VARIATIONS IN NORTHERN HEMISPHERE SNOWFALL: AN ANALYSIS OF HISTORICAL TRENDS AND THE PROJECTED RESPONSE TO ANTHROPOGENIC FORCING IN THE TWENTY-FIRST CENTURY

by

JOHN P. KRASTING

A Dissertation submitted to the

Graduate School-New Brunswick

Rutgers, The State University of New Jersey

in partial fulfillment of the requirements

for the degree of

Doctor of Philosophy

Graduate Program in Atmospheric Science

written under the direction of

Dr. Anthony J. Broccoli

and approved by

____

____

____

____

____

New Brunswick, New Jersey

October, 2008

ABSTRACT OF THE DISSERTATION

Variations in Northern Hemisphere Snowfall: An Analysis of Historical Trends and the Projected Response to Anthropogenic Forcing in the Twenty-First Century

By JOHN P. KRASTING

Dissertation Director:

Dr. Anthony J. Broccoli

Snowfall is an important feature of the Earth's climate system that has the ability to influence both the natural world and human activity. This dissertation examines past and future changes in snowfall related to increasing concentrations of anthropogenic greenhouse gases. Snowfall observations for North America, derived snowfall products for the Northern Hemisphere, and simulations performed with 13 coupled atmosphere-ocean global climate models are analyzed.

The analysis of the spatial pattern of simulated annual trends on a grid point basis from 1951 to 1999 indicates that a transition zone exists above 60° N latitude across the Northern Hemisphere that separates negative trends in annual snowfall in the mid-latitudes and positive trends at higher latitudes. Regional analysis of observed annual snowfall indicates that statistically significant trends are found in western North America, Japan, and southern Russia. A majority of the observed historical trends in annual snowfall elsewhere in the Northern Hemisphere, however, are not statistically significant and this result is consistent with model simulations.

Projections of future snowfall indicate the presence of a similar transition zone between negative and positive snowfall trends that corresponds with the area between the -10 to -15° C isotherms of the multi-model mean temperature of the late twentieth century in each of the fall, winter, and spring seasons. Redistributions of snowfall throughout the entire snow season are likely – even in locations where there is little change in annual snowfall. Changes in the fraction of precipitation falling as snow contribute to decreases in snowfall across most Northern Hemisphere regions, while changes in precipitation typically contribute to increases in snowfall. Snowfall events less than or equal to 5 cm are found to decrease in the future across most of the Northern Hemisphere, while snowfall events greater than or equal to 20 cm increase in some locations, such as northern Quebec. A signal-to-noise analysis reveals that the projected changes in snowfall are likely to become apparent during the twenty-first century for most locations in the Northern Hemisphere.

Acknowledgements

I would like to express my gratitude to the people who have made this work possible. First and foremost, this work would not have been possible without the financial support of the Graduate Assistance in Areas of National Need (GAANN) fellowship program, Cook College, and Dr. Anthony J. Broccoli, my advisor.

I would like to offer my utmost appreciation to my advisor, Dr. Anthony J. Broccoli. As a mentor, his constant support and patience has provided me with a strong base upon which to conduct this work. I have learned a great deal from his insight, and discussing the results presented in this dissertation with him over the past several years has sparked many new questions about this topic that I hope to pursue in the future.

My thanks are also expressed to my committee members – Dr. Brian Colle, Dr. David Robinson, Dr. Alan Robock, and Dr. Georgiy Stenchikov. I have valued their input through the process of conducting this research. I acknowledge the time and effort that was offered to serve on my committee.

I owe a debt of gratitude to my colleagues, Dr. Michael Previdi and Dr. Matei Georgescu, whose thoughtful comments and insight helped to improve this work.

My thanks are also expressed to Dr. David Robinson, Dr. Tom Mote, Dr. Jamie Dyer, and Dr. Michael Rawlins for supplying the observed data that were used in this study.

A special word of thanks to the computing support staff of the Department of Environmental Science, in particular Mina Azer, Jonathan Billings, Michael Ferner, Tom Grezlak, and Dana Price whose technical support and advice made much of this work possible. Additionally, I would like to thank all of my friends and colleagues at the Department of Environmental Sciences and the Center for Environmental Prediction whose companionship was an important part of my graduate education experience.

Finally, I would like to thank my friends and family, to whom this work is dedicated. I am especially grateful to Dorothy Krasting and Jill Halapchuk for their thorough editing of this manuscript. I am grateful for their never-ending advice and support. Without their inspiration, this work would not have been possible.

Dedication

To my family.

Table of Contents

Lists of Tables

List of Illustrations

Chapter 1

Review of Relevant Literature and Dissertation Overview

Future changes in snowfall as a result of anthropogenic climate change have been relatively unexplored compared to many other components of the climate system. Snowfall is one of the most complex climate features on Earth, and during its cycle - from the evaporation of water to the eventual melting on land – snow may be influenced by any one of a number of factors. Increasing carbon dioxide (CO_2) concentrations from the combustion of fossil fuels, for example, will alter several aspects of the Northern Hemisphere climate that influence snowfall – namely temperature and precipitation.

There are several ways to quantify snow and snow processes. On the ground, snow cover and snow water equivalence (SWE) may be used to examine snow that has already fallen. Since these variables are typically important on weekly or seasonal timescales, their study is most often relevant in locations that experience extended periods of time when the air and surface temperatures are below freezing. This work, however, will focus on snowfall, a related yet distinctly different atmospheric variable, that can and often does occur in locations in which the air temperature does not remain below freezing for extended periods of time

The potential impact from changes in snowfall is far-reaching as there are many meteorological, climatic, and societal dimensions to snowfall. From a climate perspective, snowfall changes potentially translate into snow cover changes that are important to the radiative processes on Earth, since freshly fallen snow reflects incoming radiation from the sun back to space. The amount of snow that falls is also an important part of the hydrologic cycle. In terms of human impact, snowfall has been linked to increases in morbidity and mortality, and there are potentially large economic disruptions associated with snow events.

Both hemispheric and regional scale processes influence snowfall in this region. The snowfall patterns also exhibit a large degree of seasonality and have significant, identifiable intra-seasonal variations. The use of climate models to understand the potential future changes in the features of eastern North American snowfall is important for impact assessment.

1.1. The Natural and Human Impacts of Snowfall

In terms of impact to the climate system, snow is an important component of the cryosphere, or the part of the Earth's surface that is frozen. Along with other components of the cryosphere, such as glaciers and sea ice, snowfall and snow cover are essential regulators of the planet's radiation budget and surface albedo (e.g., Robock 1980, Hall 2004), as well as of hydrological processes (e.g., Lettenmaier and Gan 1990). Snow is involved in several climate feedbacks (e.g., Robock 1985).

Snowfall is vital to the hydrologic processes that are present on Earth. Lettenmaier and Gan (1990) show that spring snowmelt is an important component of runoff in the western United States and this contribution may change significantly in a warmer climate. Groisman et al. (2001) report that there is a relatively small contribution to heavy runoff events from snowmelt in the eastern United States, but there is an overall contribution throughout the spring melt season.

Changes in snowfall may also influence large-scale variations in the atmospheric field. Blanford (1884) hypothesized that a link existed between heavy snowfall in the Himalaya Mountains and a weakening of the monsoon that develops each summer over India. There are indications that albedo, temperature contrasts, and complex hydrological feedbacks may play a important role in this teleconnection (e.g., Barnett et al. 1988). Cohen (1999) also provided evidence that the advection of snow-cooled air may be take place over great distances, thus potentially altering large-scale circulation regimes. Robock et al. (2003) found, however, that the strength of the Indian monsoon is predicted by warmer winter temperatures across Europe. Robock et al. (2003) demonstrated that the warmer temperatures in Europe are related to negative snow cover anomalies that are produced by circulation and temperature anomalies.

The eventual melting of snow is important to chemical and biological processes. In northern Eurasia, snowfall was shown to have an important influence on tree growth. (Vaganov et al., 1999) Increased winter precipitation, combined with a positive trend in early summer temperatures, contributed to a delay in the peak radial growth of trees. There is support that this delay is responsible for the decreased sensitivity of tree growth in this region to summer temperatures. Changes in growth resulting from changes in snowfall are important to understanding past climate variability, future changes in climate, and the carbon cycle (Vaganov et al. 1999)

Stottlemyer and Toczdlowski (1999) examined the interactions between snowmelt and stream water chemistry. It was found that in the forest floors of Michigan, the snowpack prevented the soil beneath from freezing, allowing the melt water to penetrate into the soil and ultimately into the rivers. In the winter months, this melt water was responsible for increased nitrate (NO_3^-) concentrations which were toxic to fresh water fish. In spring, the snowmelt increased soil water levels, which led to the removal of ions and dissolved organic carbon from the forest floors that are essential to the maintenance of the forest ecosystems.

Snowfall events are of particular interest since many sectors of the population are impacted in various ways. Rooney's (1967) conceptual framework of the urban snow hazard is an excellent examination of how snowfall can affect everyday life. The economic costs associated with snow removal as well as the costs involved in closing schools, businesses, and factories are to be considered when studying snowfall. Rooney also states that the transportation sector suffers extensive disruption from snowfall caused by airport and railway closures and by an increased number of traffic accidents.

Andreescu and Frost (1998) studied traffic accidents in Montréal, Canada, between 1990 and 1992. Their results indicated that the correlation coefficient between the number of traffic accidents and the snowfall amount was +0.48 and that this value was statistically significant at the 0.005% level. Eisenberg and Warner (2005) also conclude that there is an 18% higher risk of being involved in a fatal automobile collision and a 150% increase in property damage on a snowfall day as opposed to a dry day.

Removal is an important impact of snowfall events. The costs of snowfall removal are a substantial part of municipal budgets and these costs fall ultimately on the taxpayer. De Freitas (1975) argues that the "monetary costs of snowfall are significant." Snowfall removal also poses environmental challenges. Snowfall that lingers for several days following an event may accumulate dirt and pollution. In Sweden, this problem has forced municipalities to adopt a complex snow removal method that separates more polluted snow from less polluted snow (Reisenosdotter and Viklanker 2006).

In the area of health and human safety, snowfall events are associated with increased risks of sickness, injury, and even death. Rogot and Padget (1976) found that snowfall was associated with higher incidences of coronary artery disease and stroke mortality for five and six day periods after the event. Anderson and Rochard (1979) conducted a study in Toronto, Canada, from 1960 to 1974 in which they found that the daily rate of sudden death increased by 88% in men 65 years old and younger on days where there was snowfall. Based on emergency room reports from 1991 to 1994, Spitalnic et al. (1996) found that there was a 27% increase in cardiac arrests on snowfall days in Providence, Rhode Island. Gorjanc et al. (1999) examined death records in Pennsylvania between 1991 an 1996 and determined that total mortality increased on days when both snowfall was greater than 3 cm and the air temperature was below -7° C. In these cases the risks of ischemic heart disease were elevated for men as young as 35 years.

In studying the impact of snowfall, several attempts have been made to develop a classification scheme for nor'easters and strong mid-latitude cyclones. Such classification systems already exist for hurricanes (Simpson 1974) and tornadoes (Fujita 1971). Hart and Grumm (2001) used NCEP reanalysis data from 1948 to 2000 to express meteorological anomalies in the height, wind, temperature, and moisture fields as the number of standard deviations away from the mean. The more anomalous a storm was, the higher its ranking. Zelinski (2002) developed a more targeted system invoking the use of categories to classify winter storms in the eastern United States with an emphasis on nor'easters. Based on the storm's pressure anomaly relative to standard atmospheric pressure, deepening rate, and maximum pressure gradient, the storms were ranked on a 1-5 categorical system, with 5 being the most intense storm.

Kocin and Ucellini (2004) demonstrate, however, that some of the most expensive and crippling snowstorms in the northeastern United States do not rank at the top of these anomaly-based classification schemes. Since meteorological severity does not necessarily agree with total impact resulting from a snow event, Kocin and Ucellini (2004) developed the Northeast Snowfall Impact Scale, or NESIS, which attempts to factor the human impact when ranking snow events. NESIS uses the population-weighted snowfall of a given event to classify snowfall events into one of five categories, which have been calibrated using the climatology of 30 snowstorms that affected the northeastern corridor of the United States in the latter half of the 20th century. Although the NESIS approach attempts to account for the impact associated with snow events, the index does not account for the fact that technology has changed over time and may not provide an accurate comparison of events.

The inability of the scientific community to develop an appropriate ranking system of snowfall events is an example of how complex the relationships are among snow, climate, and society. Both the natural and human impacts of snowfall are far-reaching. It is these relationships with other components of the climate system as well as the effects on human health, safety, and prosperity that motivate the study of this topic. Advancing the scientific knowledge of how Northern Hemisphere snowfall might change in the future will serve to benefit the understanding of the relationship between the climate and society on the whole.

1.2. Observed Synoptic-scale Spatial and Temporal Variability of Northern Hemisphere Snowfall

Constantin François de Chassebœuf, comte de Volney, was a successful French Renaissance man who escaped persecution following the French revolutionary movement and sought refuge in the United States. Volney, as he was later known, had particular interests in geology, weather, climate, and especially wind circulation patterns. Beginning his journey in Philadelphia, Volney traveled for three years throughout the portion of the United States that lies east of the Mississippi River and later published his observations on the climate and geology of the region in his book titled *Tableau du Climat et Sol des États Unis* (Volney 1804). This work is believed to be the first written climatology of the United States.

In his 1804 book, Volney recorded some of the first fundamental observations on snowfall. Through comparisons with locations in Europe at similar latitudes, Volney was first to identify the latitudinal gradient of snowfall in eastern North America and the importance of the Atlantic Ocean as a moisture source for snowfall. Volney was also one of the first individuals to correlate northeasterly winds with periods of wet and cold weather along the coastal areas of New England.

Other early attempts were made to create a climatology of North American snowfall. Brooks (1915) developed monthly and seasonal climatologies of snowfall for the United States and more specifically for New England (1917). Brooks (1917) intuitively concluded that “air temperature is probably the first factor in determining snowfall” and that “without precipitation, however, there can be no snowfall.”

More recently, an effort has been made to understand the complexity of the relationships among temperature, precipitation, and snowfall. Davis et al. (1999) examined the temperature-snowfall relationships across Canada using monthly snowfall and temperature data that spanned the years 1945 to 1994. Their analysis indicates that there is a positive relationship between snowfall and temperature at higher latitudes and also in the mountainous regions of southwestern Canada. Negative relationships are present at more southern latitudes along the coast and also along the lee of the Rocky Mountains. Furthermore, the analysis of the seasonal cycle of snowfall also indicates that the transition seasons – early fall and late spring – are dominated by negative relationships between snowfall and temperature at most locations throughout Canada.

Namias (1960) examined snowfall over the eastern United States and determined that although there is a relationship among temperature, precipitation, and snowfall, this relationship is sometimes unclear since more complex processes may exert a discernable influence on the overall climatology of North American snowfall. Schemenauer et al. (1981) present a more sophisticated explanation for the occurrence of snowfall and discuss four associated atmospheric conditions:

1.) Sufficient moisture and active nuclei at a temperature suitable for the formation and growth of ice crystals,
2.) Sufficient depth of cloud to permit growth of snow crystals by aggregation and accretion,
3.) Temperatures below 0° C in most of the layer through which the snow falls, and
4.) Sufficient moisture and nuclei to replace losses caused by precipitation.

Schemenauer et al. (1981) indicate that snowfall amounts are related to the strength of the large-scale vertical motion and terrain lifting. The amount of snowfall produced is also related to the duration of snowfall events, which are a function of the speed and track of synoptic-scale disturbances.

In addition to the processes described by Schemenauer et al., local terrain effects, urbanization, and proximity to water also are important when considering the climatology of snowfall. Processes occurring on the mesoscale are important in understanding the complexity of snowfall. Combined with enhanced computer simulation capabilities, an understanding of the mesoscale processes associated with snowfall, have helped operational weather forecasters more accurately predict snowfall events. For example, Ucellini et al. (1995) cite forecaster recognition of mesoscale aspects as one of the main reasons why forecasting the March 1993 “superstorm” that affected the eastern half of the United States was such an unprecedented success since accurate snowfall totals were forecast well before the onset of the storm.

Changnon (2006) established a climatology of extreme snowfall events from 1948 to 2001 based on 204 first-order observation sites. In this study, mountain ranges in both the western and eastern United States have higher values of snowfall that are associated with the higher elevations. Changnon (2006) also establishes that the Great Lakes enhance regional snowfall frequency by 20 to 50%. The analysis was extended to all snow events in the United States and it was determined that the average event distribution was largely latitudinal and that elevation continued to play an important role in the frequency distribution, particularly in the western part of the country (Changnon and Changnon 2006).

Perhaps one of the most objective analyses of the characteristics of snowfall across the United States was performed by Harrington et al. (1987). Using harmonic analysis of monthly snowfall totals, they showed that the peak of the snowfall season occurs in mid-February across much of the continental United States. The snowfall peak of the snowfall season for the Great Lakes and the Pacific Northwest, however, occurs slightly earlier in February. Harrington et al. also found that large monthly snowfall values develop in the interior of the United States in October and November. Snowfall then spreads eastward as the fall and winter seasons progress forward in time. In early winter, lake-effect snowfall is also at its peak.

Harrington et al. (1987) cite several mechanisms responsible for the observed patterns of snowfall. First, the eastward progression of snowfall during the late fall is consistent with an observed eastward drift of the climatological 500-mb height trough that develops over central and eastern North America. Second, as snowfall intensifies in all areas through the winter season, the expansion of the Northern Hemisphere circumpolar jet allows cold polar air to penetrate into the continent. Third, the position of the Colorado and Alberta storm tracks in early winter allows for increased snowfall in the Midwest and upper Great Plains while forcing cold air to flow over the relatively warm waters of the Great Lakes producing lake-effect snow as cyclones pass by. Fourth, baroclinicity – which increases along the Carolinas and the eastern seaboard in January and February – promotes the formation, intensification, and maintenance of strong coastal storms and nor'easters.

North America is fortunate to have had extensive studies of snowfall performed over the 20th century. Similar studies are not as readily available for other locations in the Northern Hemisphere. With the advent of satellite retrieval of snow depth and SWE in the 1970s, global climatologies of these variables were created. Two problems, however, exist with the data. First, the record of satellite measurements is comparatively shorter than other climate records. Even where local measurements of snow depth exist prior to the 1960s, concerns about representativeness exist when spatial patterns are derived from point observations. Secondly, problems arise when inferring snowfall from these records since melting may create substantial differences between the snow depth and snowfall climatologies (e.g., Lydolph 1977).

With those cautions in mind, some relevant deductions about the climatology of snowfall can be derived for the Northern Hemisphere using snow depth data. Ye et al. (1998) state that:

> [The] mean snow depth distribution patterns [of Russia] seems to be related to climatic conditions that are determined by latitude, elevation, surface characteristics, and the proximity to open water bodies.

Kripalani et al. (1999) developed gridded data from the historical Soviet daily snow depth data (HSDSD) product and examined the climatology of Russian snow depth between 1881 and 1985. The authors concluded that large snow depths in Western Russia are associated with cyclonic circulation present over the region. The northerly flow on the western side of the associated trough favors the advection of cold air into the regions, thus preserving the snow pack.

The analysis of Kripalani et al. (1999) also identified an area of large snow depth between 40º - 120º E and 55º - 70º N, and in particular between 80º - 100ºE. In this smaller region, snow depth steadily increases from 10 cm in October to 80 cm in March before complete melting begins to take place. This study also identified an area of deep snow cover to the west of the Kamchatka Peninsula. Kripalani et al. (1999) cite the results of Lyndolph (1976), however, where the area of maximum snowfall in western Russia does not coincide with the area of maximum snow depth. The discrepancy arises as the melting of snow prevents the snow depth from rising above 20 cm.

In an effort to understand the previously discussed Blanford hypothesis (1884), Zhang et al. (2004) examined the synoptic climatology of heavy spring snow depth seasons over the Tibetan Plateau and their linkage with the strength of the Indian Monsoon. The longwave pattern at 500 mb indicates that a semi-permanent planetary trough exists throughout southeastern Asia during the spring. Zhang et al. (2004) propose that snowfall is increased in spring seasons when the India-Burma trough in the subtropical jet stream is sharpened. A sharper trough enhances cyclonic circulation and lift over the plateau while moistening the atmosphere with southerly winds. Sea surface temperature warming in the Northern Indian Ocean is believed to be responsible for the increase in the amplitude of the India-Burma trough.

1.3. Inter-annual Variability of Northern Hemisphere Snowfall

There are no direct evaluations of the changes in snowfall over the Northern Hemisphere in the Fourth Assessment Report (AR4) of the Intergovernmental Panel on Climate Change (IPCC). Rather, snowfall is incorporated into the discussions of snow cover changes and there is an implicit assumption that changes in snowfall are similar. In the report, Northern Hemisphere snow cover since the 1920s has decreased little during the heart of the winter seasons. Decreases in snow cover extent and increased variability, however, are noted in the fall and spring transition seasons. Upward trends in higher latitude temperatures are believed to be responsible for the observed trends in snow cover, as melting and precipitation type are both governed by the temperature (Lemke et al. 2007).

Several authors have examined how North American snowfall patterns have changed over time. Scott and Kaiser (2004) examined snowfall trends from 1948 to 2001 and observed that snowfall generally increased over a narrow band from Colorado to Wisconsin to the lee of Lakes Ontario and Erie. This study further noted that the snowfall season has decreased in length as much as 6 days per decade in the region extending from the Pacific Northwest to eastern Kansas, to the Ohio Valley and to the northeastern United States.

In other parts of the Northern Hemisphere, snowfall changes were examined. Laternser and Schneebeli (2003) examined long-term trends in snow cover, snow depth, and snowfall days over the Swiss Alps from 1931 to 1999. All three of these metrics exhibited gradual increases since 1930 until statistically significant decreases began in the early 1980s. The negative trends in the snow metrics were the result of a decrease in the fraction of precipitation that fell as snow. The authors caution, however, that latitude and elevation contribute significantly to variations in the trends.

Karl et al. (1992) found that the area-averaged snowfall across northern Canada (55-80º N) was increasing at a statistically significant rate of 8.8 cm decade^{-1} in the late twentieth century and that this increase was due to an increase in precipitation and not a redistribution of precipitation between snowfall and rainfall. In southern Canada (< 55º N), the trend was -0.65 cm decade^{-1}, with most of the decrease occurring during the 1980s. The climate record of precipitation in southern Canada indicates that precipitation has significantly increased by 9.7 ± 2.1% per century, implying that the fraction of precipitation falling as snow, or snow fraction, has decreased. Although statistically insignificant, analysis of the United States climate record also indicates an increase in total precipitation with decreases in both snowfall and the snow fraction since 1950 (Karl et al. 1992).

Hartley and Keables (1998) examined the snowfall climatology in New England from 1950 to 1992. They characterized the late 1950s to the late 1960s as a period of relatively increased snowfall followed by a more variable period in the 1970s and 1980s that had a discernible downward trend in snowfall. Principal component analysis was performed on the temperature, precipitation, and snowfall fields to examine the variability and relationships among the three variables. They found that temperature and precipitation only explain 42% of the total variability in snowfall.

The patterns of temperature and precipitation, however, are often influenced by both relatively high-frequency synoptic scale weather variability and low-frequency climate variability. All of these modes of variability have an influence in determining the inter-annual variability of snowfall over eastern North America. Building on the work of Karl et al. (1992), Groisman and Easterling (1994) examine snowfall and precipitation across Canada and the United States more closely. Groisman and Easterling indicate that most of the observed increases in North American annual precipitation during the twentieth century south of 55º N have occurred along the eastern sections of the continent and are consistent with a negative trend in the snow fraction in this region. Groisman and Easterling also examined the Southern Oscillation Index (SOI) and found that negative values, corresponding to El Niño conditions, are associated with the increases in precipitation.

Kunkel and Angel (1999) discuss the relationship between El Niño-Southern Oscillation (ENSO) and snowfall across the continental United States between 1951 and 1997. They found that there was a stark contrast between the composite map of snowfall for five El Niño and five La Niña years. Snowfall during El Niño events was characterized by a general decrease in snow across much of the northern United States and this decrease is statistically significant at the 10% level in the Ohio and Mississippi river basins. La Niña years indicate an opposite relationship to snowfall, although not statistically significant, and exhibit increases in snowfall in the area over the northern Great Lakes.

Hartley and Keables (1998) further examined how cyclonic activity, storm tracks, and western Atlantic sea surface temperatures (SSTs) contributed to snowfall variability in New England between 1950 and 1992. Using a composite analysis technique of high and low snowfall seasons, negative 700 mb anomalies over the continental United States and positive 700 mb anomalies suggestive of a more meridional circulation regime were associated with high snowfall winters. Low-level cold advection was enhanced by storm tracks along and just to the south of the New England coastline, and the high variability of the North Atlantic Oscillation (NAO) and the Pacific-North American (PNA) patterns in the 1970s coincides with the high variability present in the snowfall record. Most importantly, the results indicate a significant correlation of SSTs with temperature, precipitation and snowfall.

Bradbury et al. (2002) examined 500 mb analyses in order to understand the mechanisms responsible for winter precipitation variability in New England. Bradbury et al. demonstrated that when the persistent winter trough shifts eastward, precipitation is reduced in the interior parts of New England as the storm tracks were also shifted eastward off the coast. The authors argued, however, that even when the trough shifted eastward, there was little effect on stations located along the immediate coast as they were still within the range of the main storm tracks. Additionally, the authors suggested that the increases in the amplitude of the trough were also associated with the negative phase of the NAO and enhanced North Atlantic blocking activity that resulted in colder temperatures along the East Coast. While there were robust relationships among the characteristics of the trough and the patterns of temperature and precipitation, there was no clear link between snowfall itself and trough properties. Bradbury et al., however, noted that cooler regional SSTs and a negative phase of NAO accompanied this eastward shift of the trough axis and also showed the link between SSTs and snowfall that was seen in Hartley and Keables (1998).

Kirby et al. (2001) suggest that a 20- to 30-year periodicity in the circumpolar vortex established by examining δO^{18} records from a lake in Fayetteville, NY over the past 1000 years may be partly responsible for the decadal variability of snowfall over North America. Hartley and Keables (1998) caution, however, that the record of snowfall is not extensive enough to examine how all of the possible modes of low frequency variability could explain the patterns of New England snowfall.

