 across north-central Pennsylvania. *Bull. Amer. Meteor. Soc.*, **79** (5), 785-797.

Leathers, D. J. and B. L. Luff, 1997: Characteristics of snow cover duration across the northeast United States of America. *Int. J. Climatol.*, **17**, 1535-1547.

Leathers, D.J., Leathers, D.J., D. Graybeal, T.L. Mote, A.J. Grundstein and D.A. Robinson, 2004: The role of air mass types and surface energy fluxes in snow cover ablation in the central Appalachians, *J. Appl. Meteor.*, **43**, 1887-1898.

Mekis, É. and W.D. Hogg, 1999: Rehabilitation and analysis of Canadian daily precipitation time series. *Atmos.-Ocean*, **37**, 53-85.

Mellor, M., 1977: Engineering properties of snow. *J. Glaciol.*, **19**, 15-66.

Miller, J. R., and G. L. Russell, 2000: Projected impact of climate change on the freshwater and salt budgets of the Arctic Ocean by a global climate model, *Geophys. Res. Lett.*, **27** (8), 1183–1186.

Natural Resources Canada, 2003: Geomatics Canada, Earth Sciences Sector, Canada Center for Remote Sensing, GeoAccess Division [Available online: http://geogratis.cgdi.gc.ca/clf/en].

NOAA-CIRES Climate Diagnostics Center, 2005: NCEP Reanalysis Data, Boulder, CO, [Available online http://www.cdc.noaa.gov/].

Peterson, B.J., R.M. Holmes, J.W. McClelland, C.J. Vorosmarty, R.B. Lammers, A.I. Shiklomanov, I.A. Shiklomanov, S. Rahmstorf, 2002: Increasing river discharge to the Arctic Ocean. *Science*, **298**, 2171-2173.

Rango, A., 1997: Response of areal snow cover to climate change in a snowmelt-runoff model. *Ann. Glaciol.*, **25**, 232-236.

Richman, M. B., 1986: Rotation of principal components. *J.Climatol.*, **6**, 293-335.

Roads, J.O., S.-C. Chen, A.K. Guetter, and K.P. Georgakakos, 1994: Large-scale aspects of the United States hydrologic cycle. *Bull. Amer. Meteor. Soc.*, **75**, 1589-1610.

Robinson, D. A., K. F. Dewey, and R. R. Heim, 1993: Global snow cover monitoring: an update. *Bull. Amer. Meteor. Soc.*, **74**, 1689-1696.

Robinson, D. A., 1989: Construction of a United States historical snow data base. *Proc. of the 45th Eastern Snow Conf.*, 50-59.

Robinson, D. A., 1999: Northern Hemisphere snow extent during the satellite era. *Preprints: Fifth Conf. on Polar Meteorology and Oceanography*, Dallas, TX, American Meteorological Society, pp. 255-258.

Robinson, D. A. and A. Frei, 1997: A Northern Hemisphere snow cover climatology using satellite information. *Proc. of the 10th Conf. on Applied Climatology*, Reno, NV, American Meteorological Society, pp. 24-28.

Seaber, P.R., Kapinos, F.P., and Knapp, G.L., 1987: Hydrologic Unit Maps. U.S. Geological Survey Water-Supply Paper 2294, 63 p.

Serreze, M. C., J. E. Walsh, E. C. Chapin, T. Osterkamp, M. Dyugerov, V. Romanovsky, W. C. Oechel, J. Morison, T. Zhang, and R. G. Barry, 2000: Observation evidence of recent change in the northern high-latitude environment, *Climatic Change*, **46**, 159-207.

Sheridan, S.C., 2002: The redevelopment of a weather-type classification scheme for North America. *Int. J. Climatol.*, 22, 51-68.

Sheridan, S.C., 2005: Spatial Synoptic Classification homepage, Department of Geography, Kent State University. [Available online: http://sheridan.geog.kent.edu/ssc.html].

U.S. Department of Commerce, 2003: TD-3200 Surface Summary of the Day, National Climatic Data Center, NOAA. [Available online: http://www.ncdc.noaa.gov].

U.S. Department of the Interior, 1994: Hydrologic units maps of the Conterminous United States, U.S. Geological Survey, [Available online: http://water.usgs.gov/GIS/metadata/usgswrd/huc250.html].

U.S. Department of the Interior, 2003: NWISWeb water data. U.S. Geological Survey. [Available online: http://waterdata.usgs.gov/nwis].

Waggoner, P. E. and J. Schefter, 1990: Future water use in the present climate. *Climate Change and United States Water Resources,* P. E. Waggoner, Ed. John Wily and Sons, 19-39.

Willmott, C.J., C.M. Rowe, and W.D. Philpot. 1984: Spheremap. Center for Climatic Research, Department of Geography, University of Delaware, Newark DE.

Yang, D., D. Robinson, Y. Zhao, T. Estilow, and B. Ye, 2003: Streamflow response to seasonal snow cover extent changes in large Siberian watersheds. *J.Geophys. Res.*, **108**, 1-14.

Ye, H., H. R. Cho, and P. E. Gustafson, 1998: Changes in Russian winter snow accumulation during 1936-83 and its spatial patterns. *J. Climate*, **11**, 856-863.

Zektser, I. S. and H. A. Loaiciga, 1993: Groundwater fluxes in the global hydrologic cycle: Past, present and future. *J. Hydrol.*, **144**, 405-427.

Zhang, X., K. D. Harvey, W. D. Hogg, and T. R. Yuzyk, 2001: Trends in Canadian streamflow, *Water Resour. Res.*, **37**, 987-998.