There have been several studies that examined snowfall over the Greenland ice sheet. Thomas et al. (2006) examined laser altimeter measurements of surface elevations and concluded that the ice is thickening at elevations greater than 1500 m. The authors report that these results are indicative of increasing snowfall in warmer climate. Box et al. (2006) studied the mass balance of the Greenland ice sheet using the Polar MM5 model and found results consistent with the observations of Thomas et al. (2006).

Very little work has been done to assess the ability of the modern atmosphere-ocean general circulation models (AOGCMs) to capture changes in snowfall. Roesch

(2005) examined the surface albedo and snow cover in the coupled climate models that were used for the Fourth Assessment Report (AR4) released by the IPCC in 2007. One of Roesch's key findings was that snow water equivalence (SWE) was often over-simulated in the 20th century climate experiments performed for the report. Roesch argues that positive snowfall anomalies in the models are responsible for the large SWEs. It may also be possible that a cold bias present in the CMIP3 models (John and Soden, 2007) may also contribute to an overestimation of SWE. The snowfall in the models was evaluated with rather coarse precipitation data from the Global Precipitation Climatology Centre and CRU temperature data were used to indicate snow if the temperature was below freezing. A more thorough examination of the models' ability to capture the patterns, magnitudes, and inter-annual variability is an essential step in interpreting any model-simulated changes in snowfall for the 21st century.

1.4. Future Changes in Northern Hemisphere Snowfall

One of the fundamental tenets regarding snowfall is that it is governed by the coexistence of sufficiently cold temperatures and the occurrence of precipitation. When examining snowfall in a warmer climate, one of two scenarios is possible for a given location in eastern North America. First, increasing air temperatures associated with anthropogenic-induced global climate change might reduce the amount of snowfall by reducing the fraction of the total precipitation that falls as snow. In places where air temperatures are cold enough that increases in temperature will not impact the partitioning of precipitation into snow and rain, the increased moisture content of warmer climate (e.g., Held and Soden 2006) may lead to increased snowfall. It is unclear how the interplay between temperature and precipitation will vary both in space and time over eastern North America. Furthermore, based on their observations, Robinson and Frei (2000) state that snow may be a good indicator of climate change. The point at which future changes in eastern North American snowfall will emerge from the background variability of the time series has not been examined.

Climate models can provide some useful information regarding future snowfall changes. In the IPCC AR4 report, AOGCMs are shown to produce more precipitation over North America (Meehl et al. 2007) and this result is shown to be a consistent and robust feature among climate models (Held and Soden 2006). With a lesser degree of confidence, however, the report also indicates that the circumpolar vortex may be shifted northward and the incidences of North Atlantic blocking patterns may change as the AO/NAO indices are projected to favor their positive modes (Christensen et al. 2007).

Snowfall is often an implicit part of the IPCC's discussion of future snow cover. A later onset to the snowfall season in fall, coupled with earlier spring snow melts, contributes to an overall shortening of the snow season and decreased snow depth over North America. The report states, however, that projected increases in precipitation may lead to increased snow accumulations in some locations across

North America – particularly near the Arctic Ocean and in some parts of the Rocky Mountains (Christensen et al. 2007).

AOGCMs have already been used to answer some specific questions related to snowfall changes and their potential impacts. As an example, Elasser and Messerll (2001) examined how climate change might impact the tourist ski areas in the Swiss Alps. Based on a projected rise in the snow levels, Elasser and Messerll found that the number of ski areas with reliable snow cover would decrease in all of the main tourist regions of the Swiss Alps. Examining future changes in snowfall over eastern North America will have a similar degree of applicability.

Kunkel et al. (2002) analyzed results from two AOGCMs in an attempt to better understand how the frequency of lake effect snowfall events might change in a warmer climate. While there might be a period in the early part of the 21st century where the length of time that the Great Lakes remain unfrozen enhances snowfall, there is an indication that warming of the surface air temperatures will likely reduce the frequency of heavy-lake effect snowfall events by the late 21st century. Lake-effect snows are an important part of the eastern North American snowfall climatology, and further studies targeting future changes in snowfall for this area are needed.

1.5. Dissertation Overview

The overall goal of this work is to understand future changes in snowfall over the Northern Hemisphere. More specifically, the aims of this project are to:

1.) Analyze how snowfall has changed historically over the Northern Hemisphere and assess the ability of modern AOGCMs to simulate the observed climatology, seasonal variation, and inter-annual variability of snowfall.
2.) Determine how snowfall over the Northern Hemisphere will change in the 21st century and examine the mechanisms involved in these changes.

These goals are motivated by the potential impact that may arise from future changes in snowfall.

In Chapter 2, the data used in this project are detailed and evaluated. This study employs both observed snowfall data as well as results from historical climate simulations performed with a suite of state-of-the-art AOGCMs made available through the third phase of the Coupled Model Intercomparison Project, or CMIP3. The strengths and weaknesses of the data are explored and comparisons are made with model output.

In Chapter 3, historical patterns in snowfall over eastern North America are examined. Data from both observations and simulations of the 20th century climate considered in the IPCC AR4 report are utilized to determine if there are trends in snowfall and, if so, whether or not they are simulated by the models. Explanations are also sought for the observed and simulated local and regional trends in snowfall.

Trends in annual and monthly snowfall are studied in Chapter 4 using the IPCC future climate simulations performed with the suite of CMIP3 AOGCMs. The interplay between changes in snow fraction, which is related to temperature, and changes in precipitation to produce regionally averaged changes in snowfall, is also studied. Using daily output that is available for the GFDL CM2.1 model, changes in the frequency of daily snowfall events are examined. Finally, a signal-to-noise analysis is performed to determine when the trends in annual snowfall might emerge from the background variability.

In Chapter 5, a summary and discussion is provided based on the scientific experiments and analyses performed in the dissertation. The results are discussed within the larger context of regional climate change and potential impacts will be postulated. Limitations of this study are examined and recommendations for future work will be made.

Chapter 2

Observed and Simulated Snowfall Data: Description and Evaluation

2.1. North American Gridded Snowfall

Several different techniques have been developed for observing meteorological quantities related to snow. Measuring snow cover is often accomplished through the use of snow courses, snow pillows, or satellite retrieval. Measuring snowfall, however, is different. Precipitation gauges are used in some locations, and fence systems have been developed to protect the gauges from wind errors and snow drift (e.g. Rechard and Larson, 1971). Direct depth measurements of freshly fallen snow are also routinely taken at observing sites. The United States Cooperative Observer Network (COOP) instructs observers to take a daily measurement of snowfall at 7 A.M. local time. Observations are reported to the nearest .1 inch and are typically measured on a snow board to prevent an overestimation of the measurement that would occur if the observation were taken on a soft surface, such as grass.

The measurement of snowfall poses a unique set of problems for observers, as the methodology used can significantly affect the results. For example, the time of the observation is critical. If an observation is taken several hours after precipitation has stopped, snow compaction and ablation will likely result in a lower snowfall measurement. Kunkel et al. (2007) discovered inhomogeneities, however, that are present in the United States cooperative observer network (COOP) snowfall record.

These inconsistencies are most likely attributable to undocumented changes in observational practices and techniques. These inhomogeneities introduce an added amount of variance and potential biases to the snowfall records that change over time. Despite this, however, Kunkel et al. believe that an accurate representation of the variability of snowfall derived from COOP data may still be possible when the data are averaged over large areas

The observed data for this dissertation are obtained, in part, from a 1° x 1° gridded climate data set of North America developed by D. Robinson (Rutgers) and T. Mote (U. Georgia). These data, referred to as the NorAm data in this dissertation, are based on daily cooperative observations from both the United States (National Climatic Data Center, TD-3200, http://www.ncdc.noaa.gov/) and from the Meteorological Service of Canada (Bratten 1996). The United States Cooperative Observer Network was established by Congress in 1890 as part of the original United States Weather Bureau (National Weather Service 2008, http://www.weather.gov/om/coop/). While data exist from the inception of the observer program, the number of stations increased in 1948. (Dyer and Mote 2006).

The individual daily observations of snowfall form the basis of the gridded analysis. The bounded region of the gridded analysis is located within 53° - 168°W longitude and 20° - 71°N latitude and the data span the years 1951 to 1999. This time period was chosen to overlap with the availability of model data used in this study. The gridded products were created using an inverse-distance interpolation algorithm to create a Cartesian grid that was then projected onto a spherical surface (Willmott et al. 1984). First, 0.25° x 0.25° grids were created using a minimum of five observations and a maximum of 25 observations. If less than five observations were found within a radius of 100 km from a grid center, then only those available observations were used. If no observations were found within 100 km, the search radius was then set to the closest observation. These exceptions were applied mostly to northern Canada, where the density of stations decreases considerably relative to the other parts of North America. Once established, the 0.25° x 0.25° grids were converted to the final 1° x 1° grid by assigning the average value of four grid cells to their intersection point. The data were subject to quality control before individual observations were included in the regridding process. Individual observations that were inconsistent with the range of daily climate extremes for the station's state or province were removed from the analysis (Robinson 1989). Additional information on the creation of the grids is documented in Dyer and Mote (2006, 2007).

The gridded nature of the NorAm data is its primary strength. Previous studies of snowfall were often limited by examining observations from a small number of stations or by concentrating on a smaller geographical area. Use of the NorAm data allows for an in-depth, continental-scale examination of snowfall in North America that incorporates almost all the cooperative observations. A second advantage of this data is that the raw cooperative observations have been subject to a rigorous quality control process and the likelihood of error has been reduced.

The daily grids of snowfall were aggregated to produce both a monthly and an annual climatology of snowfall. Figure 2.1a illustrates the North American climatology of annual snowfall (cm) from 1951 to 1999. Annual snowfall in North America is characterized by a general increase in snowfall with latitude. Localized higher amounts of snowfall (200 to 400 cm/year) are found in eastern Québec and are associated with the prevailing North American winter storm tracks. Higher amounts are found along the Rocky Mountains in the western part of the continent, since the Pacific Ocean provides ample moisture to mid-latitude storm systems. Colder temperatures at these higher elevations also favor snow versus rain. Similarly, higher snowfall amounts are also located along the southern shore of Alaska.

The data, however, are not without limitation. The locations at which the cooperative observations are taken are primarily in regions of relatively high population density. As a result, certain biases are present in the data. First, the density of stations decreases dramatically north of the United States-Canada border - especially across the Yukon Territory. The gridded analysis of snowfall across northern Canada is based upon a fewer number of stations than those in the remainder of the North American continent. Secondly, cooperative observations are biased toward lower elevations. Larger snowfall totals that are typically found at higher elevations may be systematically underrepresented by the cooperative observer network. Thus the NorAm data are likely to underestimate snowfall in the higher elevations of western North America, since snowfall typically increases with altitude.

2.2. Derived Snowfall Data

Gridded snowfall observations, such as the NorAm data, are not readily available for other locations in the Northern Hemisphere. A technique that has been used to obtain estimates of snowfall in locations where observations are lacking is to develop a derived snowfall product by estimating the fraction of precipitation falling as snow based on monthly climatological temperatures.

One of the earliest attempts to relate monthly temperature to monthly accumulated snowfall was performed by Cehak-Trock (1958, interpreted by Legates 1990). Based on observations of temperature, precipitation, and snowfall from 23 stations across the tropical, mid-latitude, and arctic regions, the following relationship between monthly temperature (*T*) and the fraction of precipitation (*SF*) was found using a logistic curve fit:

$$SF = \frac{1}{1.0 + 1.61 \cdot (1.35)^{T}} \tag{2.1}$$

This relationship has been used in several studies of snowfall, including Legates (1990) and Rawlins et al. (2006). Legates (1990) evaluated the relationship for the entire Northern Hemisphere and found the mean absolute error of the relationship to be 0.06 mm/month.

This relationship was used to develop a derived snowfall product for the Northern Hemisphere from 1951 to 1999. Monthly grids of derived snowfall were produced from the University of Delaware Global Surface Air Temperature and Precipitation Climatology version 3.01 (Willmott and Matsuura 2002, http://climate.geog.udel.edu/~climate/, referred to as WM). A total of 7280 stations were used to produce time series of monthly and annual temperature, while 20,782 stations were used to create the precipitation fields. The actual number of stations used in a given year, however, varied substantially. Between 1,600 and 5,400 stations were used each year to create the temperature climatology, while 1,100 to 14,800 stations were used to create the precipitation climatology. The WM data were interpolated onto the native 0.5° grid of the climatology using a spherical distance weighting method (Shepard 1968). The original 0.5° data were regridded to a 1° x 1° grid and converted to Network Common Data Format (NetCDF). The regridded and packaged data were made available through the University of Washington's website: (Mitchell 2008, http://www.jisao.washington.edu/data_sets/willmott/).

Unlike the NorAm climatology, elevation was taken into account in the WM temperature data. Based on output from a digital elevation model (DEM), all temperatures were adjusted to sea-level using an environmental lapse rate of 6 °C/km prior to the spatial interpolation. After the interpolation to the 0.5° grid was complete, temperatures were adjusted back to the DEM elevation of the grid cell using the same environmental lapse rate. No elevation corrections were applied, however, for precipitation.

Climatology-Aided Interpolation (CAI) was also applied to the temperature data (Willmott and Robeson 1995). In this process, monthly temperatures at each station were differenced from their monthly climatology, and the spatial interpolation process was applied to the differences. The resulting 0.5° gridded field of differences was then added to the overall gridded climatology. Similar to the elevation corrections, the CAI procedure was applied only to the temperature data and not to the precipitation data.

The 49-year climatology of the WM derived snowfall is shown in Figures 2.1b and 2.2b The WM climatology exhibits a latitudinal gradient of annual snowfall along the eastern coast of North America, with a maximum of approximately 400 cm/year in eastern Québec. Local maxima of snowfall are also found across the Western United States and Canada, where values are similar in magnitude to those seen in the NorAm climatology.

The WM data also show other significant features of the Northern Hemisphere climatology of snowfall. Comparison with data published in the World Survey of Climatology (Wallen et al. 1970; Wallen et al. 1977; Arakawa et al. 1969) is made for the WM data. Across Europe and Asia, snowfall is well-represented. Local maxima exist in the mountainous regions of the Pyrenees, Central Alps, and Scandinavian Peninsula. The WM data also exhibit a minimum of snowfall in northwestern China. This lack of snowfall is consistent with the findings of Watts (1969), who finds that low amounts of snowfall are the result of low amounts of precipitation in this region.

Using the same relationship to determine the fraction of precipitation falling as snow (eq. 2.1), Rawlins et al. (2006) developed a derived snowfall product based on precipitation data for the former Soviet Union (Figure 2.2a). These precipitation measurements are higher resolution than the 0.5° WM precipitation climatology, and have been subject to strict observational guidelines and quality control. Using a derived snowfall product based on these precipitation measurements augments the hemispheric climatology based on the WM data.

Rawlins et al. (2006) also derived snowfall data that are based on observed daily and sub-daily precipitation measurements released by the All-Russian Research Institute for Hydrometeorological Information-World Data Center of the Federal Service for Hydrometeorology and Environmental Monitoring in Obninsk, Russia. These observations are made available through dataset TD-9813 produced by the National Climatic Data Center in Asheville, NC (http://www.ncdc.noaa.gov/). The TD-9813 data consist of precipitation measurements from 2188 stations located in the former Soviet Union. The data for some stations extend as far back as October, 1874. Precipitation measurements for some stations were missing following the dissolution of the Soviet Union in 1991. The gaps in data for these stations were filled using observations taken from the Russian synoptic data archive.

The original sub-daily and daily precipitation measurements were quality controlled. The data were homogenized to account for changes in rain gauge type and in wetting biases using a technique developed by Groisman and Rankova (2001). Corrections were applied for wind errors in accordance with Bogdanova et al (2002). These measurements were gridded to the standard 25-km Equal Area Scalable Earth (EASE) stereographic grid projection for the Northern Hemisphere using an inverse-distance weighted interpolation method by Rawlins et al. (2006). However, only gridded data that lie in the original boundaries of the TD-9813 dataset (54.10ºE to 190.17º; 35.28ºN to 80.60ºN) were used for analysis in this study. The gridded product consists of monthly accumulated precipitation totals that are based on the original observations. Monthly snowfall was then derived by estimating the fraction of precipitation that fell as snow using Eq. 2.1. This estimation also used the 0.5º x 0.5º resolution WM monthly temperature data that were sampled onto each EASE-grid location.

Rawlins et al. (2006) compared the TD-9813 precipitation with two global precipitation climatology datasets – the WM climatology and the CRU v.2.0 data set (Mitchell et al. 2004). When annual rainfall is compared among the three datasets, TD-9813 has a slight bias toward more rainfall. Therefore, it is possible that a similar bias exists for the derived snowfall product. Despite this inconsistency, however, Rawlins et al. (2006) demonstrate that the magnitude of the year to year variability of the rainfall time series is comparable among the three data sets. Each of the three data sets shows similar decreases in rainfall through the late 20th century.

The climatology of annual derived snowfall (cm/year) for the TD-9813 derived snowfall is shown in Figure 2.2a. Although extratropical storms influence weather in Siberia during the winter months, this area is relatively far removed from large sources of moisture. As a result, most locations in Siberia receive less than 200 cm/year of snowfall, and these values are consistent with qualitative accounts of Russian snowfall. In these locations, the derived snowfall is consistent with qualitative accounts of snowfall. The local maxima of annual snowfall are found in the Altai Mountains, which are located on the border of Mongolia, and also in the western part of the forested region, or taiga, that covers the Central Plateau (Danckwortt 1924).

2.3. Model Simulations of Snowfall

The use of models to simulate both past and future snowfall is a key component of this project. Models are useful in understanding what changes might have already taken place with respect to snowfall in locations where observations are either inadequate or are missing. Secondly, models are important in understanding how snowfall will change in the future.

The Fourth Assessment Report (AR4) of the Intergovernmental Panel on Climate Change (IPCC) incorporated results from various state-of-the-art climate models that were developed and run at various scientific institutions throughout the world. The models represent the efforts of thousands of individuals who contribute to the model components and who have conducted simulations of past and future climate. Results from the IPCC AR4 model simulations were archived by the Program for Climate Model Diagnostics and Inter-comparison (PCMDI) located at Lawrence Livermore National Laboratory in California. The results were made available by the World Climate Research Programme (WCRP) as part of the Third Coupled Model Inter-comparison Project, or CMIP3 (Meehl et al. 2007). Modeling centers that contributed simulations for the IPCC AR4 report also submitted results from 12 different experiments to the CMIP3 project. The details of all 12 experiments can be found in Meehl et al. (2007). In particular, the details of the two experiments used in this dissertation are quoted below:

1. Twentieth-century simulation to year 2000 (preferably starting from pre-industrial conditions in the late 1800s) with anthropogenic and natural forcings as modeling groups deemed appropriate;
2. Climate change experiment: Twenty-first century climate change simulation with SRES A1B (medium forcing, i.e., CO_2 concentration of about 700 ppm by 2100) from 2000 to 2100.

Computational space and time limitations prevented the analysis of every model that contributed to the CMIP3 project. Therefore, a subset of the models included in the CMIP3 project was selected for this research. The selection of the models was based on the work of Reichler and Kim (2008), which systematically tested the models' performance. Reichler and Kim's methodology involved computing a skill index statistic that was based on the normalized errors of 14 different climate parameters verified with data from 1979 to 2000. It is important to note that snowfall was not one of the variables used to evaluate model performance in Reichler and Kim (2008). The objective was, however, to focus on the best models overall and to examine how these models perform in simulating snowfall. The thirteen models that performed better than average in Reichler and Kim's analysis were selected for this study. These models are listed in Table 2.1, along with their spatial resolution, number of simulations for each experiment, and literature reference.

Each of the 13 models used in this project has monthly data available through CMIP3. Each of the model simulations was regridded to a uniform 1º by 1º latitude-longitude grid to facilitate comparisons among the models and the observed data. Many of the modeling centers performed multiple simulations for each of the experiments. There were a total of 41 individual simulations for the climate of the 20th century and 32 simulations for the SRES A1B scenario. Annual time series of the climate variables used in this study were created from the original monthly data.

Coupled atmosphere-ocean general circulation models represent the physical processes of the climate system in mathematical form. The major processes in the atmosphere, ocean, cryosphere, and land surface are represented by these equations. The evolution of the climate system over time can be accomplished through the numerical integration of these equations. Climate models have increased in complexity between the third and fourth assessment reports of the IPCC. This complexity is the product of both added knowledge of the climate system and the ability for faster computing. Improvements in bulk microphysical schemes in AOGCMs have resulted in changes in the way precipitation type is determined by the models. Precipitation type in AOGCMs has long been determined using a diagnostic scheme based on temperature. If the temperature in the model layers closest to the surface was below freezing, any precipitation that fell was characterized as snow. With improved resolution and more sophisticated parameterizations, some of the models used in this dissertation are capable of representing mixed phase (ice, graupel, water) clouds in their microphysical schemes. In these schemes, precipitation type is determined at the time of condensation in the cloud. Snow, ice, and rain are represented as individual mass fluxes in a vertical grid cell column. If falling snow and ice encounter a layer of above-freezing temperatures, then the mass is converted from frozen to liquid precipitation. Table 2.1 also outlines the process used in determining precipitation type for each of the 13 models.

Model resolution has also improved in recent years. Improvements in computing speed allow model developers to refine the spatial resolution of their models. Some modeling centers have increased the horizontal resolution of their models by a factor of two since the completion of the IPCC Third Assessment Report (TAR). This improvement, although beneficial, is still inadequate in representing some features that are important for snowfall. Mesoscale precipitation processes are still poorly represented in models, and features such as lake-effect snow in North America are on a scale that is too small to be simulated by the models.

Figures 2.1c and 2.2c illustrate the climatology of annual snowfall across the Northern Hemisphere for the multi-model ensemble. Snowfall increases with latitude in the Northern Hemisphere and is heaviest across western North America, northern Québec, parts of Russia, and Japan. Snowfall is enhanced by orography and is heaviest in locations that feature cold temperatures and a nearby source of moisture. Snowfall totals are found to be relatively small across the southeastern United States and also in northwestern China. A more detailed evaluation of the model-simulated climatology of snowfall relative to the observations is discussed in the next section.

2.4. Evaluation of Model Simulated Snowfall

Snow and precipitation processes that are represented in today's climate models have been developed and based on observations and physical understanding of the Earth's climate system. The most effective way to test their performance relative to the simulation of snowfall, however, is to focus on how well the models simulate the snowfall of the present climate.

When Figures 2.1a and 2.1b are compared to Figure 2.1c, there is a positive snowfall bias in the multi-model ensemble relative to the observed data. The patterns of relative maxima and minima, however, are captured reasonably well. The magnitude of annual snowfall is in better agreement over the eastern half of the continent, while the largest absolute errors are found in the western part of the continent.

The multi-model ensemble climatology of snowfall over Eurasia (Figure 2.2c) differs, however, from the TD-9813 climatology (Figure 2.2a) and the WM climatology (Figure 2.2b). The snowfall maximum in the Central Plateau, however, is significantly under-simulated by the models by as much as 250 to 300 cm. In all of the southern mountain ranges (e.g., Alatai, Tien Shan, Sayan, Aldan), snow simulation is greatly enhanced in the multi-model ensemble. Snowfall is also under-simulated in the models across central Europe, while it is overestimated across the Himalaya Mountains and Southeast Asia.

A more qualitative assessment of model performance is also conducted. Comparison with data published in the World Survey of Climatology (Wallen et al. 1970; Wallen et al. 1977; Arakawa et al. 1969) does not reveal any significant discrepancies between the multi-model ensemble and the descriptive accounts. The general latitudinal gradient of snowfall is captured by the models and the locations of maxima and minima in annual snowfall agree with the qualitative accounts. Local maxima exist near the Pyrenees, Central Alps, and across the Scandinavian Peninsula. While most of the snowfall accounts discuss the frequency of snowfall rather than accumulations, Gazzolo and Pinna (1973, reprinted in Cantu 1977) provides a detailed summary of annual snowfall for Italy. Italy typically receives 50 cm year^{-1} with some of the higher mountains in the northern part of the country receiving as much as 600 cm. While these extremes are missing from the multi-model ensemble, the simulations do capture the latitudinal gradient of snowfall in Italy. There is evidence that the elevation errors found for North America and Russia might also be present when examining snowfall in the European mountain ranges (i.e., the Swiss Alps, the Pyrenees Mountains, and the Scandinavian mountains).

Taylor diagrams of late-twentieth century snowfall were made to compare the model-simulated climatology of snowfall to the observed data. Taylor diagrams illustrate how well individual spatial patterns match an observed pattern. These diagrams are based on the relationship between the correlation coefficient, RMS difference, and the variances of two fields. For a complete discussion of the relationship between these three quantities, the reader is referred to the original paper by Taylor (2001).

Correlation between a model-simulated field of snowfall and the observed field is represented by the azimuthal location of a marker on the plot. Simulations that exhibit high pattern correlations are located closest to the x-axis, while low pattern correlations will be closer to the y-axis. The spatial variance is expressed in these figures as the normalized standard deviation, or the standard deviation of the simulated field divided by the standard deviation of the observed field. The difference between the normalized standard deviation of the simulated field compared to the observed field is expressed as a radial distance from the origin. In summary, the simulations that agree most with the observed field lie closest to the x-axis and have a radial distance closest to a normalized standard deviation of 1.

In Figure 2.5, each of the 41 simulations performed with 13 models is compared to the NorAm snowfall climatology. The Taylor diagrams demonstrate model performance of annual snowfall from 1951 to 1999 for the geographical region east of the Rocky Mountains (105° W - 50° W, 25° N – 72° N), relative to the NorAm gridded observations. In this region, pattern correlations between the models and observed data are strong, and the models exhibit spatial variances that agree well with the observed data. When the entire continent is examined (Figure 2.6, 165° W – 50° W, 25° N – 72° N), the pattern correlations are not as strong as in Figure 2.5 and the spatial variance is also greater.

Taylor diagrams show each of the 41 simulations compared to the TD-9813 derived snowfall data (Figure 2.7). The resulting pattern correlations and normalized standard deviations show less agreement than in North America. There are several reasons why there are differences between the models and TD-9813 data. First, the maximum of snowfall near the western Central Plateau of Siberia is slightly displaced to the south in the models. The coarse representation of terrain in the models tends to flatten and broaden the area of the Central Plateau. Secondly, the original TD-9813 precipitation measurements may be subject to a similar low-elevation bias that is present in the NorAm data. The difference between the low-elevation bias in the TD-9813 derived snowfall and the NorAm data, however, is that only the precipitation measurements are biased; the TD-9813 derived snowfall is based on the WM temperatures that are corrected for elevation.

The Taylor diagram for the models versus the WM snowfall climatology is shown in Figure 2.8. Across the entire Northern Hemisphere, most of the models show similar pattern correlations (0.65 to 0.85). The models disagree, however, on their simulations of the observed variance in snowfall. The two versions of the Canadian model (CGCM T47, CGCM T63) produced simulations that performed the best relative to the WM data. It is also important to note that model performance does not appear to be associated with either model resolution or the method by which precipitation type is determined.

Comparison of the maps and Taylor diagrams of the 49-year snowfall climatology of both observations and the multi-model ensemble indicate that elevation is important in assessing model performance in North America. Better agreement between the multi-model ensemble and the observed data in the eastern half of North America can be attributed to the relatively flat terrain and a higher density of stations to capture the regional variations of snowfall. The larger errors in the West are consistent with the coarse representation of the Rocky Mountains in the models as well as with biases that are present in the observational data.

The fidelity of the snowfall simulations can also be assessed by computing area mean snowfall. Table 2.2 lists the area mean annual snowfall for North America, the TD-9813 domain, Eurasia, and all land areas of the Northern Hemisphere for each of the three observational data sources and all 13 CMIP3 models. The mean annual snowfall was based on the period of years from 1951 to 1999. For North America, Eurasia, and the Northern Hemisphere, all 13 models overestimate snowfall relative the observed data. Most models also overestimate snowfall relative to the TD-9813 data, although two models slightly underestimate the spatially averaged snowfall (CGCM 3.1 T47, and CGCM 3.1 T63). Overall, these two models have smallest absolute differences with the observed data sources.

One concern associated with using observational data for snowfall, however, is that the data are biased toward lower elevations. Increased numbers of stations situated in valleys as opposed to mountain peaks lead to the prevalence of warmer temperatures in the climatological record, and, thus, a reduction in the fraction of precipitation falling as snow. As a result, snowfall is likely to be underestimated by the observational record. In order to demonstrate this elevation bias in the NorAm data, comparisons were made with data that have been corrected for elevation. The NorAm record includes temperature data, which have also been derived from monthly cooperative observations. The NorAm temperature data were subject to the same quality-control and gridding procedures that were previously discussed for snowfall. The NorAm temperature data were compared with data produced by the PRISM Climate Mapping Program at Oregon State University (Daly et al. 1994). PRISM, or the Parameter-elevation Regressions on Independent Slopes Model, uses single-point observations along with elevation data and other spatial data sets to create high-resolution gridded estimates of climate data for maximum temperature and minimum temperature – which are averaged together to produce a monthly mean temperature. The higher resolution data capture small-scale features such as elevation impacts, land-sea interactions, and night inversions.

The high-resolution PRISM data (~0.08° latitude x ~0.08º longitude) were regridded to the same 1° x 1° grid of the NorAm data for comparison. Since the PRISM data and NorAm data span different time periods, only the time interval of overlap was used (1971 to 1999). The results of the temperature comparison are shown in Figures 2.9. The NorAm data agree with the PRISM data in the eastern half of North America, while the discrepancies are largest in the mountainous terrain of western North America. It is likely, therefore, that snowfall observations also suffer from this elevation bias. Results obtained from studying snowfall in the western part of continent with the NorAm data must be interpreted with this bias. Any future gridded analyses should apply corrections to account for elevation.

Elevation errors that are present in the observational record explain only part of the difference between the observed climatology of snowfall and the model simulations. When the mean annual temperature from the multi-model ensemble is differenced with the temperature from the NorAm data, large biases are present in the mountainous regions. A secondary part of the error is the result of the poor model topography. This limitation is particularly important when examining snowfall, since the occurrence of snowfall is dependent upon temperature, which is influenced by elevation. The horizontal resolution of the climate models is on the order of 10^2 km, making the models incapable of resolving individual mountain ranges. Figure 2.9 illustrates the mean topography of the models versus USGS topography data, originally at 0.2° resolution, regridded to match the 1° resolution of the multi-model ensemble. In the models, the mountains are more broad and flat than in actuality, and individual mountain ranges such as the Appalachians, Sierra Nevada, Cascades, and the Front Range of the Rocky Mountains are not easily visible. Additionally, the presence of the Great Basin is missing from the model topography.

The difference between the USGS topography and the mean elevation of the models for the TD-9813 region is shown in Figure 2.9. In the southern mountains, model representation is more broad and flat when compared to actual topography. The finding that the number of higher-altitude locations is larger in the models can explain the positive snowfall bias in the multi-model ensemble. Elevation errors are smaller for the Central Plateau, where snowfall is significantly underestimated in the models. The model representation of topography, however, shows a lower and less expansive Central Plateau than is seen in the USGS topographic data.

Chapter 3

Late Twentieth Century Variations in Northern Hemisphere Snowfall

3.1. Motivations and Overview

In the review of relevant literature presented in Chapter 1, it was noted that the IPCC Fourth Assessment Report has no direct discussion of snowfall changes. The discussion is focused on snow cover, and it is implied that the changes in snowfall would be similar. The report states that midwinter snow cover across the Northern Hemisphere has changed little since the 1920s, although there has been increased variability in the fall and spring transition seasons (Lemke et al. 2007). Many of the previous studies of historical snowfall changes are also limited by either geographical extent or the availability of data. An in-depth examination of late-twentieth century snowfall changes across the Northern Hemisphere is reported in this chapter.

Scott and Kaiser (2004) examined cooperative snowfall observations across the United States from 1948 to 2001. The study found that the snowfall season has decreased in length by as much as 6 days per decade across a large area extending from the Pacific Northwest, Kansas, the Ohio Valley, and the Northeastern United States. North of this area, however, annual snowfall was shown to increase, especially near Colorado, Wisconsin, and the lee of the Great Lakes. Similarly, Karl et al. (1992) found that the area-averaged snowfall increased at a rate of 8.8 cm decade^{-1} in northern Canada between 55 and 80 °N latitude. In southern Canada, however, snowfall was found to decrease at a rate of 0.65 cm decade^{-1}.

In other parts of the Northern Hemisphere, snowfall was also examined. Laternser and Schneebeli (2003) studied snow cover, snow depth, and snow days in the Swiss Alps from 1931 to 1999. These three variables were found to have increased slightly between 1930 and 1980, before decreasing significantly between 1980 and 1999. In Greenland, Thomas et al. (2006) found that the thickness of the Greenland ice sheet increased at high elevations. These results are indicative of snowfall increases across central Greenland.

The data used in this chapter to examine snowfall are the same as those used in the previous chapter. Observational data, derived snowfall products, and model simulations are used to address two scientific objectives. First, the nature of the trends in late-twentieth century snowfall is determined and the responsible mechanisms for these changes are identified. Second, the ability of the CMIP3 models to simulate the trends in late-twentieth century snowfall is assessed. Understanding model performance in simulating the late-twentieth century trends in snowfall is important in interpreting the results obtained through simulations of the future climate.

3.2. Grid Point Analysis of Annual Trends

The purpose of this analysis is to identify any large scale patterns of snowfall changes during the last half of the twentieth century. Trends in annual snowfall are computed for the period 1951 to 1999. Trends in annual snowfall are examined on a grid point basis for the observed and simulated data used in this study. The starting year is chosen to coincide with the increased reliability in the NorAm and TD-9813 data. The ending year is selected to include all of the CMIP3 models that are used in this dissertation. December 1999 is the last month that appears in all 41 historical climate simulations, since the dividing point between historical and future radiative forcing varies among the models.

The nature of snowfall data, like other climate variables, often complicates the traditional t-test method of determining statistical significance. First, snowfall data exhibit autocorrelation. As a result, the individual observations in a time series are not independent, thus the actual number of degrees of freedom that should be used to determine statistical significance is less than *n-2*. Secondly, the snowfall data are not normally distributed since negative values of snowfall are not physically possible. As a result, snowfall observations fail the assumptions of the classical t-test that the data are independent of each other and that the data are normally distributed.

In order to determine statistical significance of the grid point trends in annual snowfall, a moving block bootstrapping technique (Wilks 1997) is used. In this method, a linear least-squares regression is computed for a given time series and the residuals are computed between the observed data and the linear fit. These residuals are then randomly resampled back onto the linear fit to create a synthetic time series. The residuals are resampled onto the linear fit in "blocks" to account for autocorrelation. A block length of four years is used throughout this study. The four-year block length is selected to ensure that any year-to-year dependence of snowfall is taken into account. This resampling is used to create 1,000 synthetic time series. Linear-least squares regression is performed on the 1,000 synthetic time series to create a distribution of plausible trends. The original least-squares regression of the original time series is deemed statistically significant at the 95% confidence interval if the zero trend falls within the upper or lower 2.5% of this distribution.

The bootstrap method of determining statistical significance avoids the assumptions of normality and independence that must be made for a traditional t-test. The block approach to resampling the residuals to create the synthetic time series preserves any autocorrelation that is present in the data. The Monte Carlo aspect of this method is preferred, as the creation of 1,000 plausible trends provides an estimate of a probability density function that is based on the data. The method described is used for determining statistical significance of a single time series. When this method is applied on a grid point basis, an additional stipulation is followed. The same time indices are used to create the synthetic time series at each grid point. Therefore, the set of plausible trends for the gridded data preserves the spatial coherence of the data.

The trends in annual snowfall for the NorAm data are shown in Figure 3.1. Only statistically significant trends ($p=0.05$) are shaded. The most notable features are increases in snowfall at higher latitudes and decreases in snowfall in the midlatitudes and in the western part of the continent. The largest decreases are found along the Rocky Mountains and central Ontario. Increases in snowfall are found, however, across the northern high plains of the United States and in the vicinity of the Great Lakes. Trends in the southeastern United States are weak. Across Alaska, there is an area of positive trends, bounded by negative trends on either side.

Trends in annual snowfall are also computed for the TD-9813 derived snowfall data (Figure 3.2). The derived snowfall product allows for the examination of snowfall trends across the former Soviet Union during the late-twentieth century. Trends across this region are either weak or negative, with the exception of a few locations. Most of the negative trends are statistically significant, especially along the southern mountains that border Kazakhstan and Mongolia, and near the northern coast of the Arctic Ocean. Positive trends are found in the western Siberian lowlands to the south of the Kara Sea. Positive trends are also found in easternmost Siberia bordering the Bering Sea.

Grid point trend analysis is also performed for the WM derived annual snowfall (Figure 3.3). The examination of snowfall with this product allows for the analysis of trends across the entire Northern Hemisphere. When the trends from WM data are compared to the NorAm data, many similarities are noted. Strong positive trends are found at higher latitudes and across northern Québec in both the WM and the NorAm data. Trends across Alaska are consistent with the NorAm data, and increases in snowfall are also found near the Great Lakes. Negative trends that extend westward to British Columbia are also found in Ontario. The negative trends in the Rocky Mountains are weaker, however, in the WM data when compared to the NorAm data. In some locations in the Rocky Mountains south of 50° N latitude, trends are near zero or even slightly positive. This difference may result from elevation corrections applied to the WM data, but not to the NorAm data. Trends across Alaska are consistent with the NorAm data, and increases in snowfall are also found near the Great Lakes. The WM trends are also largely in agreement with trends in the TD-9813 data. The most notable difference between the two data sets is that the area of positive trends in northern Siberia is slightly larger in the WM data than in the TD-9813 data.

Across the remainder of the Northern Hemisphere, positive trends are found in the WM data over Greenland, with the exception of the extreme southern tip. Positive trends are also found in northern Scandinavia and central Siberia, including the Taimyr Peninsula. Negative trends dominate much of central Europe, extreme eastern Russia, and Japan.

One of the primary objectives of this chapter is to evaluate the ability of the CMIP3 models used in this study to reproduce the observed trends in snowfall for the late twentieth century. Trends in annual snowfall are computed for the multi-model ensemble mean of the 41 individual climate simulations, and the results are compared to the trends derived from the observational data. While the other three data sources represent a single realization of the climate of the late twentieth century, the multi-model ensemble is an average of 41 individual simulations of the historical climate. Examining a multi-model ensemble mean allows the internal variability of climate system to be averaged out, thus highlighting the forced trends in annual snowfall. The shaded areas in Figure 3.4 indicate locations in which the trends in the multi-model ensemble are stronger than ±1 cm decade^{-1}. Decreases in snowfall are noted across western North America and across the northeastern United States and eastern Canada. Decreases are also found across much of central Europe, the Tibetan mountains, and Japan.

The robustness of trends in the CMIP3 models was also examined. Figures 3.5 and 3.6 show the percentage of model simulations that have statistically significant trends. The objective of this analysis is to gain insight into how the individual model simulations reproduce the observed trends of the late twentieth century. In this analysis, the computational cost of performing the bootstrap method of statistical significance for all 41 CMIP3 model simulations was too great. Therefore, a traditional t-test was used to assess statistical significance in the simulations.

Figures 3.5 and 3.6 illustrate that in locations where trends in the multi-model ensemble mean are largest, the percentage of model simulations that show statistically significant trends is greatest. Even in locations where the multi-model ensemble trends are smaller, Figures 3.5 and 3.6 provide additional information. In extreme northern Québec, for example, trends are weak in the multi-model ensemble, but are positive in the NorAm and WM data. Figure 3.6 indicates, however, that at least some model simulations do, in fact, produce statistically significant positive trends in this region.

To further understand the extent to which the models and observations agree in simulating the late twentieth century variations in snowfall, area-mean annual snowfall is shown in Figure 3.7 for North America (A), the TD-9813 domain (B), all of Eurasia (C), and the entire land area of the Northern Hemisphere (D). In Figure 3.7, the anomalies relative to the 1951-1999 mean annual snowfall are shown. For 3 of the 4 areas, observations and model simulations show highly variable time series of annual snowfall, with a slight indication of a downward trend. The exception is over the TD-9813 domain, where snowfall increases slightly after 1985. This result compares well with Figures 3.3 and 3.4, which also show weak positive trends in annual snowfall across Russia. Across the entire Northern Hemisphere (Figure 3.7, D), annual snowfall decreases slightly despite increases at high latitudes.

3.3. Regional Analysis of Trends in Annual Snowfall

The primary motivation for this analysis is to compare the observed and simulated trends across the Northern Hemisphere. As previously discussed in greater detail in Chapter 2, the AOGCMs used in this study lack the resolution to represent all of the mesoscale features responsible for the climatology of snowfall in a given location on the grid point scale. The assumption is made that models perform better when several grid points are averaged together to produce a regional time series of snowfall.

The Northern Hemisphere is divided into 20 regions for this analysis. The selection of these regions is somewhat subjective. Regions are selected based on several factors, including population, economic importance, or indications that changes in snowfall had already occurred or are projected to occur in the future. The 20 regions are shown graphically in Figure 3.8. Table 3.2 describes the geographical boundaries of the regions.

The regions are defined on a uniform 1° x 1° latitude-longitude grid. The observed and simulated snowfall data are regridded to the uniform grid before they are averaged over a given region. Only model grid boxes which contain a minimum of 50% land area are included in the regional averaging process. Spatial averaging is performed on each of the regions to create a single annual time series of snowfall for each of the regions. This method is performed for all 41 model simulations and for the observed data. Table 3.2 also outlines the source of the observational data used for each of the regions.

Trends in annual snowfall for these regions are computed using an ordinary least-squares regression that is consistent with the methods described in section 3.2. The horizontal solid lines in Figures 3.9 through 3.28 represent the regression coefficient for the observed trends in each region and the 95% confidence intervals are denoted by the horizontal dotted lines. The markers indicate the ordinary least-squares regression coefficients for the 41 model simulations and the whiskers indicate the full range of 95% confidence intervals from all simulations performed with a given model.

The trends in regional snowfall are consistent with the spatial patterns described in the previous section. Regions 8 and 9 show statistically significant negative trends in the observed data (Figures 3.16 and 3.17). These regions are located along the western coast of North America where robust negative trends are also found in the models. Similarly, regions 12 and 14 show statistically significant decreases in annual snowfall near Japan and Turkey, respectively. The positive trend across Scandinavia (region 18, Figure 3.26), however, is shown to be significant. One of the reasons for the absence of significant trends across the remainder of the Northern Hemisphere is that the regional time series of annual snowfall exhibit a large degree of inter-annual variability. The weak climate change signal during this period is also a contributing factor.

A second objective of this analysis is to compare the model-simulated trends with the observed trends. While the amount of overlap between the observed and simulated trend confidence intervals provides some indication of the amount of consistency between the two, Lanzante (2005) cautions against the direct comparison of two sets of error bars. In order to prevent erroneous conclusions about the consistency between the modeled and observed snowfall trends, a t-test is employed to determine if there is a statistically significant difference between the two regression coefficients.

The test to compare two slopes is described in Zar (1974). In this test, the null hypothesis states that the slopes are the same. The t-statistic is described as:

$$t = \frac{b_1 - b_2}{s_{b_1 - b_2}} \tag{3.4}$$

where

$$S_{b_1-b_2} = \sqrt{\frac{(S_{Y\cdot X}^2)_p}{(\sum x^2)_1} + \frac{(S_{Y\cdot X}^2)_p}{(\sum x^2)_2}} \tag{3.5}$$

and

$$(S_{Y\cdot X}^2)_p = \frac{(\text{residual SS})_1 + (\text{residual SS})_2}{(\text{residual DF})_1 + (\text{residual DF})_2} \tag{3.6}$$

The critical value for t should be based on a two-tailed distribution and $n_1 + n_2 - 4$ degrees of freedom. The shape of the markers for the model simulations in Figures 3.8 to 3.28 signifies the results of the test. A star indicates that the modeled and simulated trends are statistically the same at the 95% confidence interval, while an open circle indicates that the trends are statistically different.

The results of this analysis indicate that for most of the regions in the Northern Hemisphere, the modeled and observed trends are generally consistent. The largest discrepancies are found in Québec (regions 5 and 6) and British Columbia (regions 9 and 10). In these regions, a majority of the modeled trends are found to be statistically different from the observed trends. The reasoning for this difference may be explained by the nature of snowfall in these regions. While some areas are subject to inconsistencies in trends, in general the CMIP3 models adequately represent the trends in annual snowfall across the remainder of the Northern Hemisphere during the last 50 years of the twentieth century.

3.4. Discussion of Analysis Results

A central theme of this project is that temperature and precipitation are primarily responsible for the occurrence of snowfall in a particular region. The air must be sufficiently cold and precipitation must take place in order for snowfall to occur. Changes in both temperature and precipitation strongly influence changes in snowfall.

Trends in temperature and precipitation are qualitatively assessed based on the work presented in the Chapter 3 of the IPCC Fourth Assessment Report (Trenberth et al, 2007). Trends in annual temperature are positive at nearly every location in the Northern Hemisphere during the 49-year period of this analysis. The spatial pattern of trends in annual temperature is characterized by the occurrence of the strongest positive temperature trends at higher latitudes. Moderate trends in temperature are found in western North America and these areas coincide with the areas of the largest decreases in snowfall. Moderate increases in temperature are also found in central Europe and slightly stronger temperature increases are found near the Himalaya Mountains. In both these locations, decreases in annual snowfall are found in the multi-model ensemble. Weaker positive trends in temperature are found across the southeastern United States.

Annual trends in precipitation are far less uniform. A general pattern of negative trends at lower latitudes and positive trends at higher latitudes emerges, but this pattern is less uniform than the pattern found for snowfall. Increases in annual precipitation are found across Alaska, northern Québec, Greenland, Scandinavia, and much of Eurasia. In these locations, temperatures are typically well below freezing during the winter and the positive increases in precipitation coincide with the positive trends in annual snowfall. Positive trends in precipitation, however, are found near the Himalaya Mountains where annual snowfall demonstrates a negative trend. Decreases in precipitation are found across western North America and also in southern Europe. Coupled with warming in these two regions, decreased precipitation helps to explain the negative trends in snowfall.

This analysis during the late twentieth century shows that changes in temperature and precipitation influence trends in snowfall amounts. Increasing temperatures reduce the fraction of precipitation that falls as snow. Changes in precipitation, however, have their greatest influence at high latitudes and at high elevations. In these locations, the temperature is typically below freezing and even modest increases in temperature are not enough to alter the fraction of precipitation that falls as snow. Therefore, either increases or decreases in annual precipitation translate almost directly into increases or decreases in snowfall. Since precipitation trends at high latitudes are mostly positive, positive trends in snowfall are found in these locations. It is also likely that changes in temperature and precipitation have opposite effects in determining snowfall trends.

The trends in both temperature and precipitation are weak for the late twentieth century relative to the trends that are expected to occur in the twenty-first century. The weaker trends are presumably the result of relatively weak net forcing of the climate system during this period (1951 to 1999). The forcing from greenhouse gases emitted into the atmosphere during this time period is distinguishable from background internal variability of the climate system, but trends in late-twentieth century snowfall are due in part to internal climate variability.

Grid point analysis of the NorAm snowfall in Section 3.2 shows positive trends in snowfall since 1950 in the vicinity of the Laurentian Great Lakes. Several studies have examined recent increases in lake-effect snow. It is believed that the combination of both warmer lake temperatures and a shorter period throughout the winter when the lakes remain frozen contributes to the increases. This feature is absent from the multi-model analysis of snowfall. The typical grid spacing of an AOGCM is approximately 2° latitude by 2° longitude. At this resolution, global climate models are incapable of representing mesoscale processes related to lake-effect snow. The lakes themselves are absent from the models and the physical interaction between the air and the water cannot be resolved. Although the hypothesis of warmer water temperatures contributing to enhanced lake-effect snow cannot be tested with global models, this mechanism can be tested by using Hudson Bay as an analog.

Most snowfall that occurs along the lee side of Hudson Bay in western Québec occurs in the fall prior to the freezing of the bay. Polar air interacts with the warmer water to produce snowfall downwind of Hudson Bay. Once the water freezes, the moisture source for the snowfall is eliminated. This mechanism is analogous to the lake-effect snow that occurs off the Great Lakes. A recent paper by Gagnon and Gough (2005) reported that the freeze dates for Hudson Bay have been occurring later in the fall. This extended period of time in which the water remains unfrozen is indicative of the possibility of additional snowfall occurring as the atmosphere and the water have more time to interact in the fall season.

The inability of AOGCMs to simulate lake-effect snow underscores one of the greatest limitations of examining snowfall with global models. Not every region in the Northern Hemisphere has a snowfall climatology that is dominated by large-scale synoptic weather systems and modes of atmospheric variability. There are many locations where mesoscale processes – such as lake-effect snow or terrain enhancement – are responsible for a large fraction of the total annual snowfall. In these locations, AOGCMs prove to be inadequate in assessing both past and future changes in snowfall. A potential area of future work would be to develop regional climate models that are capable of resolving these mesoscale interactions that contribute to snowfall.

The Arctic Oscillation (AO) is the dominant mode of atmospheric variability in the winter across the Northern Hemisphere. The positive phase of the AO is characterized by decreased pressure in the Arctic and increased pressure in the mid-latitudes. The link between the AO and snowfall is important. The AO influences temperature and precipitation, as the mid-latitude storm tracks are related to the AO. For example, when the AO is in its positive mode, storm tracks across North America are shifted northward and much of the United States and Canada experience warmer temperatures. When the AO is negative, however, cold air is common across much of North America.

There is conflicting evidence, however, concerning the possibility that changes have already occurred to the AO as a result of increased greenhouse gases. Some researchers (e.g. Feldstein 2002) argue that the AO has trended toward its positive phase in the last 50 years. Miller et al. (2006) report, however, individual models vary significantly in reproducing this positive trend of the late 20th century. Furthermore, Joyce (2002) suggests that the sensitivity of winter precipitation to the AO across the United States has increased since the early twentieth century. A trend toward a more positive AO, coupled with an increased sensitivity to the index, may lead to decreased snowfall across central and eastern North America. Changes in circulation associated with the phase of the AO may be an important secondary factor in assessing past and future changes in snowfall across parts of the Northern Hemisphere. One potential area of future work would be to use regional climate models to examine mesoscale interactions as they relate to snowfall changes for both past and future climates.

Chapter 4

Future Changes in Northern Hemisphere Snowfall

4.1. Motivation

In the 21^{st} century, changes in many aspects of the climate system are expected to become more pronounced than in the 20^{th} century. The primary reason is that concentrations of anthropogenic greenhouse gases are expected to continue to increase over the coming century. This chapter addresses scientific questions and issues related to the nature and causes of future snowfall changes in the Northern Hemisphere. In particular, this chapter outlines the trends in snowfall for the 21^{st} century, examines the influence of temperature and precipitation changes on snowfall, and determines when such changes in snowfall may become apparent.

Knowles et al. (2006) hypothesized that future changes in snowfall are likely in the Western United States, based on trends and principles derived from the analysis of historical snowfall. This work was based on cooperative station data, which spanned the years 1948 to 2004. Knowles et al. found that the fraction of precipitation that fell as snow decreased as a result of warmer temperatures. The analysis found that snowfall decreased the most when the temperature changes over the period were slight to moderate (0 to +3ºC). The mean temperatures in these locations were close to the freezing point, and a small increase in temperature was enough to impact the fraction of precipitation that fell as snow. Knowles et al. (2006) also found that increases in liquid-equivalent snowfall occurred in locations where the mean minimum winter temperature (November through March) on precipitation-producing days was less than -5ºC. Knowles et al. (2006) hypothesize that midwinter warming will substantially affect snowfall that occurs in the Western United States during the 21st century. This decrease in snowfall will result in an increased risk for water shortages and flooding.

In the Northeastern United States, Hayhoe et al. (2007) examined the past and future changes in hydrological climate features based on simulations that were performed with nine of the CMIP3 models. Their results project a 50% (85%) decrease in area-averaged SWE relative to 1961– 1990 values by the year 2099 in the B1 (A1F1) climate change scenario. The results of this study also indicate a shortening of the snowfall season and a 25% (50%) reduction in the number of snow days for this scenario. The results also highlight a smaller reduction in the number of snow days across Pennsylvania and New Jersey. Hayhoe et al. (2007) argue that, at present, there are already a limited number of days that are sufficiently cold to support snowfall. In the future, the number of these days is expected to decrease, but the increases in precipitation raise the likelihood that snow will actually occur on colder days. More work is needed to understand and quantify snowfall changes in more locations in the Northern Hemisphere on regional scales. Additionally, the causes for such changes must be further understood.

The primary tools in studying future snowfall changes are coupled atmosphere-ocean climate models that are forced with time-varying concentrations of greenhouse gases and aerosols. The same set of 13 CMIP3 climate models identified in Chapter 3 is used, and simulations using the IPCC SRES A1B climate change scenario are analyzed. The A1B scenario is characterized by "rapid and successful economic development" as an emphasis is placed on "market oriented solutions" to problems and a "high consumption of resources" (IPCC 2004). Meehl et al. (2007) report that the mean increase in surface air temperature through 2099 for the A1B scenario is 2.7 °C, with CO_2 concentrations by the end of the 21st century that are in the middle of distribution of all of the SRES scenarios. The forcing to the climate system is midrange compared to the other scenarios, yet large enough so that statistically meaningful changes in snowfall are likely to be found.

Future trends in annual and seasonal snowfall are presented on a grid point basis in section 4.2. In sections 4.3 and 4.4, trends in monthly snowfall are examined, and the effects of changes in temperature and precipitation on snowfall are analyzed. Changes in the frequency of daily snowfall events are studied in section 4.5. In section 4.6, a signal-to-noise analysis is presented in order to understand when the changes in snowfall might become perceptible. In section 4.7, the results of the analysis of future snowfall are compared to the results presented in other papers that have studied snow cover and snow water equivalent.

4.2. Grid point Trends in Snowfall

Annual trends in snowfall are calculated for each model and for each grid point. The annual trends in snowfall for the 21st century shown in Figure 4.1 are for the multi-model ensemble. Trends in accumulated annual snowfall are calculated using the linear least-squares regression method described in section 3.2, and the results are presented in units of cm decade^{-1}. Small trend values between -1 and 1 cm decade^{-1} are not shaded in order to emphasize coherent patterns. In this figure, annual snowfall is projected to decrease in the 21st century across much of North America and Europe. Snowfall is projected to increase, however, at higher latitudes and across the interior parts of Greenland and Asia. The pattern of future snowfall change is also similar to the one for the 20th century that was presented in Chapter 3. Both patterns show a transition zone between positive and negative trends in snowfall.

The key result of a transition zone between positive and negative trends in annual snowfall can be explained for the same reasons discussed in Chapter 3. The transition zone occurs for two reasons:

1.) Precipitation type is insensitive to temperature at higher latitudes, as the air is sufficiently cold so that even a modest increase in temperature is unable to raise levels above freezing.
2.) Widespread precipitation increases at higher latitudes are projected to occur in the 21st century.

In locations poleward of the transition zone, any change in precipitation translates directly into a change in snowfall. Decreases in snowfall south of the transition zone are likely to be driven primarily by warmer temperatures, decreasing the fraction of precipitation falling as snow, with changes in precipitation being a secondary effect. These changes in temperature and precipitation will be further examined in Section 4.4.

Decreases in snowfall may significantly impact the local water resources in the Western United States. Annual snowfall is an integral part in determining the amount of spring runoff. In this region, some evidence is provided that changes in runoff resulting from decreased snowfall are already occurring. The future decreases in annual snowfall increase the likelihood that the impact on local water resources will become more apparent in the 21st century.

These results also fit into the broader context of high-latitude climate change. Substantial increases in snowfall are projected to occur over much of Greenland. The potential exists for increase in snowfall to provide a positive contribution to the mass of the Greenland ice sheet. The increased ice and snow may partially offset the substantial melt that is expected to occur during the 21st century (Thomas et al. 2006; Box et al. 2006). These results support previous findings related to the nature of the ice balance over Greenland and underscore the necessity of including snowfall changes in future studies of the Greenland ice sheet.

The trends in snowfall for the 21st century are projected to be stronger than the trends that were found for the late twentieth century. The stronger trends are a reflection of the expected larger forcing of the climate system in the 21st century. The magnitude of the changes in temperature and precipitation – the two main factors that determine snowfall – is greater in the 21st century. These findings suggest that any consequences that were experienced during the late 20th century as a result of changes in snowfall are likely to become more prominent in the future.

Snowfall is also examined on a seasonal basis. Seasonal trends in snowfall are examined from the multi-model ensemble for the fall (September, October, November), winter (December, January, February), and spring (March, April, May) seasons. Trends are computed using a least-squares regression, and the seasonal trends in snowfall are shown in Figures 4.2 – 4.4. These results show that substantial changes are projected to occur within the snowfall season. Fall trends are predominately negative across most of the Northern Hemisphere, with the exceptions of Greenland and extreme Northern Russia. In the winter season, however, positive trends are found in more locations – especially at higher latitudes. The most notable trend increases are simulated across Northern Québec, Greenland, and Central Russia. In the spring, snowfall decreases across much of the Northern Hemisphere.

In most locations throughout the Northern Hemisphere, temperatures during the transition seasons are warmer than the temperatures during the winter. It is much more likely, therefore, that increases in temperature will influence the fraction of precipitation falling as snow in the transition seasons to a greater degree than in the winter. Furthermore, increases in high latitude precipitation are greatest during the winter and have a lesser positive contribution during the remaining seasons.

These results indicate that total annual snowfall in some locations will not change, but the seasonal distribution of snowfall will be different. At this point, it is unclear if a redistribution of snowfall throughout the snowfall season will alter other features of the climate system. Potential influences that a temporal redistribution of snowfall might have on climate features, such as runoff and the long-term storage of soil moisture, should be studied further.

Temperatures from the multi-model ensemble of the late twentieth century climate simulations are also shown in Figures 4.2 to 4.4. The mean temperatures for the 30-year period from 1970 through 1999 are plotted. Comparison of temperatures with snowfall trends shows a relationship between the two. Positive trends in snowfall are found in locations and seasons where the average temperature is below -15 °C. Decreases in snowfall are found, however, when the average temperature is greater than -10 °C. This result implies that the transition zone between negative and positive trends is likely to occur somewhere between these two values. This relationship is valid across much of the Northern Hemisphere, with slight differences noted near the Tibetan Plateau where extreme elevation differences are an added factor.

The result that the transition zone between positive and negative trends is located between the -10 and -15 ºC isotherms of the late twentieth century climate cannot be directly compared with the results of Knowles et al. (2006). Primarily, Knowles et al (2006) compared late-twentieth century observations of temperature to late-twentieth century trends in snowfall on a regional scale in the Western United States. This dissertation, however, compares model simulations of the late-twentieth century temperatures to future trends in snowfall on the hemispheric scale. Secondly, this dissertation assumes a 10:1 snowfall-to-precipitation ratio, where as Knowles et al. (2006) used only precipitation liquid equivalents. There are some key points, however, that can be made when examining the two studies. Knowles et al. (2006) found that the transition zone between positive and negative trends in SFE occurred when the average minimum temperature on precipitation-producing days was -5 ºC. The reason why the results from Knowles et al. (2006) differ from the findings presented in this section is unclear. One possible explanation would be that the method of using minimum temperatures on precipitation days results in a warmer temperature than using the method of the seasonal average temperature. Precipitation tends to be associated with warm advection, and thus temperatures on precipitating days are likely to be warmer.

Räisänen (2008) also examined the relationship between temperature and the transition zone between positive and negative trends in SWE. Räisänen found that -20 ºC winter isotherm corresponds well with the transition zone, and this result is more consistent with the values obtained in this analysis. This slight difference between the results of this analysis and those of Räisänen is expected, since colder temperatures would be required to maintain the snow pack and lead to a positive trend in SWE.

4.3. Monthly Trends in Snowfall

To examine the projected changes in monthly snowfall, a method was employed to facilitate the decomposition of the snowfall change into the parts that are attributable to changes in temperature and precipitation (see section 4.4). The change in monthly snowfall is defined by:

$$\Delta S = S_1 - S_0 \tag{4.1}$$

where S_0 and S_1 are the 20-year averaged snowfall values from 2001 to 2020 and 2080 to 2099, respectively. In order to compare these changes to the trends in snowfall presented in the previous section, the difference between the mean snowfall values in the two periods can be converted to a rate of change by dividing the number of years that separate the two periods, or 80 years in this case. This method of examining future changes in snowfall is chosen to facilitate the decomposition of the snowfall change into the parts that are attributable to changes in temperature and precipitation (see section 4.4).

The results of this analysis are shown in Figures 4.5 to 4.24. Each color identifies results from a different model, where the symbols indicate different simulations performed with the same model. The box plots for each month show the full range of trends from the model simulations, as well as the median and 25th/75th percentile values. In the lower latitude regions, this analysis indicates that snowfall will likely decrease throughout the entire snowfall season. The strongest changes in midwinter snowfall are projected to occur along the western coast of North America where some individual months show median values of -20 to -30 cm decade^{-1}. Similarly, decreases in midwinter snowfall are found throughout much of central and southern Europe. It is also important to note that the fall decreases in snowfall are greater than the decreases in the spring months. One potential reason for this finding is that the snow-albedo effect (e.g., Lynch et al. 1998; Yang et al. 2001) is stronger in the spring and results in colder temperatures in the spring than in the fall.

In some regions, however, increases in midwinter snowfall are projected to occur. These regions are located at higher latitudes, such as Northern Québec, Siberia, and Greenland. In these locations, midwinter increases in snowfall are accompanied by decreases in transition season snowfall. In Québec and Siberia, however, annual trends are close to zero. These results provide evidence that midwinter increases in snowfall are likely to be offset by the decreases in snowfall that will occur during the transition seasons.

One of the key findings of this analysis is that there are no regions outside Greenland where snowfall increases in every month of the snowfall season. Increases in snowfall over Greenland throughout the entire snowfall season are a result of sufficiently cold temperatures, increased precipitation, and close proximity to the mid-latitude storm track. Although trends are positive in every month for Region 20, trends are very weak in the transition seasons. In the remaining regions across the Northern Hemisphere, the effects of reduced snowfall from warmer temperatures are felt at some point during the snow season.

The significance of this result is important to the overall discussion of snowfall changes. The results lend support to previous studies that argue for a shortening of the overall snowfall season. Furthermore, this is evidence of a concentration of snowfall into a shorter period of time in locations where annual trends are very weak, or close to zero (e.g., Regions 5 and 6). This concentration of snowfall may be more difficult to deal with from the human perspective, as local resources for snow removal and adaptation will likely need to be revised to handle the change in the distribution of snowfall.

The results further emphasize the importance of the snow-albedo feedback. The presence of a snow pack reflects shortwave radiation and cools the lower atmosphere, thus leading to a tendency for more snowfall to occur. A reduction of snowfall in the fall seasons may lead to a later development of the snow pack. As a result, warmer near-surface temperatures may persist later into the snowfall season. This mechanism is partially responsible for the decreases in autumn snowfall that are projected to occur across the Northern Hemisphere.

In summary, the findings of this analysis confirm that substantial changes are likely to occur within the snowfall season. Most locations will experience decreases in snowfall during the fall and spring transition seasons, while some locations will experience positive trends in snowfall during the middle of the winter season. The combination of increases and decreases in snowfall occurring at different times during the snow season appear to offset each other in some regions.

4.4. Decomposition of Temperature and Precipitation Effects on Snowfall

The occurrence of snowfall depends upon both precipitation and temperature. As the global climate warms, the accompanying changes in temperature and precipitation may have opposite effects on snowfall. In a warmer climate, the fraction of precipitation falling as snow would be expected to decrease, provided that temperatures are relatively close to the freezing point. This result is most likely to occur in regions to the south and also in the transition seasons. At higher latitudes where the temperature remains well below freezing throughout the snow season, increases of temperature have little direct impact on snowfall. Anthropogenic climate change also intensifies the hydrological cycle, enhancing precipitation in some locations. The interplay between the effects of changing temperature and the effects of changing precipitation is examined in this section.

In order to understand the interplay between temperature and precipitation, the simulated trends in monthly snowfall (ΔS), are decomposed as follows. Snowfall is expressed as the product of precipitation and the snowfall fraction:

$$S = P * SF \tag{4.2}$$

If $S_1, P_1,$ and SF_1 represent the monthly snowfall, precipitation, and fraction of the precipitation falling as snow averaged over the last 20 years of the 21st century, and $S_0, P_0,$ and SF_0 are the same quantities averaged over the first 20 years, then the changes in these quantities can be written as:

$$\Delta S = S_1 - S_0 \tag{4.3}$$

$$\Delta P = P_1 - P_0 \tag{4.4}$$

$$\Delta f = SF_1 - SF_0 \tag{4.5}$$

Through some algebraic manipulation, the difference in snowfall between the two periods can be written as

$$\Delta S = S_0 - S_1 = P_0 \Delta SF + SF_0 \Delta P + \Delta P \Delta SF \tag{4.6}$$

Thus, the changes in snowfall can be decomposed into three parts: the effects of changes in snow fraction with precipitation held constant, the effects of precipitation changes with snow fraction held constant, and the interaction between changes in snow fraction and precipitation. If it is assumed that snow fraction depends primarily on temperature, the three components can be regarded as the contributions of temperature changes, precipitation changes, and the interaction between temperature and precipitation changes to the change in monthly snowfall. As in section 4.3, these contributions are expressed as trends by dividing them by the time interval between the midpoints of the two averaging periods, which is 80 years, Figures 4.25 – 4.44 show the decomposition of the trends in monthly snowfall for the 20 regions used in this study.

The negative contributions of temperature are greatest during the transition seasons, as well as in midwinter in regions to the south. The negative contribution occurs throughout the entire snowfall season in most regions, with the exception of midwinter in the higher latitude regions (Regions 6, 7, 13, 19, and 20). Warmer temperatures reduce the probability that precipitation will fall in the form of snow. In the higher latitude locations, the effect becomes smaller since even modest increases in temperature would not impact the snow-to-precipitation ratio. This is a key result of this study, since it implies that temperature changes have a similar robust influence on snowfall in all of the regions. If acting alone, temperature increases of the 21st century would lead to decreases in snowfall across the Northern Hemisphere.

The contributions of ΔP to the snowfall trends are not as uniform, although some generalizations are apparent. The decompositions show that the individual effect of increasing precipitation offer mainly positive contributions to snowfall trends. The effects of ΔP are largest in the Arctic and Arctic subpolar regions and become negative in the midlatitudes. Most of the positive contributions from ΔP occur in midwinter. There are a few regions that show negative contributions from ΔP, and these regions are located in central and southern Europe.

Since changes in precipitation have a first-order effect on determining changes in snowfall, it is important to understand the mechanisms responsible for the spatial variations in precipitation. The result that a majority of the contributions of ΔP are toward positive changes in ΔS is related primarily to an enhancement of the meridional water vapor transport. The relationship between the convergence of water vapor transport and precipitation minus evaporation (P-E) can be used to gain insight into the pattern of precipitation trends in the 21st century simulations. The global moisture budget (e. g., Trenberth et al. 1995) can be written as:

$$\frac{\partial w}{\partial t} + \nabla \cdot \frac{1}{g}\int_0^{p_s} q\mathbf{v}\, dp = E - P \qquad (4.7)$$

where *w* is the column-integrated water vapor (precipitable water), q is the specific humidity, *v* is the wind vector, *E* is the surface evaporation, and *P* is precipitation. Over long time periods, the tendency term of precipitable water is small so that the divergence of the vertically-integrated moisture flux balances the difference of evaporation and precipitation. The relationship between time-averaged water vapor transport and P-E makes it possible to infer changes in the former from changes in the latter.

Held and Soden (2006) found that the general pattern of *P-E* changes little in a warmer climate and that this response is robust across climate models. Held and Soden (2006) also found, however, that the magnitude of *P-E* is enhanced. They explain this response as resulting from warming-induced increases in lower tropospheric water vapor, which simply enhance the existing pattern of meridional water vapor transport. This mechanism influences the spatial pattern of snowfall trends. Since *P-E* increases with latitude over the region of interest and evaporation changes do not vary as much in space as precipitation changes, 21st century trends in winter precipitation are greater at higher latitudes. Thus, the effects of temperature changes on snowfall trends are more likely to be offset by the effects of precipitation changes at higher latitudes.

As noted previously, however, the changes in snowfall related to the changes in ΔP are less uniform. The reason is that despite a large-scale amplification of *P-E*, the underlying patterns of precipitation changes are more complex. Across much of

North America and Asia, projections of 21st century climate show a poleward shift of the mid-latitude jet stream (e.g., Yin 2005). This poleward shift of the storm tracks further increases precipitation at higher latitudes when combined with the changes to P-E.

In order to understand the relative importance of changes in snow fraction versus changes in precipitation, monthly scatter plots of the two effects versus ΔS are shown (Figures 4.46 and 4.47). The decomposition analysis is performed on all of the individual model simulations, and every pairing of the two quantities for all 20 regions is plotted for each month. The figures indicate that the effects of changes in the snow fraction on snowfall are present in almost every month. The strongest links are present during midwinter where the correlation coefficients approach 1. In contrast, however, the scatter plots of ΔS versus ΔP show less of a linear relationship between than ΔS and ΔSF, as the dependence of ΔS on ΔP varies significantly from region to region. Correlation values for ΔS and ΔP are less than those found for ΔS and ΔSF, suggesting that in a general sense, regional variations in snowfall trends are more closely determined by temperature changes than precipitation changes.

The term representing the interaction of changes in precipitation and snow fraction ($\Delta P \Delta SF$) is negative and its magnitude is generally smaller than the other terms. The predominance of negative values of this term results from the nearly uniform negative values of ΔSF and widespread positive values of ΔP. This term is small, however, and does not have a first-order effect in determining the overall sign of ΔS.

The interplay of the effects of ΔT and ΔP in producing the simulated values of ΔS is not uniform at all latitudes. In the middle latitudes, there is often a balance between the competing effects of ΔP and ΔT that results in either a small positive or a negative overall trend in snowfall. At midlatitudes, the contribution of ΔP is smaller in magnitude than ΔT and the net result is a larger negative snowfall trend. In the northern regions, the positive values of ΔP substantially overwhelm the negative values of ΔT, especially from midwinter through early spring.

The importance of the interplay varies significantly from region to region. In western North America and central Europe, decreases in monthly snowfall are primarily the result of warmer temperatures and are only marginally affected by increases in precipitation. In the Mid-Atlantic regions (Regions 3 and 4), the competing effects are more balanced. This result adds support to the argument presented in Hayhoe et al. (2006), that the probability of precipitation occurring on cold days in these regions may stabilize a decrease in the number of snow days by the end of the 21st century. In the higher latitude regions, such as Greenland, any decreases in snowfall associated with warming are largely overwhelmed by increases in snowfall related to precipitation increases.

4.5. Changes in the Frequency of Daily Snowfall Events

The frequency of daily snowfall events exceeding a given threshold during the first and last twenty years of the 20th century is examined. Daily snowfall data from the models used in this study are not available in the CMIP3 archive. Daily data were obtained, however, for GFDL CM2.1 model simulation of the A1B scenario. As a result, the remaining 12 members of the multi-model ensemble used throughout this study are not examined.

The daily snowfall frequencies are analyzed at each grid point for the thresholds of 5 and 20 cm in a 24-hour period. In this analysis, frequencies are computed for events exceeding the given thresholds for two 20-year periods: 2001-2020 and 2081-2100. Figure 4.47 illustrates the differences in the frequencies for the two 20-year periods and also provides the raw frequency count for each threshold between 2001 and 2020.

Most midlatitude, subpolar, and high-latitude regions experience snowfall events that exceed 5 cm during the first 20 years of the 21st century. Large snowfalls that exceed 25 cm, however, are confined to the near-coastal sections of North America, southeastern Greenland, the Tibetan Plateau, and eastern Asia. Decreases in the number of daily events exceeding 5 cm are found across most locations between 20º and 60º N latitude, while increases in the number of events increases to the north. The decreases in events exceeding 5 cm are largest over the northeastern United States, southern Canada, the Rocky Mountains, southeastern Greenland and the Tibetan Plateau. The largest increases in the frequency of 5 cm snowfalls are located in the Central Plateau of Russia, the Kamchatka Peninsula, Alaska, and northern Greenland.

Decreases in the frequency of the events exceeding 20 cm are largest across eastern and western North America, southeastern Greenland, and across southeastern Asia. The largest increases in the number of events exceeding 20 cm are in southern Québec, the coasts of British Columbia and Alaska, and in some parts of the Tibetan Mountains.

The most striking feature is the decreases in the frequency of 5 cm snowfall combined with the increases in 20 cm snowfall that takes place over Québec. This location is coincident with decreases in annual snowfall during the 21st century. This implies that a greater fraction of the annual snowfall in this region will occur in larger, but relatively less frequent, events. The coexistence of increasingly frequent 24-hour large snowfall events with decreased total snowfall is a consequence of the interplay between temperature and precipitation. Precipitation increases at higher latitudes make larger snowfall events possible when the temperature is sufficiently cold to snow. There are indications that a similar shift toward larger, less frequent events will take place along the coast of British Columbia and in eastern China.

4.6. Signal-to-noise Analysis

The model simulations of snowfall during the 21st century that are discussed in earlier sections are characterized by the combination of the unforced variability of snowfall and the signal of a warming climate. A signal-to-noise analysis is performed on the time series of annual snowfall for each of the 20 regions used in this study. The purpose of this analysis is to determine when the simulated trend in snowfall emerges above the background noise of unforced variability. In practical terms, this point of emergence represents a time when the simulated changes in snowfall might become noticeable.

The signal-to-noise analysis is different from the trend analysis performed in previous sections because it assumes the vantage point of a 21st-century observer who has available only the time series of snowfall up to a particular point. The analysis utilizes pairwise t-tests to examine the statistical difference between sets of 21-year periods and the first 21 years of the 20^{th} century. The analysis first determines if the final 21 years (2080 – 2100) of the transient climate simulation and a reference period (1901 – 1921) are statistically different. The 1901 – 1921 reference period was chosen as the first set of 21 years are present in all of the individual model simulations. If a difference is detected between the two periods, the 21-year window is shifted backward by one year, so that the analysis period ranges from (2079 – 2099). This shift continues backward in time until there is no longer a statistical difference between the two periods. The emergence year is reported as the concluding year in the earliest 21-year period in the transient climate simulation that is significantly different from the years 1901-1921. If there is no statistical significance between final 21 years of the transient climate simulation and the reference period, then no emergence year is given. Thus, the emergence year represents the point in the simulated snowfall time series in which the mean snowfall over the previous 21 years is significantly different from the first 21 years of the 20th century and remains significantly different through the remainder of the time series.

The signal-emergence analysis is performed for all of the individual simulations of the CMIP3 models used in this study. In Figures, 4.49 to 4.54, the annual snowfall anomalies relative to the 1951 to 1999 base period are plotted. The observed anomalies are plotted in red for comparison. Each symbol along the x-axis denotes an emergence year from one of the simulations performed with the AOGCMs, with colors denoting models and symbols denoting different simulations performed with the same models. The median emergence year for each region is also noted on the figures, where “ND” indicates “no date” of emergence.

In the Mid-Atlantic and northeastern parts of the United States (Regions 1-4), the median emergence years are between approximately 2025 and 2050. In northeastern Canada, however, emergence years are much later in the 21st century with no signal emerging from a majority of individual model simulations in Region 6. Similarly, other arctic sub-polar and high latitude regions show emergence years that occur late in the 21st century. Some regions, however, show emergence years that will occur early in the 21st century. Such regions include western North America (Regions 8-10), central Europe (Regions 16-18), Japan (Region 12), and the area near Turkey (Region 15)

The result that some regions do not show an emergence year is common at higher latitudes. The absence of emergent trends in these regions arises because of the large variability of snowfall and the offsetting effects of decreasing snowfall during the transition seasons and increasing snowfall during midwinter – especially in Region 6. In regions where the signal emergence year occurs early in the twenty-first century, the trends in snowfall are particularly pronounced and are not obscured by a large year-to-year variability in annual snowfall. The presence of an early emergence year is extremely important, since it does not preclude the possibility that noticeable changes have already occurred. The years shown in Figures 4.49 to 4.54 represent the median from all model simulations, and the emergence years for some ensemble members have already passed. In these regions, changes in snowfall are likely to become prominent in the coming decades. As a result, some of the factors affected by snowfall (e.g., changes in runoff, lower stream flows in spring) may become apparent in the near future. The analysis shows that in some locations, strong trends in snowfall are likely to occur, and the effects of a changing climate may soon become apparent to snowfall observers.

In order to determine how these results compare to the signal emergence of temperature, the same analysis was repeated for the time series of annual temperature for each of the regions in this study. Table 4.1 shows the median emergence year for temperature compared to the median emergence year for snowfall in each of the 20 regions. In general, the positive trends in temperature become apparent at an earlier point in time when compared to the trends in snowfall.

To further illustrate this point, Figure 4.55 shows the mean of the standardized anomalies for snowfall obtained from the individual simulations are compared to the mean of the standardized anomalies of temperature based on a 1901-1930 base period. In this figure, Region 3 (Mid-Atlantic United States) is shown as an example where the snowfall signal emerges in the first half of the 21st century. The temperature anomalies in Region 3 become noticeable at an earlier point in time than snowfall. By the year 2030, temperature anomalies are 2.5 standard deviations above the mean, while snowfall is only 1 standard deviation below the mean. Similar results are also shown for Region 17 (Southern Europe). In contrast to temperature changes, snowfall changes take longer to become apparent. Therefore that annual snowfall may not be a particularly sensitive indicator of climate change during the next few decades.

4.7. Comparison of Results to Changes in SWE

One question that follows the study of future snowfall changes is how these results might compare to changes in snow cover in a warmer climate. Some work on the topic of snow cover changes has already been done. It is the objective of this section to compare the results in this dissertation to previous work regarding changes in snow cover.

Roesch (2006) compared the climatology of snow cover in 15 of the CMIP3 models to observations of the late twentieth century snow cover obtained from the United States Air Force Environmental Technical Applications Center (USAF-ETAC). Roesch found that the climate models used in his study overestimate snow water equivalence (SWE) in Eurasia during the winter months. Additionally, spring snowmelt is slower in the models than in the observational data. The inter-model standard deviation is low during the start of the snow cover season and gradually increases as the snow season progresses. This result is indicative that feedback processes involving snow cover are present in the models and that they vary significantly by model.

Räisänen (2008) also examined SWE in the CMIP3 models and drew comparisons with observations. Some slight differences in methodology exist, however, between the two studies. Räisänen used a larger ensemble of the CMIP3 models (20 versus the 15 used by Roesch). Räisänen also examined observational data from the Former Soviet Union Hydrological Snow Surveys (FSUHSS) data set. The FSUHSS data is of higher resolution compared to the USAF-ETAC data used by Roesch. Räisänen found that the models do overestimate SWE relative to observations in the early and later parts of the snow season. In contrast with Roesch, however, Räisänen found agreement between the models and the FSUHSS data in the middle of the snow season. Räisänen was able to replicate the key result of Roesch that the spring snow melt in the models is too slow compared to the rates found in observations.

Both Roesch and Räisänen discuss the potential feedback mechanism that may be present in the spring regarding snow cover and surface temperature. Both authors agree that it is unclear whether or not a cold bias in the models leads to increased SWE in the models during the spring, or vice versa. Räisänen found that temperatures across Eurasia in the spring are as much as 2º to 3° C colder in the models than in the CRU (University of East Anglia) data for the period of 1950 to 1999. Räisänen and

Roesch cite this cold bias as being partly responsible for the slow snow melt in spring.

The evaluation of the models in Chapter 2 indicates that the simulations of snowfall are exaggerated because of a cold bias in the models (e.g., John and Soden, 2007). Understanding the nature of the feedbacks between snow and temperature requires further examination to determine if such processes are accurately represented in the climate models. Further refinement of these processes will better improve future climate projections of temperature, particularly for the transitional seasons.

In addition to studying recent changes in SWE, Räisänen also examined future projections of SWE. There are several similarities between the future projections of SWE and the future projections of snowfall discussed in this chapter. Räisänen illustrates that SWE increases at higher latitudes and decreases at lower latitudes in the middle of the winter season, with decreases nearly everywhere during the fall and spring. Similarly, it is shown in section 4.1 that there is a transition between positive and negative trends in snowfall, although this feature is present throughout the entire snow season. Räisänen's results show that the transition between positive and negative changes in SWE in mid-winter broadly coincides with the NDJFM average -20°C isotherm of the late twentieth century. In this dissertation, it is shown that the transition line between positive and negative trends in annual snowfall coincides with the -10 to -15°C isotherm in all three seasons. Much colder temperatures are necessary in order to counteract melting processes and allow persistent snow cover to increase. In order for a location to receive only more snowfall, temperatures need not be as cold.

Räisänen shows that even in locations where SWE increases during the middle of the snow season, there are decreases in SWE during the fall and spring that shorten the overall snow cover season. The only exceptions are in extreme northern North America and Siberia, where the snow cover season is extended. In contrast to these results, however, the area that receives an increase in snowfall throughout the season is more expansive than the areas that see an increase in SWE. Räisänen cites both decreases in transitional season snowfall and increased melting of snow throughout the season as responsible mechanisms for this difference between SWE and snowfall.

Räisänen's results show similar changes in SWE when the early and late parts of the 21st century are compared to the late 20th century. The magnitude of the changes in the early part of 21st century, however, is smaller than that of the latter part of the 21st century. This result agrees with a lower signal-to-noise ratio in the early part of the 21st century. The quantitative signal-to-noise analysis of regional snowfall in section 4.4 supports this argument. Snowfall is influenced similarly by a weaker signal-to-noise ratio in the early part of the 21st century and it is not until the late 21st century, if at all, that the trend signal emerges from the background variability of the snowfall time series in a region.

Chapter 5

Summary and Conclusions

This dissertation examines both past and future changes in snowfall on a hemispheric and regional basis across the Northern Hemisphere. The focus of this work is to detail the climate mechanisms responsible for snowfall changes that are the result of increasing concentrations of anthropogenic greenhouse gases. Changes in snowfall affect natural processes such as spring snow melt and radiation, as well as human activities – including snow removal and tourism. The primary findings of this study are summarized.

Gridded cooperative snowfall observations (Dyer and Mote, 2006, 2007) were obtained for North America for the period 1951 to 1999. For other locations in the Northern Hemisphere, derived snowfall products were used that are based on monthly temperature and precipitation data from the Willmott-Matsuura climatology (Willmott and Matsuura 2002, http://climate.geog.udel.edu/~climate/). A derived snowfall product developed for the former Soviet Union was also used (Rawlins et al. 2006).

The observed climatology of snowfall was then compared to AOGCM simulations of snowfall obtained through the World Climate Research Programme's Third Coupled Model Inter-Comparison Project (WCRP CMIP3, Meehl et al. 2007). Simulations from 13 state-of-the-art climate models were used throughout this study. The selection of the models was based on the work of Reichler and Kim (2008), which systematically evaluated the performance of the CMIP3 models.

In both the observed and simulated climatology of snowfall across the Northern Hemisphere, a general latitudinal gradient exists. The highest amounts of annual snowfall are found in the polar and arctic subpolar regions, while lesser amounts are found in the midlatitude regions. The interior sections of the Northern Hemisphere continents typically receive less snowfall than coastal locations, as the availability of moisture is enhanced near the oceans. Snowfall is also greater in areas of higher terrain. Overall, western North America, northeastern Québec, Greenland, Tibet, and Japan receive the most annual snowfall, while northern China, parts of Siberia, central Europe, and the southeastern United States receive the least amounts of annual snowfall.

The comparison of the observed and simulated snowfall climatology indicates that the models are consistent with observations for most locations in the Northern Hemisphere. The most sizeable differences between the models and the observations, however, occur in higher elevation locations. This result is related to two factors. First, the resolution of AOGCMs limits their representation of mountainous terrain. These locations are represented as broad areas of increased elevation, which significantly affects how the models simulate temperature and precipitation. The general effect is a contribution to an over-estimation of snowfall in these locations. Secondly, observations of snowfall are biased toward lower elevation locations and do not capture the true magnitude of higher elevation snowfall. In addition to elevation concerns, most of the CMIP3 models exhibit a cold bias (John and Soden,

2007) that further contributes to a model overestimation of snowfall in most locations.

Using the observed and simulated snowfall data for the late-twentieth century, trends in annual snowfall were examined. Trends were examined on a grid point basis in order to demonstrate the large-scale patterns associated with the recent variability in snowfall. Trends in the late twentieth century annual snowfall indicate the presence of a transition zone between positive trends at higher latitudes and negative trends in the middle latitudes. This transition zone is located above 60°N latitude and is a robust feature of climate models and observations. This transition zone is the result of high latitude increases in precipitation in locations where warming during the late twentieth century does not alter the fraction of precipitation falling as snow.

Since the AOGCMs lack the spatial resolution necessary to accurately reproduce climate simulations on a grid point scale, regional averaging was employed to facilitate a fairer comparison between the observed and simulated trends. The data were averaged to create 20 regions across the Northern Hemisphere. As a result of this analysis, statistically significant trends in annual snowfall are found for several regions in the Northern Hemisphere – including western North America, Japan, and southern Russia. Furthermore, most of the trends in snowfall from the individual model simulations are statistically consistent with the observed trends during this period

An important finding of the analysis of the late twentieth century trends is that very few locations exhibit statistically significant trends. The combination of a weak trend signal resulting from greenhouse gas increases and large amounts of inter-annual variability in the time series of snowfall contributes to the lack of statistical significance in most locations. Analysis of trends from the multi-model ensemble, however, averages out the internal variability of the climate system and forced trends in snowfall do emerge – namely decreases in western North America and central Europe along with increases over Greenland.

The analysis of late-twentieth century trends also highlighted the mechanisms that are responsible for the observed changes in snowfall between 1951 and 1999. Late twentieth-century trends in annual snowfall are influenced by temperature increases and precipitation changes. There is also evidence, however, that mesoscale interactions – such as lake-effect snowfall – contribute to the observed trends in snowfall.

As noted previously, the cause of such changes is increased greenhouse gas forcing to the Earth's climate system. In the twenty-first century, this forcing will be greater, and thus, the effects will be more pronounced. Using the same methodology of the analysis of historical trends in snowfall, changes in snowfall were assessed for the twenty-first century. More analysis of changes in snowfall on the seasonal and monthly time scales was performed in order to answer questions related to snowfall changes occurring within the snowfall season. The findings of the grid point analysis of future trends in snowfall reveal that the transition zone between positive and negative trends in annual snowfall is also found for twenty-first century. The grid point trends in annual and seasonal snowfall were compared to the late twentieth century multi-model ensemble climatology of temperature. The transition zone between positive and negative trends of snowfall in the multi-model ensemble corresponds to the area bounded by the -10 and -15 ºC isotherms in each of the fall, winter, and spring seasons.

Trends in snowfall were examined on a monthly basis for each of the individual simulations performed with the CMIP3 models. This analysis was conducted for the same 20 regions identified in the analysis of late twentieth century trends in annual snowfall. The results of this analysis demonstrate that most regions experience decreases in snowfall during the fall and spring transition seasons. Some regions, however, show increases in mid-winter snowfall. In some of these locations where midwinter increases in snowfall are found (i.e. Region 6, northern Québec), grid point trends in annual snowfall are weak. This key result implies that decreases in snowfall that occur during the fall and spring seasons may be offset by increases in snowfall during the middle of the snowfall season.

This result prompted a more thorough analysis of the mechanisms responsible for the transitional season decreases and the midwinter increases in snowfall. The future changes in snowfall were decomposed into parts related to changes in the fraction of precipitation falling as snow and to changes in precipitation in order to determine which has a greater effect on trends in snowfall. The decomposition analysis indicates that changes in the snow fraction uniformly contribute to decreases in snowfall. Precipitation changes are generally positive, although their contributions are less uniform and greatest at higher latitudes.

Analysis of daily snowfall events was performed for the GFDL CM2.1 model in order to determine if changes in the frequency of large daily snowfall events relative to small daily snowfall events contribute to the overall changes in snowfall. The results of this analysis found that the frequency of small daily snowfall events (less than 5 cm) decreases across most mid-latitude Northern Hemisphere locations, but increases at arctic sub-polar and high latitude locations. In contrast, however, the frequency of large daily snowfall events (greater than 20 cm) does increase in northern Québec. This result implies that in this location where changes in total annual snowfall are weak relative to the other regions in this study, the snowfall may be compacted into less frequent but larger events.

A signal-to-noise analysis was also performed to determine when future changes in snowfall will emerge from background climate variability. The results of this signal-to-noise analysis indicate that the changes in annual snowfall will emerge for most regions from the background climate variability at some point during the twenty-first century. In some locations, however, where the inter-annual variability is large or the trend signal is weak, or both these conditions occur, no signal emergence year was found. These areas include Region 6, northern Québec; Region 14, southern Siberia; and Region 19, southeastern Greenland. The emergence years of the changes in snowfall, however, are found to be later than the emergence years of the changes in temperature.

The implications of this signal-to-noise analysis are important in broader discussions of anthropogenic climate change. The question as to whether or not snowfall has been an indicator of climate change still remains inconclusive. Despite signal emergence years that are projected to occur during the 21st century for most regions in the Northern Hemisphere, inter-annual variability will remain large. Therefore, it will still be possible for regions to experience high snowfall years in locations where there is a clear downward trend in annual snowfall.

A comparison was also made between the results of this study and other studies that examined similar changes in snow cover and SWE. The qualitative comparison that was performed did not reveal any significant discrepancies between these analyses of snowfall and previous studies related to snow cover. A transition zone between positive and negative trends in snow cover was also found in other studies, although studies found that this transition zone for SWE occurs at colder temperatures that were found for snowfall. Additionally, these studies project that changes in snow cover will also emerge from the background climate variability during the 21st century.

There are several factors to consider when evaluating and interpreting these results. Changes in snowfall are influenced by uncertainties in the observations. Kunkel et al. (2007) argue that cooperative observer data are subject to inconsistencies in the way in which the data are measured and recorded. Additionally, a low-elevation bias is present in the data since there are more measurements taken at lower elevations. The relatively fewer number of measurements at higher elevations presents a problem in studying snowfall changes. Changes at higher altitudes are potentially important from the standpoint of water resources, as the snow pack and subsequent snow melt are important to stream flows and the potential risks of flooding. Obtaining reliable snowfall data from locations across the Northern Hemisphere that are in a form that is ready for analysis is difficult. In studying historical changes in snowfall, there is a need for a hemispheric or global gridded climatology of snowfall that is quality-controlled and that accounts for elevation.

The accelerated loss of sea ice poses additional concerns in interpreting the results of this work. Reduced sea ice concentrations at higher latitudes increases the moisture flux from the ocean to the atmosphere. This would further enhance precipitation at higher latitudes. Recent observations of sea ice loss during the summer months are greater than what is simulated by AOGCMs. Therefore, the models may be underestimating snowfall increases at higher latitudes.

Observational practices in snowfall will continue to be an issue in future analyses of snowfall. The first-order observing stations in the United States are largely automated, and the traditional human measurements of snowfall are no longer taken. There are plans, however, to introduce ultrasonic or infrared snow depth sensors which are able to automate snow depth measurements. Such sensors may eventually improve upon traditional observational techniques, as the risk of human-induced errors is greatly reduced.

One of the major assumptions made in this study is the uniform 10:1 snow-to-liquid ratio in converting snowfall mass flux from the models to a snowfall depth. Research has shown that this ratio has substantial regional variations. Baxter et al. (2005) presents a climatology of the snowfall to liquid ratio derived from 30 years of cooperative observer data. Their findings illustrate that the empirically derived mean snowfall to liquid ratio is 13:1, which is 30% larger than the commonly used 10:1 ratio. Baxter et al. (2005) also show a range of ratios from 6:1 to 14:1, with the larger values found across parts of western Pennsylvania and east-central Oregon.

It is also possible that the snowfall to precipitation ratio varies as a function of the season. Since temperatures in the fall and spring seasons are warmer than in the midwinter, it is therefore likely that snowfall occurring in the middle of the snowfall season is less dense than snowfall that takes place in the transition seasons. It is also common for snow events to occur within the same season that have significantly different snowfall to precipitation ratios. Such variations are often the result of meteorological factors (i.e. temperatures in snow growth regions) and are difficult to simulate with climate models. Understanding of the regional variability of the snow-to-liquid ratio would further clarify the findings of this study.

Similar to the discussion of the snowfall to precipitation ratio, snow density may be a factor in interpreting the high latitude increases in snowfall uncovered in this study. As the temperature warms in the lower troposphere, the density of snowfall increases. In locations where temperature increases do not influence precipitation, the assumption of a 10:1 snowfall to precipitation ratio may overestimate the magnitude of the snowfall changes in these regions.

The limitation of the coupled AOGCMs used in this study must also be considered when interpreting the results of this work. Most of the AOGCMs used in this study diagnose snowfall using simple algorithms based on the temperature in the lower troposphere. While some of the newest AOGCMs now incorporate explicit mixed-phase microphysical schemes that keep independent water and ice budgets, it is unclear if such schemes significantly improve the simulation of snowfall in the models. Further research should investigate the performance of more detailed representations of mixed-phase precipitation on a model's ability to simulate snowfall.

Terrain representation was discussed throughout this dissertation as a significant limiting factor of the models in accurately simulating snowfall on the grid point scale. The results presented in Chapter 3 also illustrate that the potential exists for important changes in snowfall to occur on spatial scales that are too fine for the current grid spacing of modern AOGCMs. It is not possible for the typical AOGCM to represent mesoscale processes that are responsible for a large fraction of the total annual snowfall in some locations. One such example is the lee of the North American Great Lakes, where mesoscale processes heavily influence the snowfall climatology. Understanding how snowfall has changed and will continue to change in response to mesoscale mechanisms remains an active area of research for many locations of the Northern Hemisphere.

Regional climate modeling has the potential to answer the question concerning how mesoscale-induced snowfall might change in the future. Mesoscale processes important to snowfall are, in theory, better represented in the higher resolution models. Mesoscale models have an improved representation of terrain that is important in understanding the altitude variations of snowfall. Furthermore, the improved simulation of air interactions with water bodies and urban heat island effects would further increase the models' ability to simulate snowfall. Downscaling of AOGCM output for use in regional climate simulations of snowfall should be a priority in research related to this topic.

This dissertation provides further understanding of the mechanisms responsible for snowfall changes that are in response to increased amounts of anthropogenic greenhouse gases. These changes also have the potential to significantly impact natural climate processes – such as water resources and the Earth's radiation budget. The effect of snowfall changes on human activity is also of practical importance because of snow removal, health and safety issues, and tourism. As future assessments of the Earth's climate are performed, the understanding of the nature and mechanisms of changes in snowfall should continue to be examined.

References

Anderson TW, Rochard C, 1979: Cold snaps, snowfall and sudden death from ischemic heart disease. *Canadian Medical Association Journal,* **121**, 1580-3.

Andreescu, M., D. Frost, 1998: Weather and traffic accidents in Montreal, Canada. *Climate Res.,* **9**, 225-230.

Arakawa, H., S. Taga, J. F. Flores, V. F. Balagot, and M. Sukanto, 1969: Climates of Northern and Eastern Asia. World Survey of Climatology, Volume 8, Elsevier, New York.

Barnett, T. P., L. Dümenil, U. Schlese, E. Roeckner, and M. Latif, 1989: The Effect of Eurasian Snow Cover on Regional and Global Climate Variations. *J. Atmos Sciences,* **46**, 661-686.

Baxter, M. A., C. E. Graves, and J. T. Moore, 2005: A climatology of snow-to-liquid ratio for the contiguous United States. *Weather and Forecasting*, **20**, 729-744.

Blanford, H. F., 1884: On the Connexion of the Himalaya Snowfall with Dry Winds and Seasons of Drought in India. *Proc. Royal Society of London,* **37**, 3-22.

Bogdanova, E. G., B. M. Ilyin, and I. V. Dragomilova, 2002: Application of a comprehensive bias-correction model to precipitation measured at Russian North Pole drifting stations. *J. Hydrometeorol.,***3**, 700-713.

Box, J. E., D. H. Bromwich, B. A. Veenhuis, L. S. Bai, J. C. Stroeve, J. C. Rogers, K. Steffen, T. Haran, and S. H. Wang, 2006: Greenland ice sheet surface mass balance variability (1988-2004) from calibrated polar MM5 output. *J.Clim.,***19**, 2783-2800.

Bradbury, J. A., B. D. Keim, and C. P. Wake, 2002: U.S. East Coast trough indices at 500 hPa and New England winter climate variability. *J. Clim.,***15**, 3509-3517.

Bratten, R. O., 1996: An updated Canadian snow depth database, contract report. *Atmos. Environ*. Serv., 24 pp.

Brooks, C. F., 1917: New England snowfall. *Mon. Weather Rev.,* **45**, 271-285.

Burnett, A. W., M. E. Kirby, H. T. Mullins, and W. P. Patterson, 2003: Increasing Great Lake-effect snowfall during the twentieth century: A regional response to global warming? *J .Clim.,***16**, 3535-3542.

Cantu, V., 1977: The climate of Italy. *Climates and Central and Southern Europe, World Survey of Climatology Volume 6,* C. C. Wallen, Ed., 127-173.

Cehak-Trock, H., 1958: DerfesteNiederschlagimatlantischenKlimagebiet, *Arch. Meteorologie, Geophysik, Bioklimatologie*, *Serie B*, **8(**3/4), 352–368.

Changnon, S. A., 2006: Frequency Distributions of Heavy Snowfall from Snowstorms in the United States. *J. Hydrol. Eng.,* **11**, 427.

Changnon, S., D. Changnon, 2006: A Spatial and Temporal Analysis of Damaging Snowstorms in the United States. *Nat. Hazards,* **37**, 373-389.

Christensen, J.H., B. Hewitson, A. Busuioc, A. Chen, X. Gao, I. Held, R. Jones, R.K. Kolli, W.-T. Kwon, R. Laprise, V. MagañaRueda, L.Mearns, C.G. Menéndez, J. Räisänen, A. Rinke, A. Sarr and P. Whetton, 2007: Regional Climate Projections. In: Climate Change 2007: The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change [Solomon, S., D. Qin, M. Manning, Z. Chen, M. Marquis, K.B. Averyt, M. Tignor and H.L. Miller (eds.)]. Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA.

Cohen, J., D. Entekhabi, 1999: Eurasian snow cover variability and Northern Hemisphere climate predictability. *Geophys. Res. Lett.,* **26**, 345-348.

Collins, W.D., et al., 2004: *Description of the NCAR Community Atmosphere Model (CAM3.0)*. Technical Note TN-464+STR, National Center for Atmospheric Research, Boulder, CO, 214 pp.

da Silva, R. R., G. Bohrer, D. Werth, M. J. Otte, and R. Avissar, 2006: Sensitivity of ice storms in the southeastern United States to Atlantic SST - Insights from a case study of the December 2002 storm. *Mon. Weather Rev.,* **134**, 1454-1464.

Daly, C., R. P. Neilson, and D. L. Phillips, 1994: A Statistical-Topographic Model for Mapping Climatological Precipitation over Mountainous Terrain. *J. Appl. Meteorol.,*, 140-158.

Danckwortt, P. W., 1923: The Economic Geography of Siberia. *Geographical Review,* **13**, 314-315, doi:10.2307/208456.

Davis, R. E., M. B. Lowit, P. C. Knappenberger, and D. R. Legates, 1999: A climatology of snowfall-temperature relationships in Canada. *J.Geophys.Res.,* **104**, 11985-11994.

De Freitas, C. R., Estimation of the Disruptive Impact of Snowfalls in Urban Areas. *J. Appl. Meteorol.,***14**, 1166-1173.

Dyer, J. L., T. L. Mote, 2006: Spatial variability and trends in observed snow depth over North America. *Geophys. Res. Lett.,***33**, L16503.

--2007: Trends in snow ablation over North America. *Int. J. Climatol.,***27**, 739-748.

Eisenberg, D., K. E. Warner, 2005: Effects of snowfalls on motor vehicle collisions, injuries, and fatalities. *Am. J. Public Health,* **95**, 120-124.

Elsasser, H., P. Messerli, 2001: The Vulnerability of the Snow Industry in the Swiss Alps. *Mountain Research and Development,* **21**, 335-339.

Feldstein, S. B., 2002: The Recent Trend and Variance Increase of the Annular Mode. *J. Clim.,* **15**, 88-94.

Flato, G.M., 2005: *The Third Generation Coupled Global Climate Model (CGCM3)* (and included links to the description of the AGCM3 atmospheric model). http://www.cccma.bc.ec.gc.ca/models/cgcm3.shtml.

Fujita, T. T., Satellite and Mesometeorology Research Project, and University of Chicago Department of the Geophysical Sciences, 1971: *Proposed Characterization of Tornadoes and Hurricanes by Area and Intensity.* University of Chicago, .

Gagnon, A. S., W. A. Gough, 2005: Climate change scenarios for the Hudson Bay region: An intermodel comparison. *Clim. Change,* **69**, 269-297.

Gazzolo, T., M. Pinna, 1973: *La Nevosita in Italia nel Quarantennio1921-1960.* MinisteriodeiLavoriPubblici, ServizioIdrografico, 211 pp. pp.

GFDL GAMDT (The GFDL Global Atmospheric Model Development Team), 2004: The new GFDL global atmosphere and land model AM2-LM2: Evaluation with prescribed SST simulations. *J. Clim*., 17, 4641–4673.

Gordon, H.B., et al., 2002: *The CSIRO Mk3 Climate System Model*. CSIRO Atmospheric Research Technical Paper No.60, Commonwealth Scientific and Industrial Research Organisation Atmospheric Research Aspendale, Victoria, Australia, 130 pp, http://www.cmar.csiro.au/e-print/open/gordon_2002a.pdf.

Gorjanc ML, Flanders WD, VanDerslice J, Hersh J, and Malilay J, 1999: Effects of temperature and snowfall on mortality in Pennsylvania. *American Journal Of Epidemiology,* **149**, 1152-60.

Groisman, P. Y., D. R. Easterling, 1994: Variability and Trends of Total Precipitation and Snowfall Over the United-States and Canada. *J. Clim.,***7**, 184-205.

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.

Groisman, P. Y., E. Y. Rankova, 2001: Precipitation trends over the Russian permafrost-free zone: Removing the artifacts of pre-processing. *Int. J. Climatol.,***21**, 657-678.

Hall, A., 2004: The role of surface albedo feedback in climate. *J. Clim.,***17**, 1550-1568.

Harrington, M., 1894: Rainfall and snow of the United States. *US Weather Bureau Bulletin C,* **80**.

Harrington, J. A., R. S. Cerveny, and K. F. Dewey, 1987: A Climatology of Mean Monthly Snowfall for the Conterminous United States: Temporal and Spatial Patterns. *J. App. Meteor.,***26**, 897-912.

Hart, R. E., R. H. Grumm, 2001: Using normalized climatological anomalies to rank synoptic-scale events objectively. *Mon.Weather Rev.,* **129**, 2426-2442.

Hartley, S., M. J. Keables, 1998: Synoptic associations of winter climate and snowfall variability in New England, USA, 1950-1992. *Int.J.Climatol.,***18**, 281-298.

Hayhoe, K., C. P. Wake, T. G. Huntington, L. F. Luo, M. D. Schwartz, J. Sheffield, E. Wood, B. Anderson, J. Bradbury, A. DeGaetano, T. J. Troy, and D. Wolfe, 2007: Past and future changes in climate and hydrological indicators in the US Northeast. *Clim. Dyn.,***28**, 381-407.

Held, I. M., B. J. Soden, 2006: Robust Responses of the Hydrological Cycle to Global Warming. *J. Clim.,***19**, 5686-5699.

Hurrell, J. W., 1995: Decadal Trends in the North Atlantic Oscillation: Regional Temperatures and Precipitation. *Science,* **269**, 676-679.

John, V. O., B. J. Soden, 2007: Temperature and humidity biases in global climate models and their impact on climate feedbacks. *Geophys. Res. Lett.,***34**, L18704.

Joyce, T. M., 2002: One Hundred Plus Years of Wintertime Climate Variability in the Eastern United States. *Journal of Climate,* **15**, 1076-1086.

K-1 Model Developers, 2004: K-1 Coupled Model (MIROC) Description. K-1 Technical Report 1 [Hasumi, H., and S. Emori (eds.)]. Center for Climate System Research, University of Tokyo, Tokyo, Japan, 34 pp.,http://www.ccsr.u-tokyo.ac.jp/kyosei/hasumi/MIROC/tech-repo.pdf.

Karl, T. R., P. Y. Groisman, R. W. Knight, and R. R. Heim Jr, 1992: Recent Variations of Snow Cover and Snowfall in North America and Their Relation to Precipitation and Temperature Variations. *J. Clim.,***6**, 1327-1344.

Kirby, M. E., H. T. Mullins, W. P. Patterson, and A. W. Burnett, 2001: Lacustrine isotopic evidence for multidecadal natural climate variability related to the circumpolar vortex over the northeast United States during the past millennium. *Geology,* **29**, 807-810.

Knowles, N., M. D. Dettinger, and D. R. Cayan, 2006: Trends in snowfall versus rainfall in the Western United States. *J. Clim.,***19**, 4545-4559.

Kocin, P. J., L. W. Uccellini, 2004: A snowfall impact scale derived from Northeast storm snowfall distributions. *Bull. Amer. Meteor. Soc.,***85**, 177-194.

Kripalani, R. H., A. Kulkarni, 1999: Climatology and variability of historical Soviet snow depth data: some new perspectives in snow - Indian monsoon teleconnections. *Clim. Dyn.,***15**, 475-489.

Kunkel, K. E., J. R. Angel, 1999: Relationship of ENSO to snowfall and related cyclone activity in the contiguous United States. *J. Geophys. Res.,* **104**, 19425-19434.

Kunkel, K. E., M. A. Palecki, K. G. Hubbard, D. A. Robinson, K. T. Redmond, and D. R. Easterling, 2007: Trend identification in twentieth-century US snowfall: The challenges. *J. Atmos. Ocean. Technol.,***24**, 64-73.

Kunkel, K. E., N. E. Westcott, and D. A. R. Kristovich, 2002: Assessment of potential effects of climate change on heavy lake-effect snowstorms near Lake Erie. *J. Great Lakes Res.,* **28**, 521-536.

Lanzante, J. R., 2005: A cautionary note on the use of error bars. *J. Clim.,* **18**, 3699-3703.

Laternser, M., M. Schneebeli, 2003: Long-term snow climate trends of the Swiss Alps (1931-99). *Int. J. Climatol.,* **23**, 733-750.

Legates, D. R., C. J. Willmott, 1990: Mean seasonal and spatial variability in gauge-corrected, global precipitation. *Int. J. Climatol.,* **10**, 111-127, doi:10.1002/joc.3370100202.

Lemke, P., J. Ren, R.B. Alley, I. Allison, J. Carrasco, G. Flato, Y. Fujii, G. Kaser, P. Mote, R.H. Thomas and T. Zhang, 2007: Observations: Changes in Snow, Ice and Frozen Ground. In: Climate Change 2007: The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change [Solomon, S., D. Qin, M. Manning, Z. Chen, M. Marquis, K.B. Averyt, M. Tignor and H.L. Miller (eds.)]. Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA.

Lettenmaier, D., T. Gan, 1990: Hydrologic sensitivities of the Sacramento-San Joaquin River basin, California, to global warming. *Water Resources Research,* **26**, 69-86.

Lydolph, P.E.,1977: Climates of the Soviet Union. World Survey of Climatology, Volume 7, Elsevier, New York.

Lynch, A. H., D. L. McGinnis, and D. A. Bailey, 1998: Snow-albedo feedback and the spring transition in a regional climate system model: Influence of land surface model. *J. of Geophys. Res.*, **103**, 29037-29049.

Laternser, M. S., 2003: Long-term snow climate trends of the Swiss Alps (1931-99). *Int. J. Climatol.,* **23**, 733-750.

McFarlane, N.A., G.J. Boer, J.-P. Blanchet, and M. Lazare, 1992: The Canadian Climate Centre second-generation general circulation model and its equilibrium climate. *J. Clim.*, **5**, 1013–1044.

Martin, G. M., M. A. Ringer, V. D. Pope, A. Jones, C. Dearden, and T. J. Hinton, 20060401: The Physical Properties of the Atmosphere in the New Hadley Centre Global Environmental Model (HadGEM1). Part I: Model Description and Global Climatology. *J. Clim.,* **19**, p1274.

Matsuura, S., K. Matsuyama, S. Asano, T. Okamoto, and Y. Takeuchi, 2005: Fluctuation of the seasonal snowpack in a mountainous area of the heavy-snow district in the warm-temperate zone of Japan. *J. Glaciol.,* **51**, 547-554.

Meehl, G. A., C. Covey, T. Delworth, M. Latif, B. McAvaney, J. F. B. Mitchell, R. J. Stouffer, and K. E. Taylor, 2007: The WCRP CMIP3 multimodel dataset - A new era in climate change research. *Bull. Amer. Meteor. Soc.,* **88**, 1383-1394.

Miller, R. L., G. A. Schmidt, and D. T. Shindell, 2006: Forced annular variations in the 20th century intergovernmental panel on climate change fourth assessment report models. *J.Geophys.Res.-Atmos.,* **111**, D18101.

Mitchell, T. D., P. D. Jones, 2005: An improved method of constructing a database of monthly climate observations and associated high-resolution grids. *Int.J.Climatol.,* **25**, 693-712.

Namias, J., 1960: Snowfall over the eastern United States: factors leading to its monthly and seasonal variations. *Weatherwise,* **13**, 238–247.

Notaro, M., W. -. Wang, and W. Gong, 2006: Model and Observational Analysis of the Northeast U.S. Regional Climate and Its Relationship to the PNA and NAO Patterns during Early Winter. *Monthly Weather Review,* **134**, 3479-3505.

Pope, V.D., M.L. Gallani, P.R. Rowntree, and R.A. Stratton, 2000: The impact of new physical parametrizations in the Hadley Centre climate model: HadAM3. *Clim. Dyn.*, **16**, 123–146.

Räisänen, J., 2008: Warmer climate: less or more snow? *Clim. Dyn.,* **30**, 307-319.

Rawlins, M. A., C. J. Willmott, A. Shiklomanov, E. Linder, S. Frolking, R. B. Lammers, and C. J. Vorosmarty, 2006: Evaluation of trends in derived snowfall and rainfall across Eurasia and linkages with discharge to the Arctic Ocean. *Geophys .Res. Lett.,* **33**, L07403.

Rechard, P. A., L. W. Larson, 1971: The use of snow fences for shielding precipitation gauges. *Proc. 39th Western Snow Conference*, pp. 56-62, Billings, Mont.

Reichler, T., J. Kim, 2008: How well do coupled models simulate today's climate? *Bull. Amer. Meteorol.Soc.,* **89**, 303.

Reinosdotter, K., M. Viklander, 2006: Handling of urban snow with regard to snow quality. *J.Environ.Eng.- ASCE,* **132**, 271-278.

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

Robinson, D. A., A. Frei, 2000: Seasonal Variability of Northern Hemisphere Snow Extent Using Visible Satellite Data. *The Professional Geographer,* **52**, 307-315.

Robinson, D. A., 1987: Snow cover as an indicator of climate change. AD-A186 880, National Technical Information Service, Springfield VA.

Robock, A., 1980: The seasonal cycle of snow cover, sea ice, and surface albedo. *Mon. Weather Rev.,* **108**, 267-285.

--1985: An updated climate feedback diagram. *Bull. Amer. Met. Soc.,***66**, 786-787.

Robock, A., M. Mu, K. Vinnikov, D. Robinson, 2003: Land surface conditions over Eurasia and Indian summer monsoon rainfall, *J. Geophys. Res.*, **108**(D4), 4131, doi:10.1029/2002JD002286.

Roeckner, E., K. Arpe, L. Bengtsson, M. Christoph, M. Claussen, L. Dümenil, M. Esch, M. Giorgetta, U. Schlese, and U. Schulzweida, 1996: The atmospheric general circulation model ECHAM-4: Model description and simulation of present-day climate. Reports of the Max-Planck-Institute, Hamburg, No.218, 90 pp.

Roeckner, E. K., et al., 2003: *The Atmospheric General Circulation Model ECHAM5. Part I: Model Description*. MPI Report 349, Max Planck Institute for Meteorology, Hamburg, Germany, 127 pp.

Roesch, A., 2006: Evaluation of surface albedo and snow cover in AR4 coupled climate models. *J. Geophys. Res.,***111**, D15111.

Rogot E, Padgett SJ, 1976: Associations of coronary and stroke mortality with temperature and snowfall in selected areas of the United States, 1962-1966. *American Journal Of Epidemiology,* **103**, 565-75.

Rooney, J. F., 1967: Urban snow hazard in the United States – appraisal of disruption. *Geog. Rev.,* **57**, 538-559.

Schemenauer, R. S., M. O. Berry, and J. B. Maxwell, 1981: Snowfall Formation. *Handbook of Snow: Principles, Processes, Management, & Use,* D. M. Gray and D. H. Male, Eds., 1st ed. Pergamon Press, 129-152.

Schneebeli, M., M. Laternser, and W. Ammann, 1997: Destructive snow avalanches and climate change in the Swiss Alps. *Eclogae Geologicae Helvetiae,* **90**, 457-461.

Scott, D., Kaiser D, 2004: U.S. snowfall trends and variability. *Bull. Amer. Meteor. Soc.,* **85**, 501-3.

Shepard, D., 1968: A two-dimensional interpolation function for irregularly-spaced data. , 517-523.

Shibata, K., et al., 1999: A simulation of troposphere, stratosphere and mesosphere with an MRI/JMA98 GCM. *Papers in Meteorology and Geophysics*, **50**, 15–53.

Simpson, R. H., 1974: The hurricane disaster potential scale. *Weatherwise,* **27**, 169-186.

SpitalnicS.J.,Jagminas L, and Cox J, 1996: An association between snowfall and ED presentation of cardiac arrest. *The American Journal Of Emergency Medicine,* **14**, 572-3.

Stottlemyer, R., D. Toczydlowski, 1999: Seasonal change in precipitation, snowpack, snowmelt, soil water and streamwater chemistry, northern Michigan. *Hydrol. Process.,***13**, 2215-2231.

Taylor, K. E., 2001: Summarizing multiple aspects of model performance in a single diagram. *J. Geophys. Res.,* **106**, 7183-7192.

Thomas, R., E. Frederick, W. Krabill, S. Manizade, and C. Martin, 2006: Progressive increase in ice loss from Greenland. *Geophys. Res. Lett.,***33,** L10503.

Trenberth, K.E. and C. J. Guillemot, 1995: Evaluation of the global atmospheric moisture budget as seen from analyses. *J. Clim.*, **8**, 2255-2272.

Trenberth, K.E., P.D. Jones, P. Ambenje, R. Bojariu, D. Easterling, A. Klein Tank, D. Parker, F. Rahimzadeh, J.A. Renwick, M. Rusticucci, B. Soden and P. Zhai, 2007: Observations: Surface and Atmospheric Climate Change. In: Climate Change 2007: The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change [Solomon, S., D. Qin, M. Manning, Z. Chen, M. Marquis, K.B. Averyt, M. Tignor and H.L. Miller (eds.)]. Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA.

Uccellini, L. W., P. J. Kocin, R. S. Schneider, P. M. Stokols, and R. A. Dorr, 1995: Forecasting the 12-14 March 1993 Superstorm. *Bull. Amer. Meteorol. Soc.,***76**, 183-199.

Vaganov, E. A., M. K. Hughes, A. V. Kirdyanov, F. H. Schweingruber, and P. P. Silkin, 1999: Influence of snowfall and melt timing on tree growth in subarctic Eurasia. *Nature,* **400**, 149-151.

Volney, C., 1804: The View of the Climate and Soil of the United States of America. *C. Mercier and Co, London,* 446 p.

Wallen, C. C., T. W. Johannessen, G. Manley, R. Arlery, and A. L. Escardo, 1970: *Climates of Northern and Western Europe.* World Survey of Climatology, Volume 5, Elsevier.

Wallen, C. C., M. Schuepp, H. Schirmer, W. Okolowicz, V. Cantu, and D. Furlan, 1977: *Climates of Central and Southern Europe.* World Survey of Climatology, Volume 6, Elsevier.

Watts, I. E. M., 1969: Climates of China and Korea. *World Survey of Climatology Volume 8: Climates of Northern and Eastern Asia,* H. Arakawa, Ed., Elsevier, 1-118.

Wilks, D. S., 1997: Resampling hypothesis tests for autocorrelated fields. *J. Clim.,***10**, 65-82.

Willmott, C. J., K. Matsuura, 2001: Arctic terrestrial air temperature and precipitation: Monthly and annual time series (1930–2000), version 1. *http://climate.geog.udel.edu/~climate/,* .

Willmott, C. J., S. M. Robeson, 1995: Climatologically aided interpolation (CAI) of terrestrial air temperature. *Int. J. Climatol.,***15**, 221-229.

Willmott, C. J., C. M. Rowe, and W. D. Philpot, 1984: Spheremap, software. Center for Climate Research, Dept.of Geography, Univ. of Delaware

Yang, F. L., A. Kumar, W. Q. Wang, H. M. H. Juang, and M. Kanamitsu, 2001: Snow-albedo feedback and seasonal climate variability over North America. *J. Clim*.,**14**, 4245-4248.

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

Yin, J. H., 2005: A consistent poleward shift of the storm tracks in simulations of 21st century climate. *Geophys. Res. Lett.,***32**, L18701.

Zar, J. H., 1974: Biostatistical analysis. Prentice Hall, Englewood Cliffs, NJ, 620 pp.

Zhang, Y. S., T. Li, and B. Wang, 2004: Decadal change of the spring snow depth over the Tibetan Plateau: The associated circulation and influence on the East Asian summer monsoon. *J. Clim.,***17**, 2780-2793.

Zielinski, G. A., 2002: A classification scheme for winter storms in the eastern and central United States with an emphasis on nor'easters. *Bull. Amer. Meteor. Soc.,* **83**, 37-52.

	Country	Atmos. Res. (lat. x lon.)	# of 20c Runs	# of 21c Runs	P-Type Method	Reference
CGCM 3.1 (T47)	Canada	~2.8° x 2.8°	5	5	DS	McFarlane et al. 1992; Flato 2005
CGCM 3.1 (T63)	Canada	~1.9° x 1.9°	1	1	DS	McFarlane et al. 1992; Flato 2005
CSIRO-MK3.0	Australia	~1.9° x 1.9°	3	1	PS	Gordon et al. 2002
ECHAM5	Germany	~1.9° x 1.9°	4	4	PS	Roeckner et al. 2003
GFDL CM2.0	USA	2.0° x 2.5°	3	1	DS	GFDL GAMDT 2004
GFDL CM2.1	USA	2.0° x 2.5°	3	1	DS	GFDL GAMDT 2004
INGV SXG	Italy	~1.1° x 1.1°	1	1	DS	Roeckner et al. 1996
MIROC 3.2 (HI-RES)	Japan	~1.1° x 1.1°	1	1	DS	K-1 Developers 2004
MIROC 3.2 (MED-RES)	Japan	~2.8° x 2.8°	3	3	DS	K-1 Developers 2004
MRI CGCM 2.3.2	Japan	~2.8° x 2.8°	5	5	DS	Shibata et al. 1999
NCAR CCSM3	USA	~1.4° x 1.4°	8	7	PS	Collins et al. 2004
UKMO HADCM3	UK	2.5° x 3.75°	2	1	DS	Pope et al. 2000
UKMO HADGEM1	UK	~1.3° x 1.9°	2	1	PS	Martin et al. 2004

Table 2.1. CMIP3 models used in this study, including country, model resolution, and the number of historical and future simulations analyzed. "DS" denotes a diagnostic scheme for determining precipitation type, while "PS" denotes prognostic schemes.

	North America (cm)	TD-9813 Domain (cm)	Eurasia (cm)	N. Hemisphere (cm)
NorAm	129	---	---	---
TD-9813	---	165	---	---
WM	148	100	73	76
CGCM 3.1 (T47)	155	147	91	88
CGCM 3.1 (T63)	176	154	98	95
CSIRO-MK3.0	209	204	118	113
ECHAM5	207	217	120	117
GFDL CM2.0	273	241	174	155
GFDL CM2.1	253	244	157	142
INGV SXG	252	260	142	139
MIROC 3.2 (HI-RES)	171	181	106	99
MIROC 3.2 (MED-RES)	187	295	112	106
MRI CGCM 2.3.2	182	203	118	111
NCAR CCSM3	205	241	123	118
UKMO HADCM3	221	197	136	127
UKMO HADGEM1	229	223	134	126

Table 2.2. Area-averaged annual snowfall (cm) for North America, TD-9813 domain, Eurasia, and the Northern Hemisphere. Annual snowfall averaged over the period 1951 to 1999. Values reported for the three observational data sources (where available), and for the 13 CMIP3 models.

	Latitude Range	Longitude Range	Observed Data
Region 1	31° N – 36° N	83° W – 75° W	NorAm
Region 2	36° N – 41° N	80° W – 72° W	NorAm
Region 3	41° N – 46° N	76° W – 60° W	NorAm
Region 4	46° N – 51° N	72° W – 59° W	NorAm
Region 5	51° N – 56° N	63° W – 55° W	NorAm
Region 6	56° N – 61° N	71° W – 61° W	NorAm
Region 7	50° N – 63° N	87° W – 75° W	NorAm
Region 8	35° N – 48° N	125° W – 115° W	NorAm
Region 9	48° N – 58° N	135° W – 120° W	NorAm
Region 10	58° N – 63° N	165° W – 135° W	NorAm
Region 11	60° N – 68° N	160° E – 170° W	WM
Region 12	30° N – 48° N	128° E – 148° E	WM
Region 13	65° N – 75° N	70° E – 90° E	TD-9813
Region 14	45° N – 55° N	83° E – 105° E	TD-9813
Region 15	35° N – 43° N	38° E – 48° E	WM
Region 16	43° N – 55° N	0° E – 33° E	WM
Region 17	55° N – 63° N	4° E – 33° E	WM
Region 18	63° N – 73° N	4° E – 33° E	WM
Region 19	60° N – 70° N	45° W – 30° W	WM
Region 20	67° N – 85° N	30° W – 13° W	WM

Table 3.1. Geographical boundaries of the 20 regions used in this study and the sources of observed data for each of the regions.

	Signal-Emergence Year Temperature	Signal-Emergence Year Snowfall
Region 1	1992	2042
Region 2	1998	2029
Region 3	1997	2025
Region 4	1987	2053
Region 5	2002	2060
Region 6	1997	N.E.
Region 7	1985	2088
Region 8	1986	2012
Region 9	1999	2017
Region 10	2000	2058
Region 11	1997	2099
Region 12	1990	2012
Region 13	1989	2091
Region 14	1990	N.E.
Region 15	1985	2021
Region 16	1999	2015
Region 17	2003	2008
Region 18	1996	2044
Region 19	1993	N.E.
Region 20	1984	2051

Table 4.1. Signal-emergence years for annual temperature and annual snowfall. Regions were no emergence year is found are denoted by "N.E."

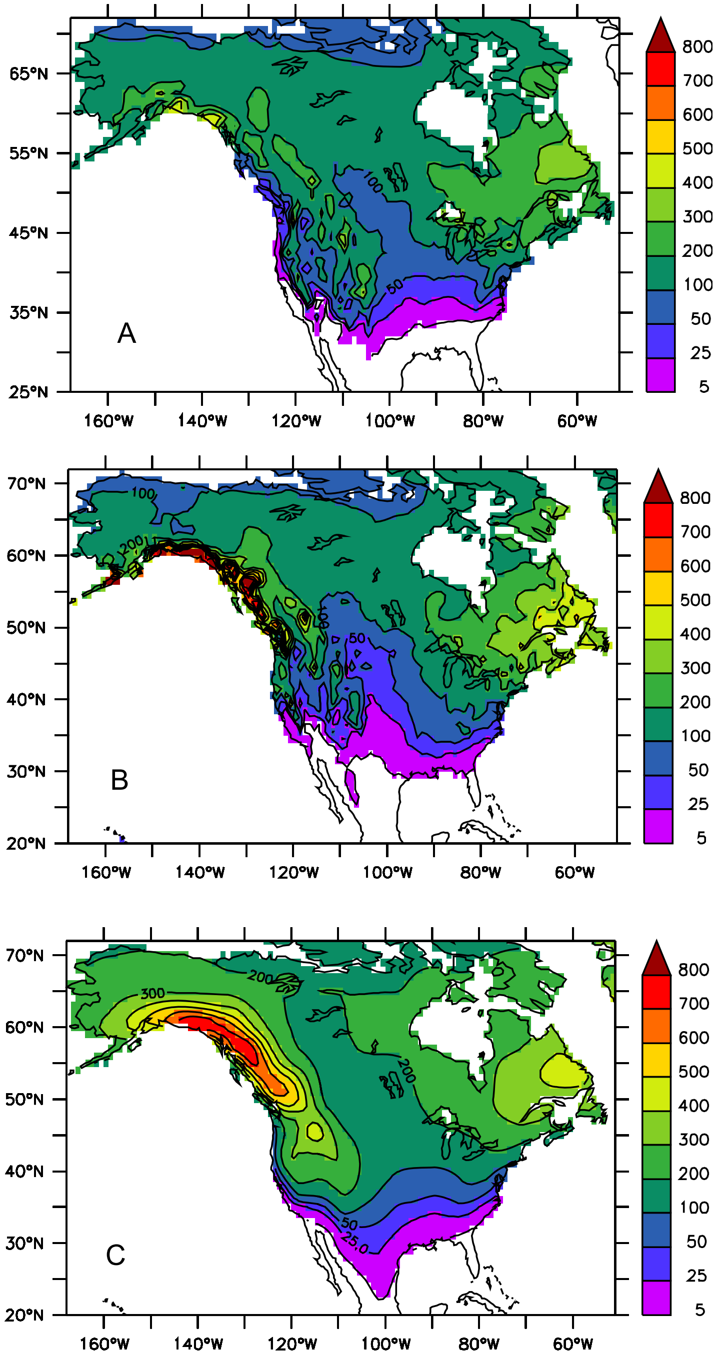


Figure 2.1. North American climatology of snowfall (cm/year) for the NorAm gridded observations (A), WM derived snowfall (B), and the multi-model ensemble (C).

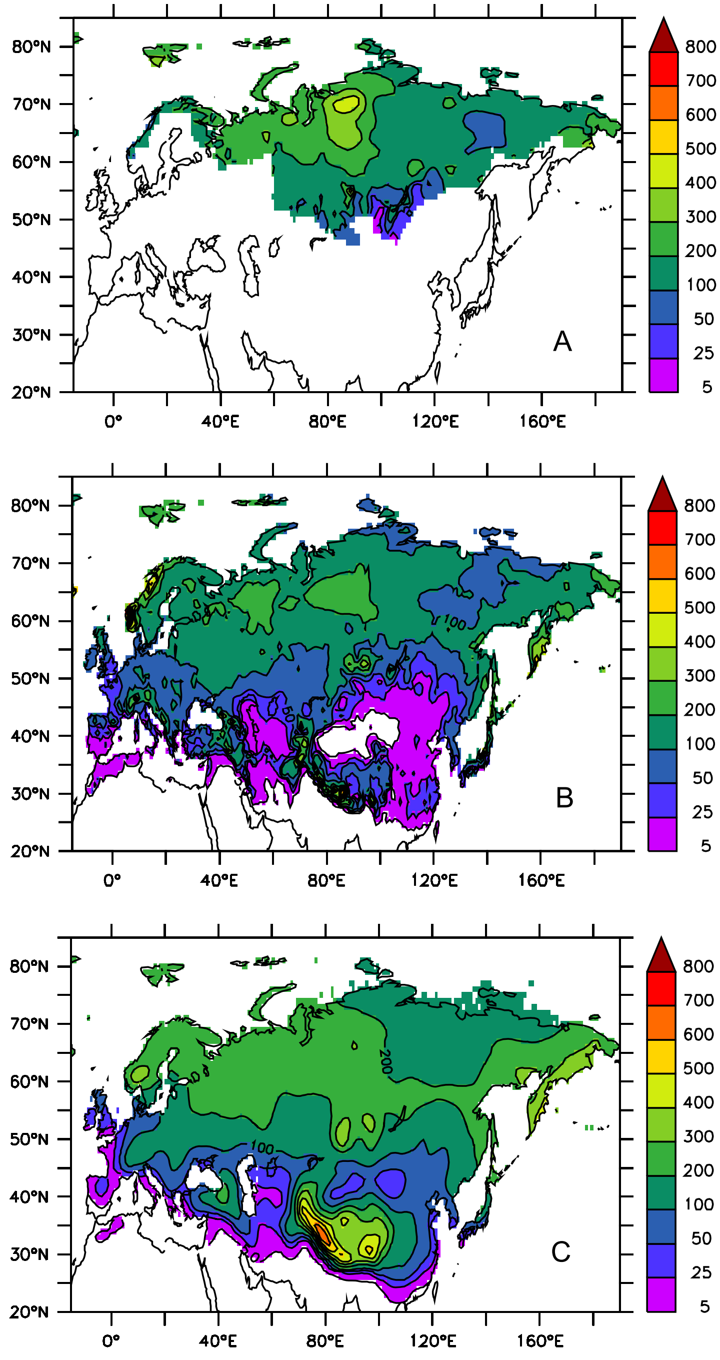
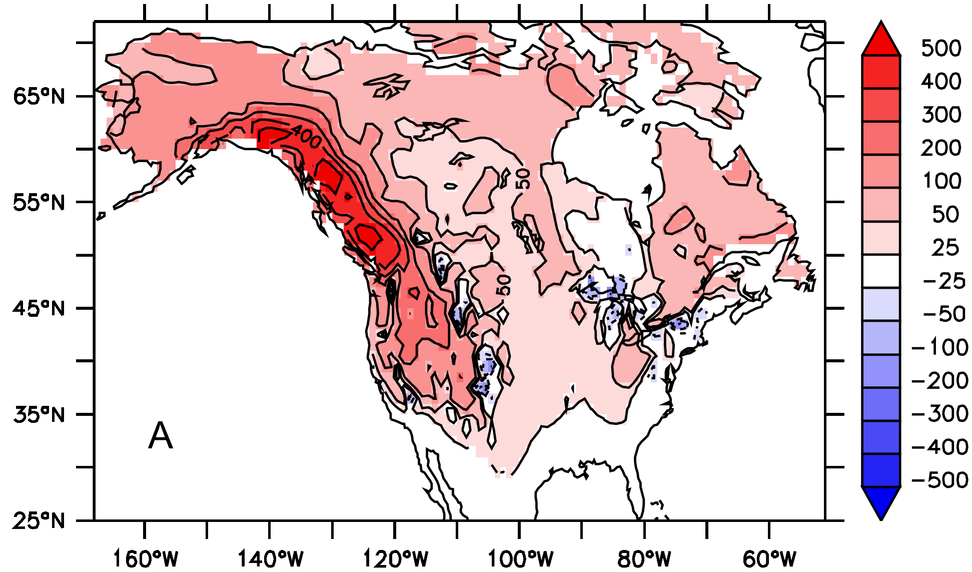


Figure 2.2. Eurasian climatology of snowfall (cm/year) for the TD-9813 derived snowfall (A), WM derived snowfall (B), and the multi-model ensemble (C).



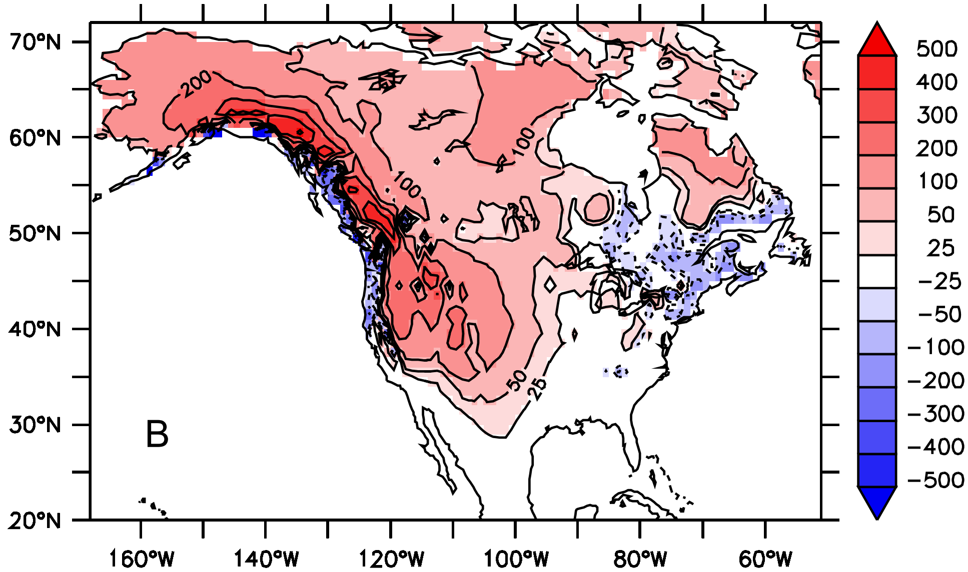


Figure 2.3. Multi-model ensemble climatology of annual snowfall minus (A) NorAm climatology; (B) WM climatology.

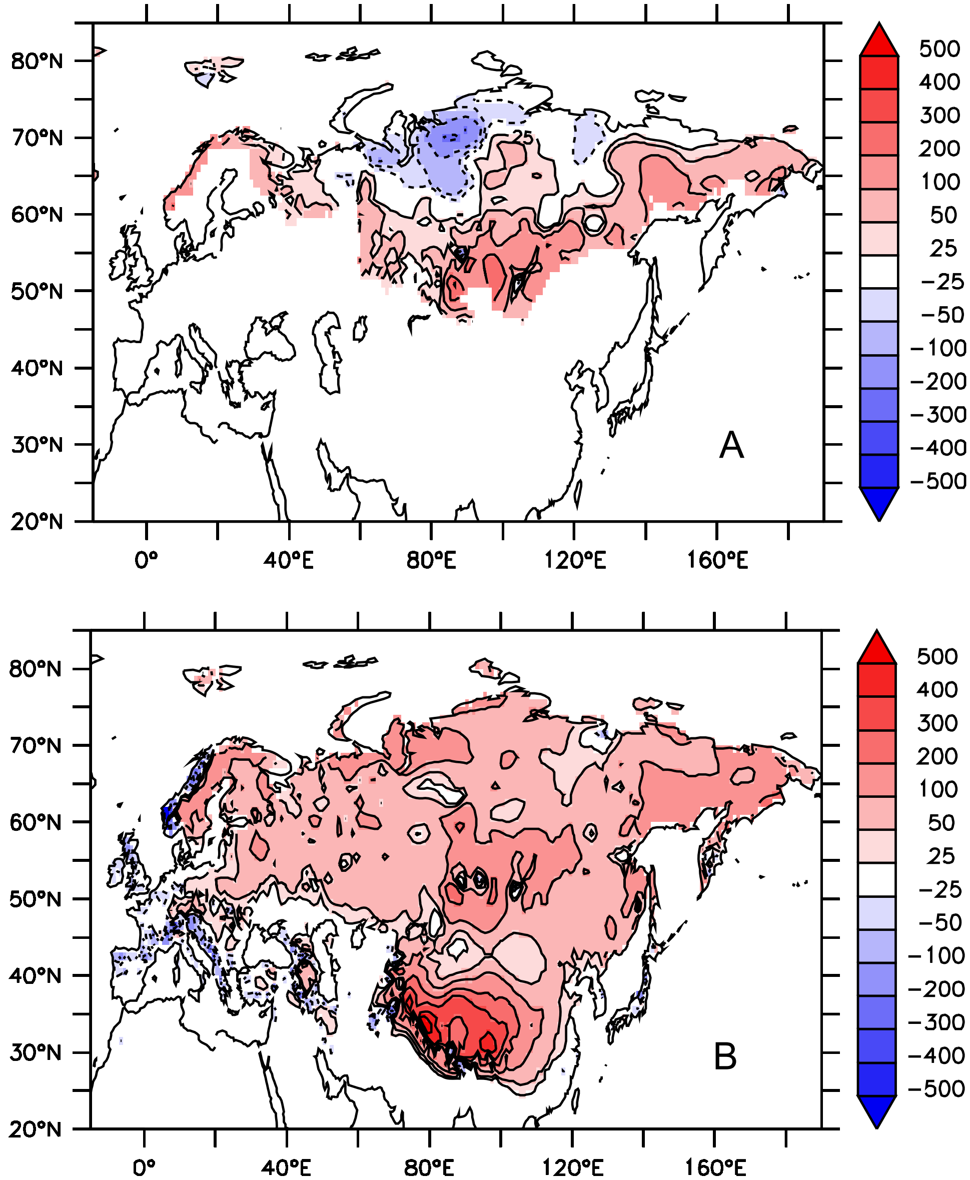


Figure 2.4. Difference (cm) between the multi-model ensemble climatology of annual snowfall and (A) TD-9813 climatology; (B) WM climatology.

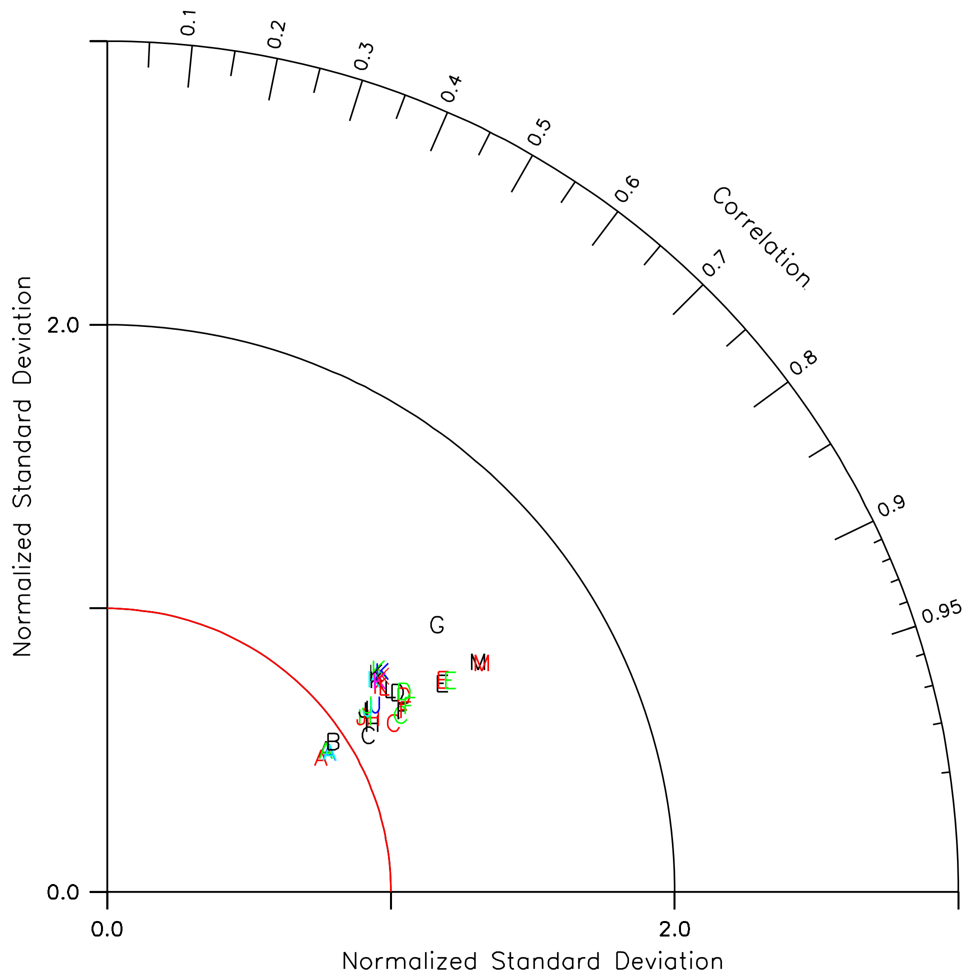


Figure 2.5: Taylor diagram of 1951 to 1999 annual snowfall, CMIP3 models versus NorAm observations east of the Rocky Mountains (105° W - 50° W, 25° N – 72° N). A - CGCM T47; B - CGCM T63, C - CSIRO Mk3.0, D - ECHAM5, E - GFDL CM2.0, F - GFDL CM2.1, G - INGV SXG, H - MIROC 3.2 Hi-Res., I - MIROC 3.2 Med-Res., J - MRI-CGCM, K - NCAR CCSM3, L - UKMO HadCM3, M - UKMO HadGEM1. Different colors represent individual simulations performed with the same model.

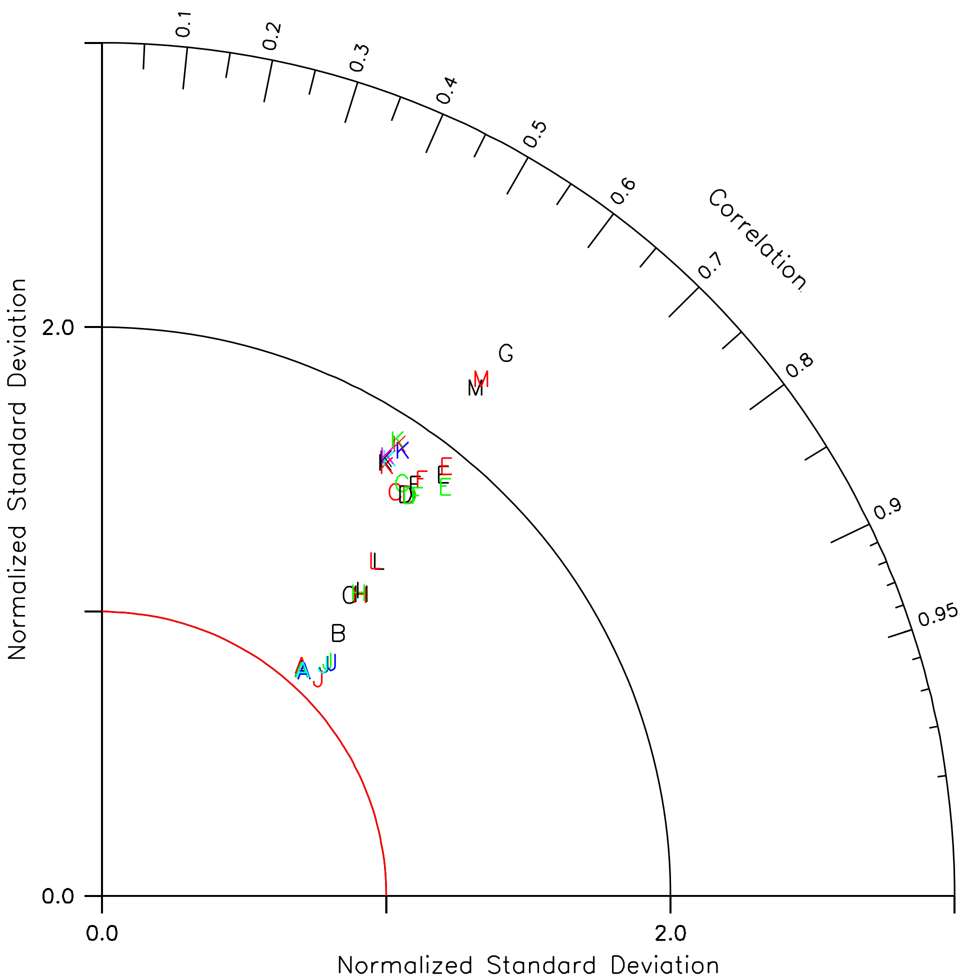


Figure 2.6: Same as Figure 2.5, except for the entire NorAm domain (165° W – 50° W, 25° N – 72° N) .

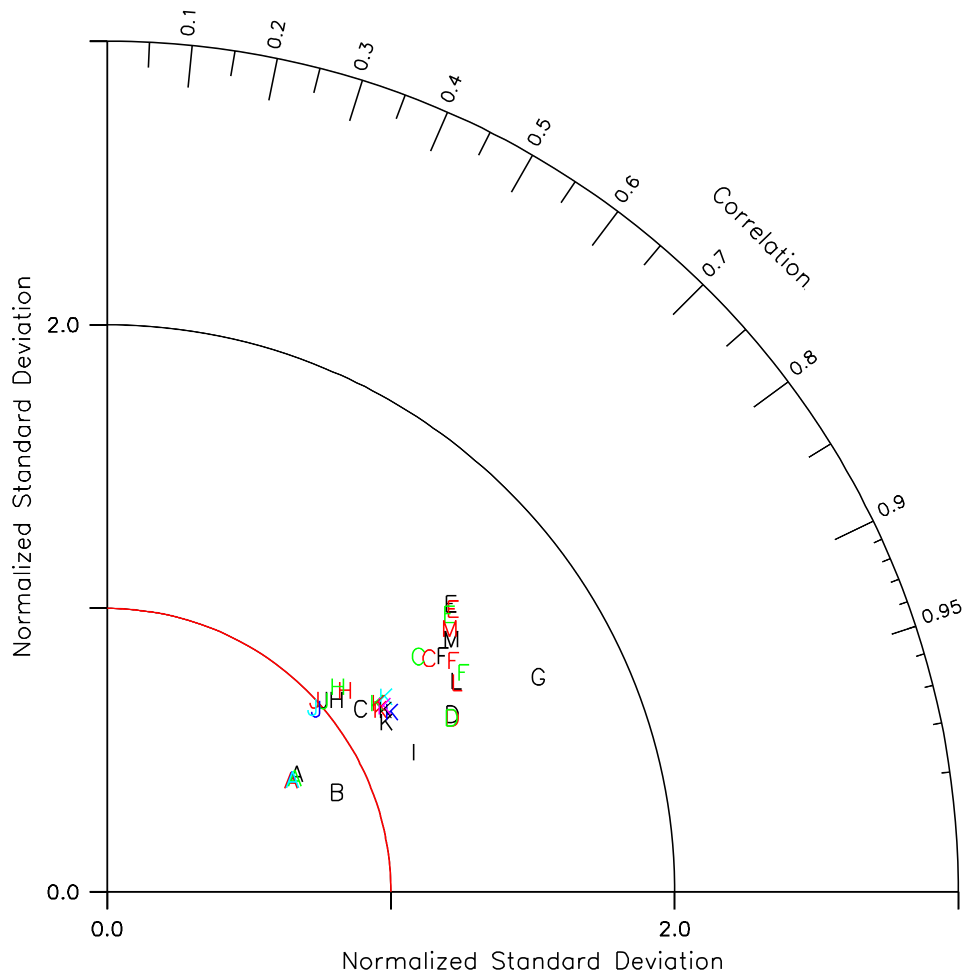


Figure 2.7: Same as Figure 2.5, except for CMIP3 models versus TD-9813 data.

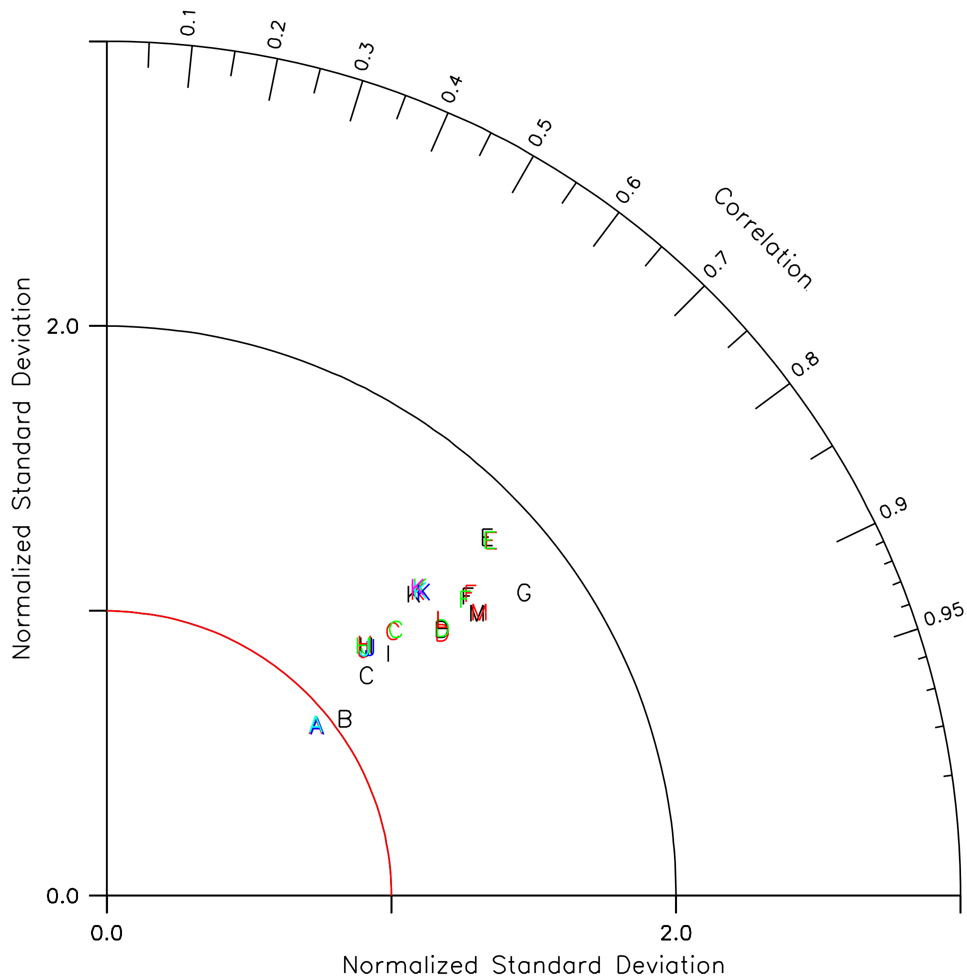


Figure 2.8: Same as Figure 2.5, except for CMIP3 models versus WM data.

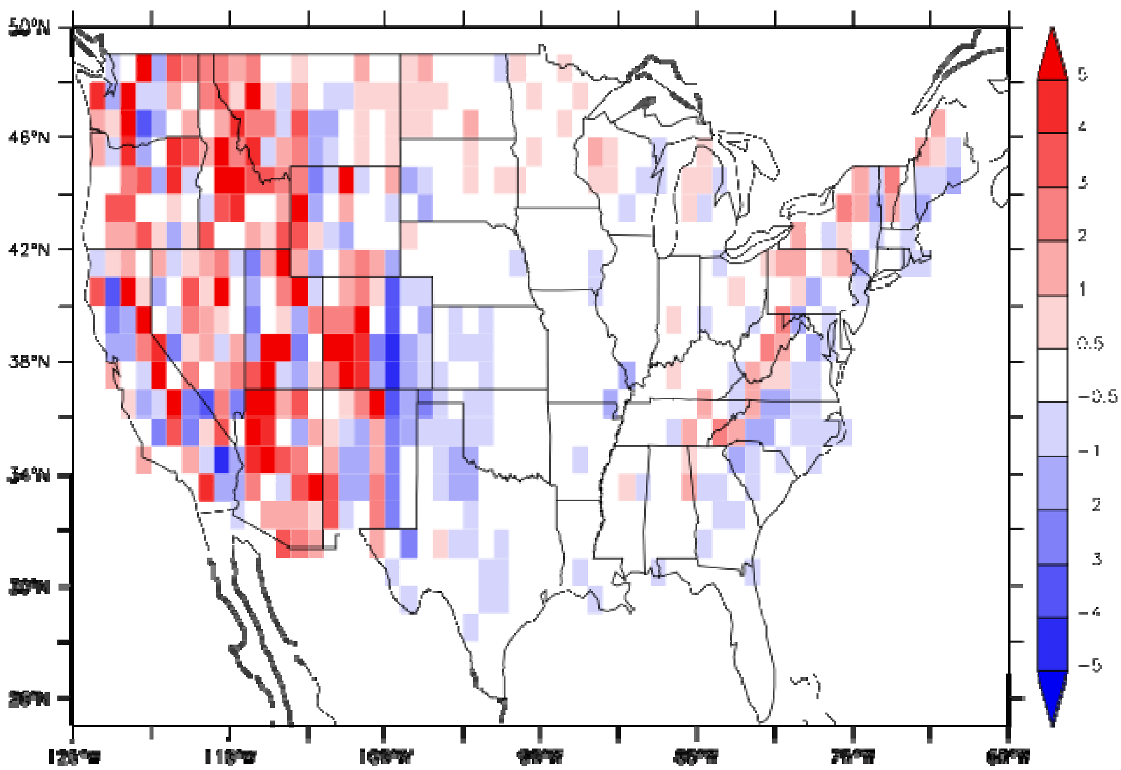


Figure 2.9. NorAm annual temperature (K) minus PRISM annual temperature (K), 1971 to 1999.

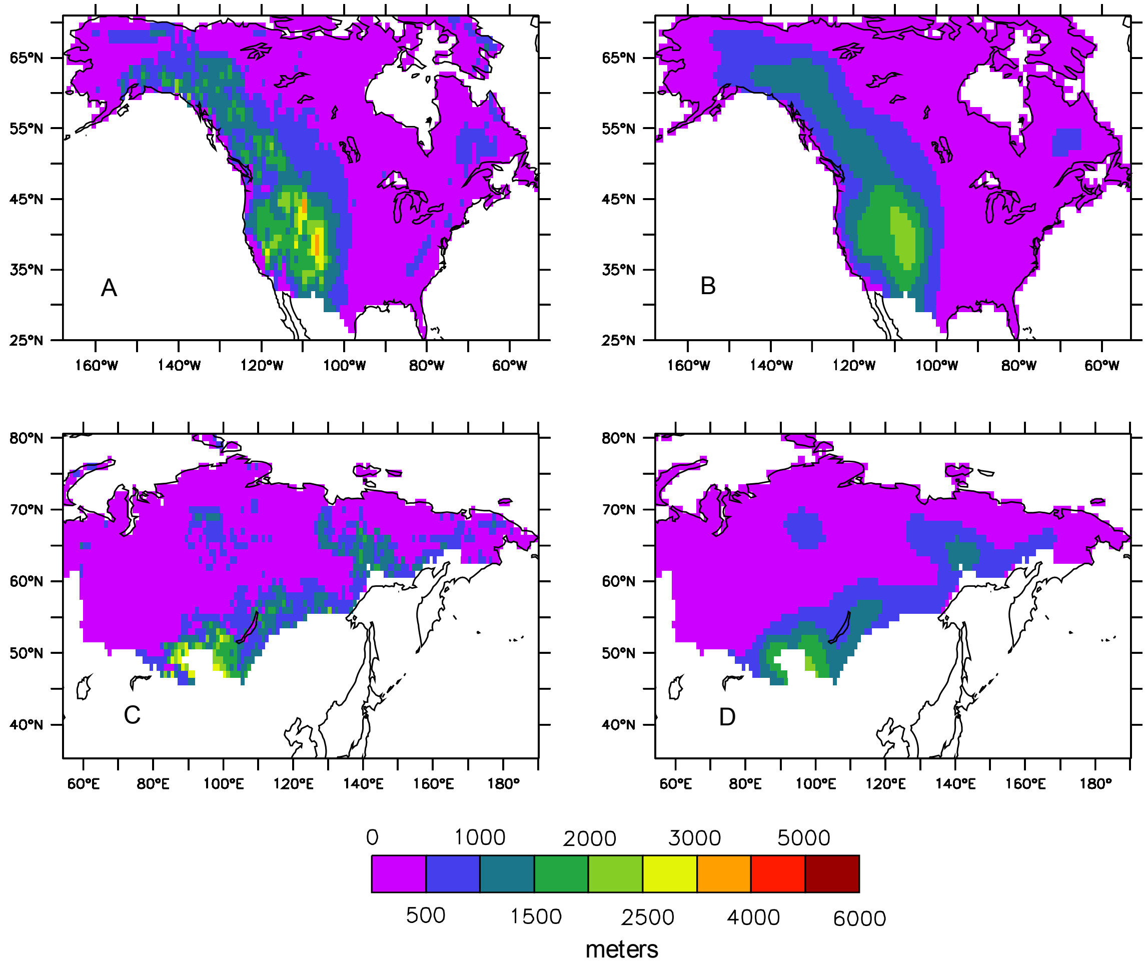


Figure 2.10: Comparison of model vs. actual topography (m). A.) USGS topography for NorAm domain; B.) model mean topography for NorAm domain; C.) USGS topography for TD-9813 domain; D.) model mean topography for TD-9813 domain.

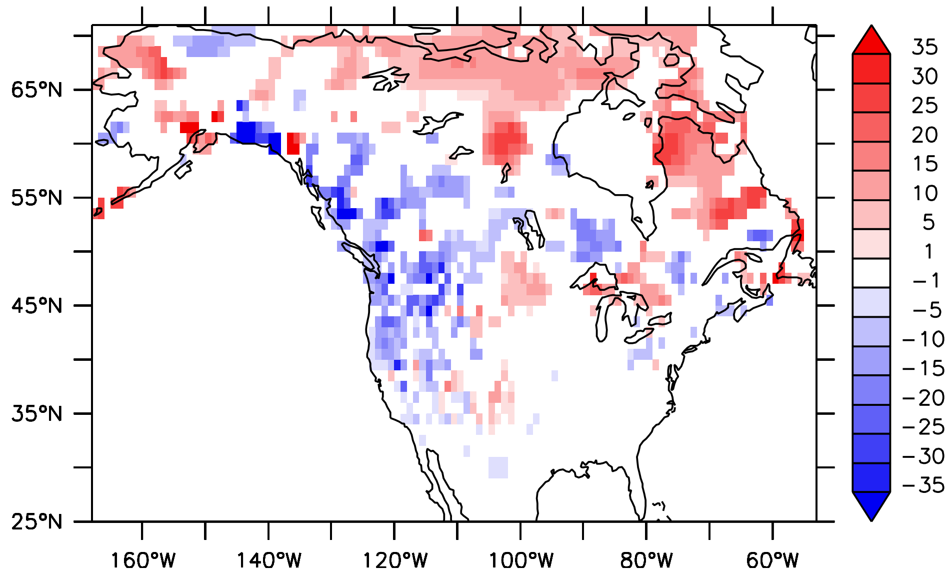


Figure 3.1. NorAm trends in observed annual snowfall, 1951 to 1999. Units are in cm/decade. Only statistically significant trends (p = 0.05) are shaded.

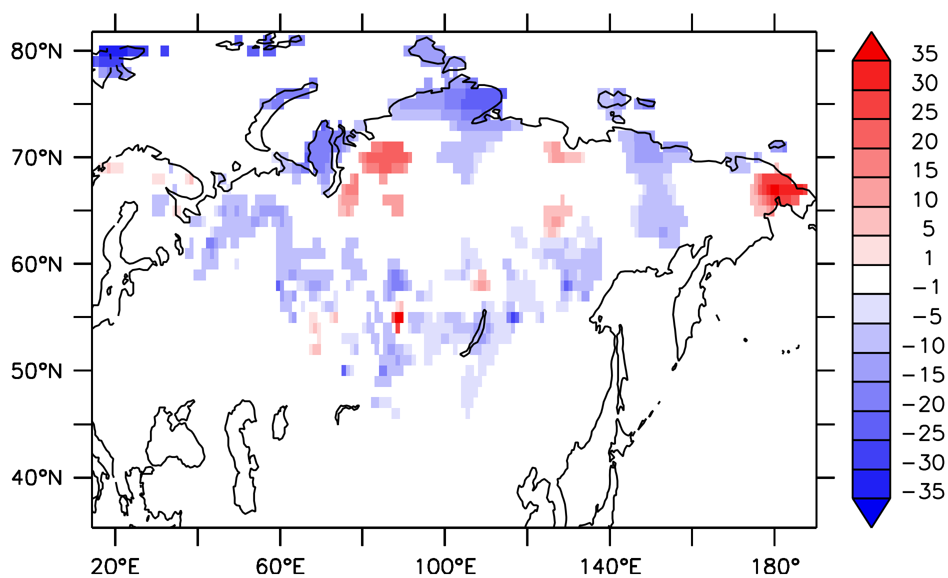


Figure 3.2. TD-9813 trends in derived annual snowfall, 1951 to 1999. Units are in cm/decade. Only statistically significant trends (p = 0.05) are shaded.

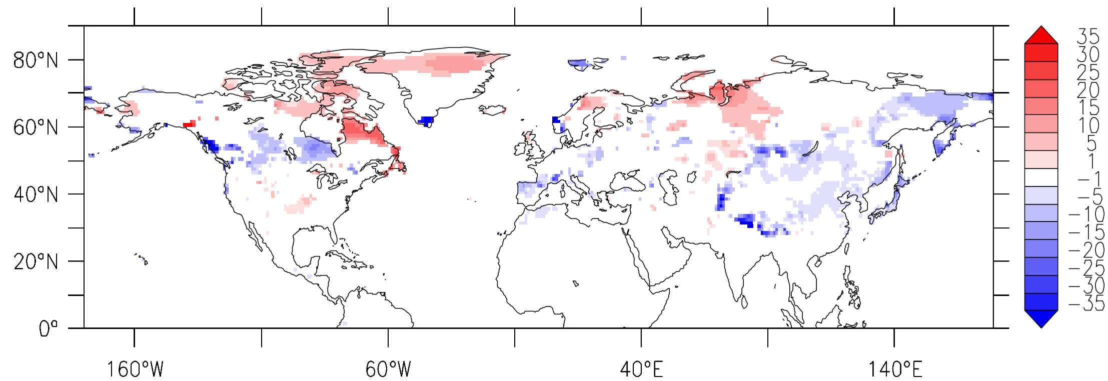


Figure 3.3. WM trends in derived annual snowfall, 1951 to 1999. Units are in cm/decade. Only statistically significant trends (p = 0.05) are shaded.

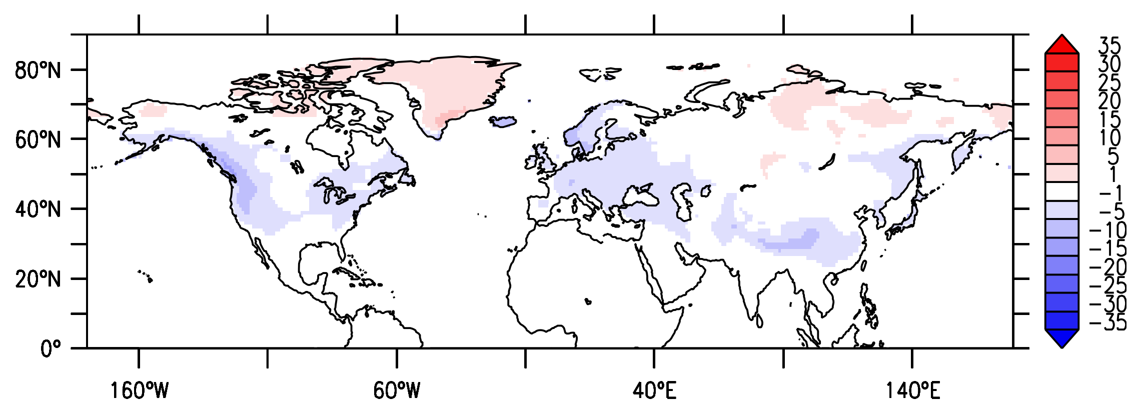


Figure 3.4. Multi-model ensemble mean trends in annual snowfall, 1951 to 1999. Units are in cm/decade.

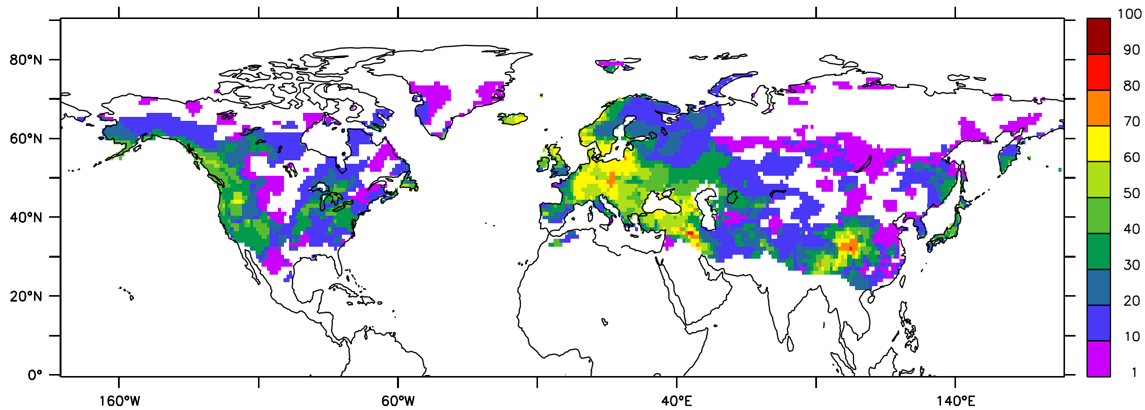


Figure 3.5: Percentage of individual model simulations showing statistically significant negative trends in annual snowfall. Locations receiving less than 5 cm of annual snowfall are masked.

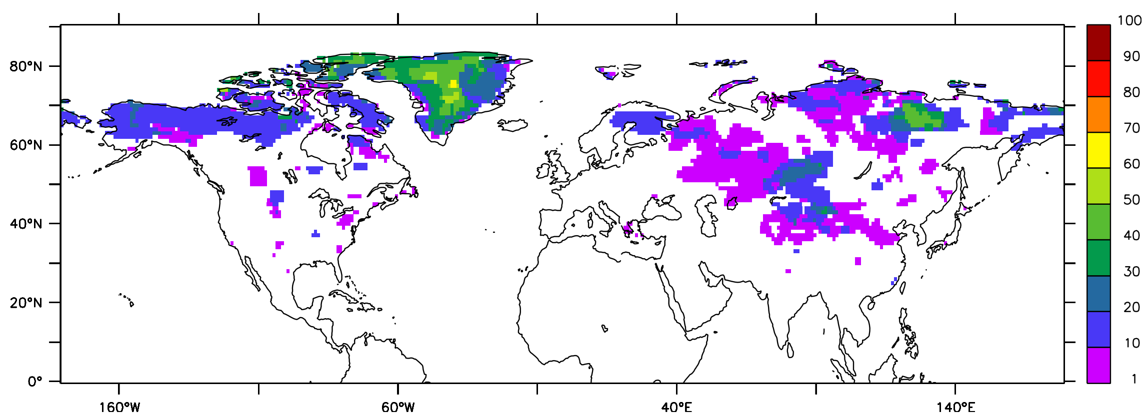


Figure 3.6: Percentage of individual model simulations showing statistically significant positive trends in annual snowfall. Locations receiving less than 5 cm of annual snowfall are masked.

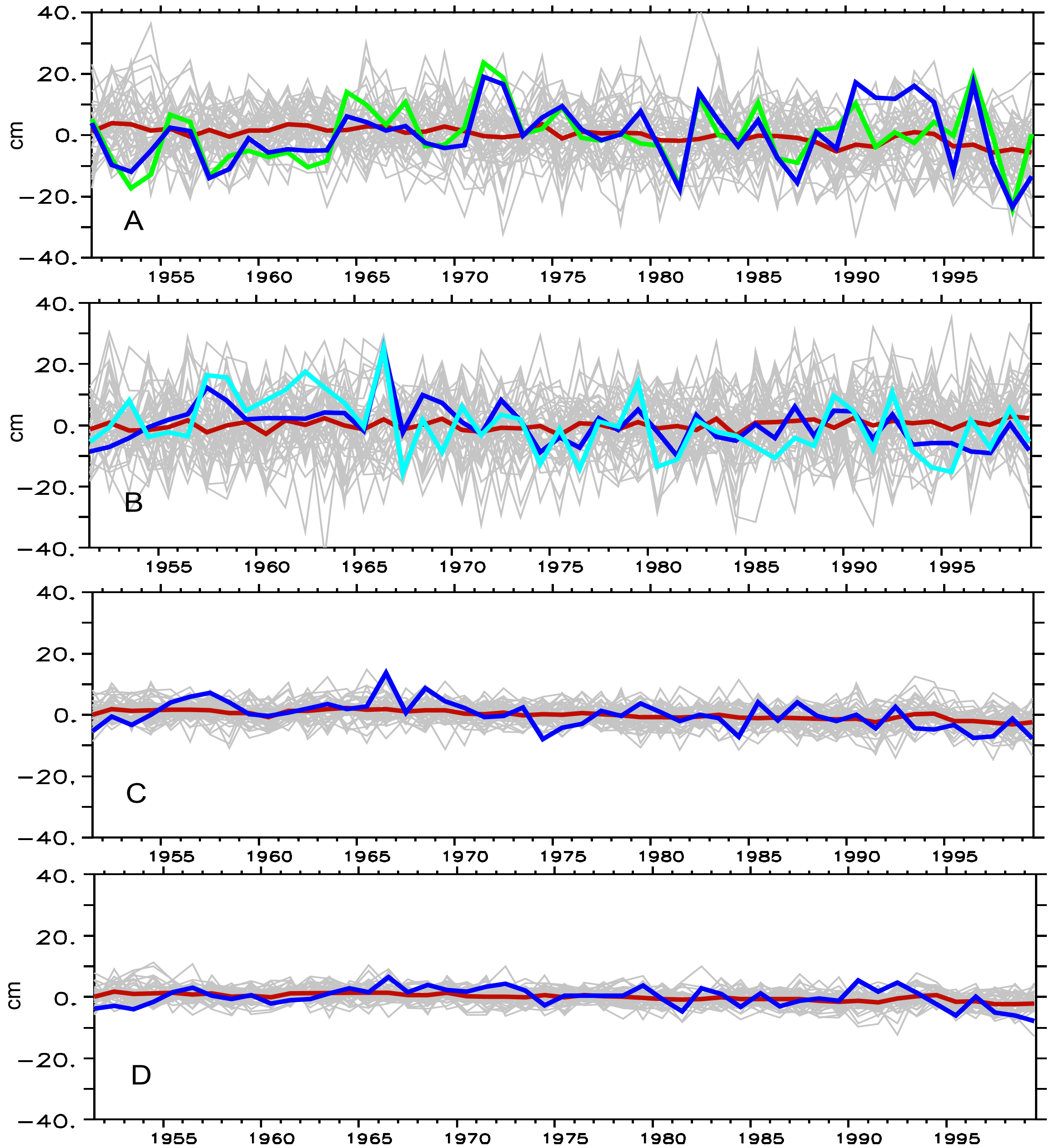


Figure 3.7. Time series of annual snowfall anomalies (based on 1951 to 1999 mean) for North America (A), Eurasia (B), TD-9813 domain (C), and the Northern Hemisphere (D). Individual model simulations (gray), multi-model ensemble (red), Willmott-Matsuura derived snowfall (blue), NorAm observations (green), and TD-9813 derived snowfall (cyan) are shown.

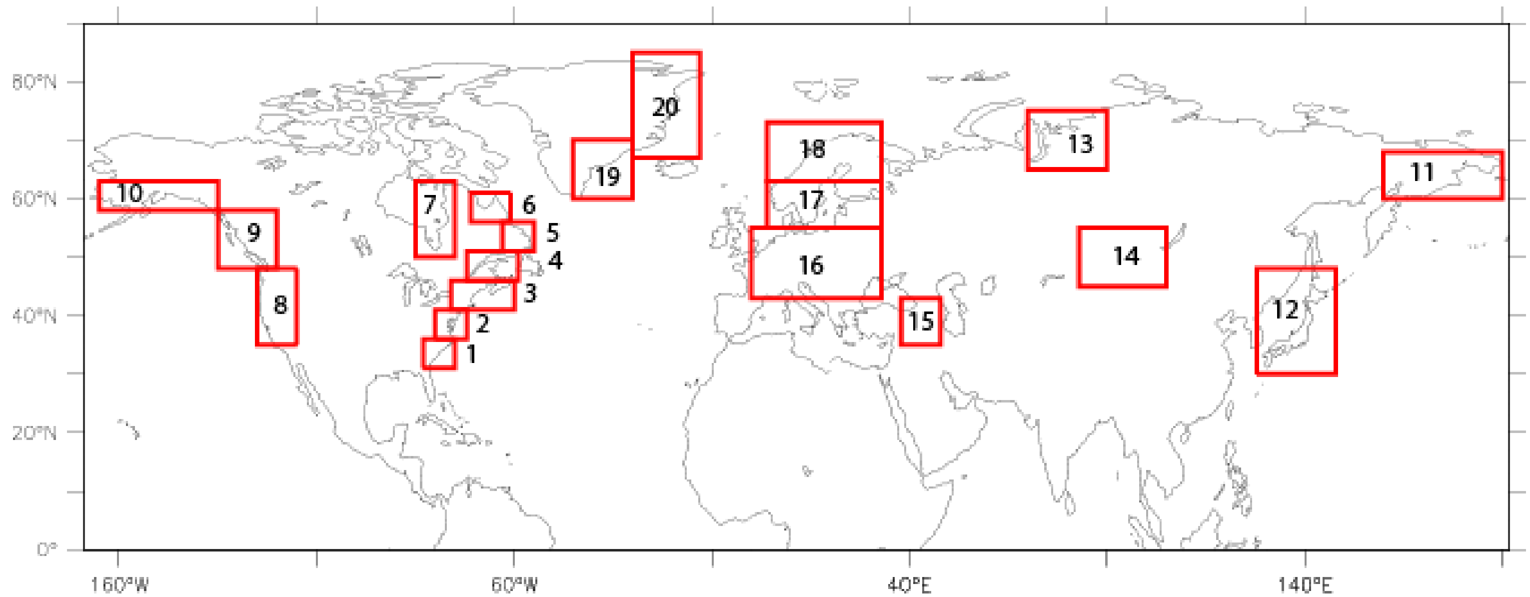


Figure 3.8. Northern Hemisphere regions used in this study.

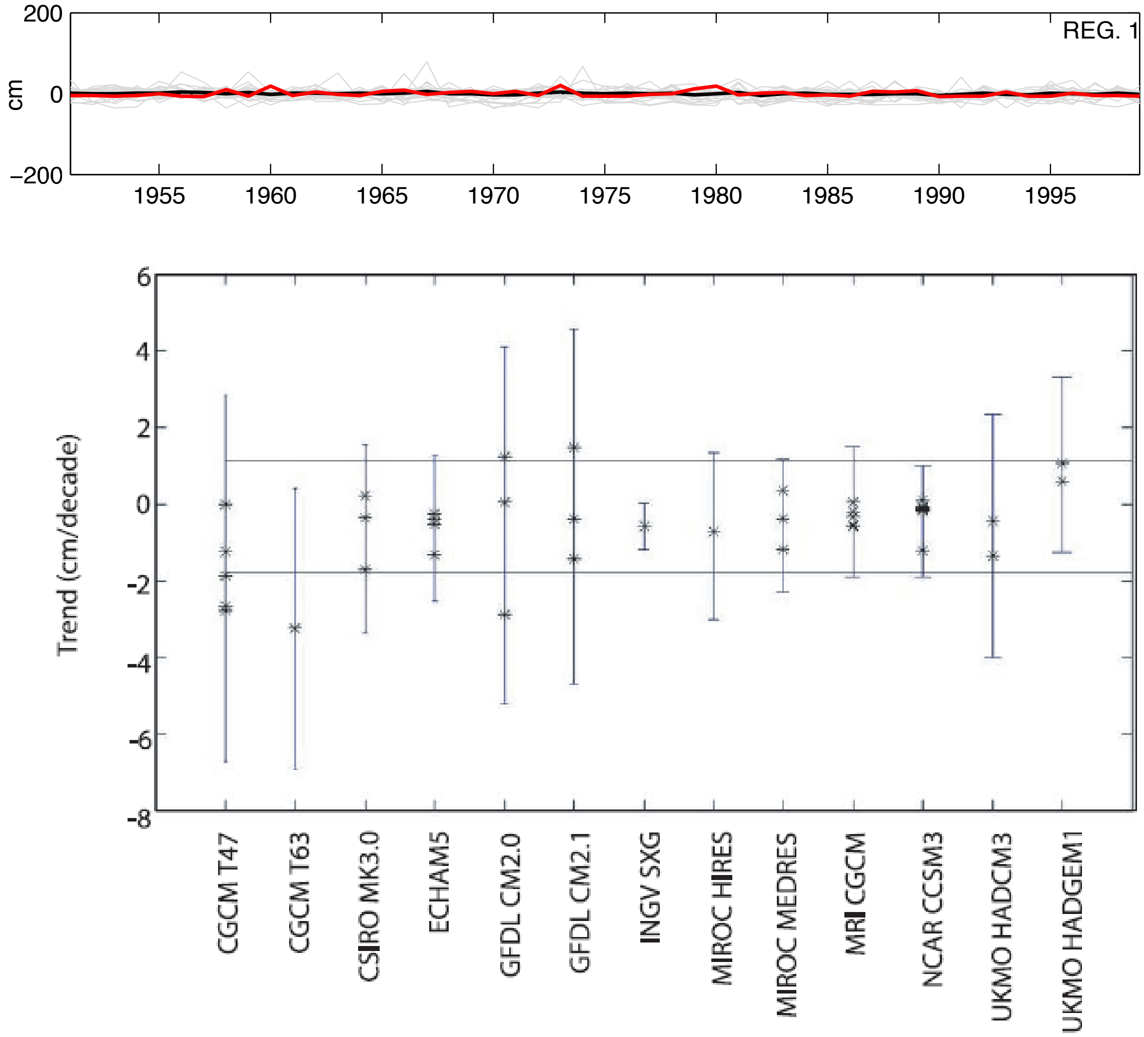


Figure 3.9. Region 1. a.) Time series of annual snow anomalies for individual simulations *gray*, multi-model ensemble *black*, and observations, *red*. b.) Linear trends for each of the model simulations (markers), and observed trend (middle horizontal line). 95% confidence intervals denoted by whiskers (models), and upper/lower horizontal line (observations). Stars indicate simulated trends are statistically the same as the observed trend (p=0.05), open circles indicate statistically different trends.

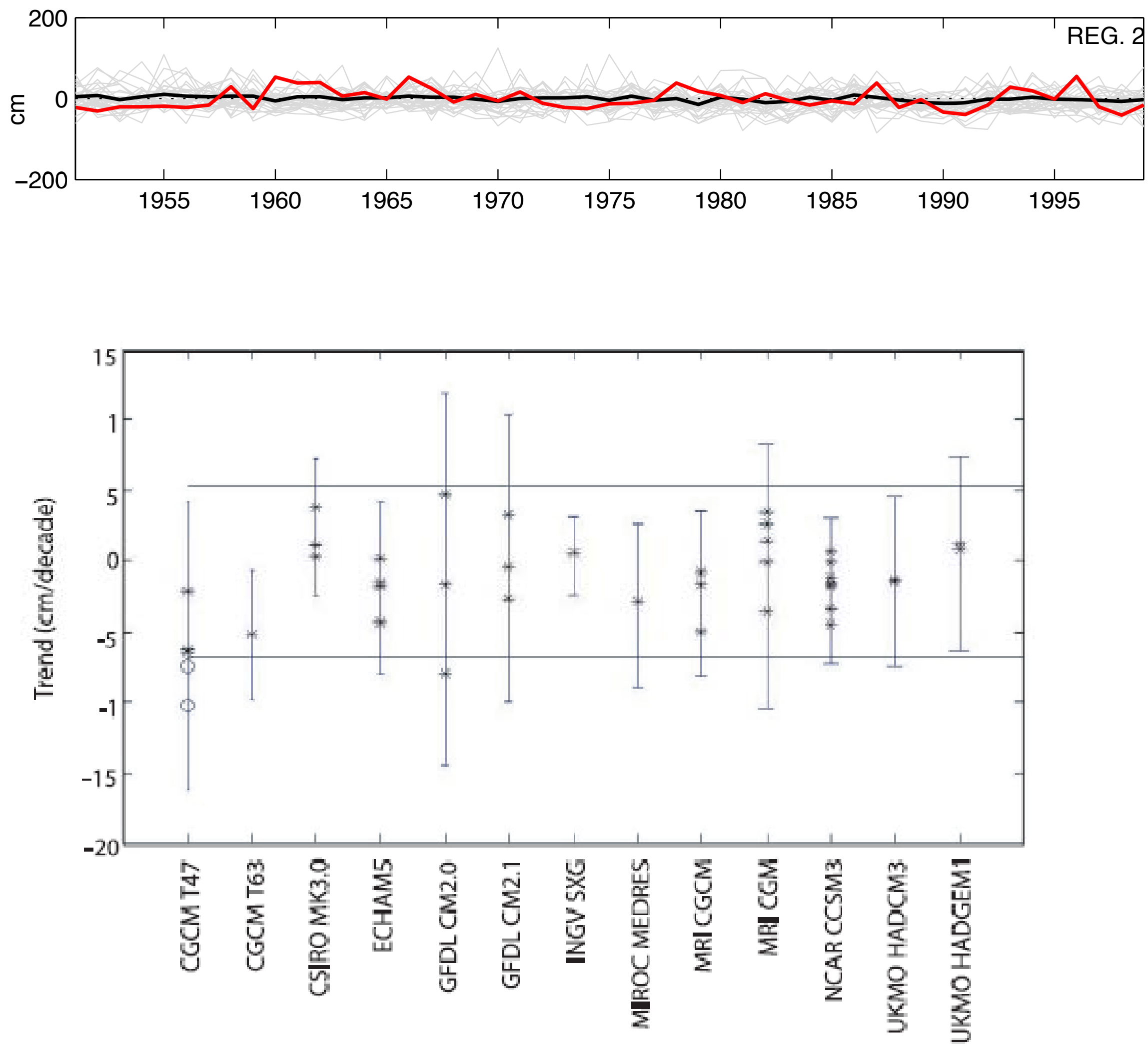


Figure 3.10. Same as Figure 3.8, except for Region 2.

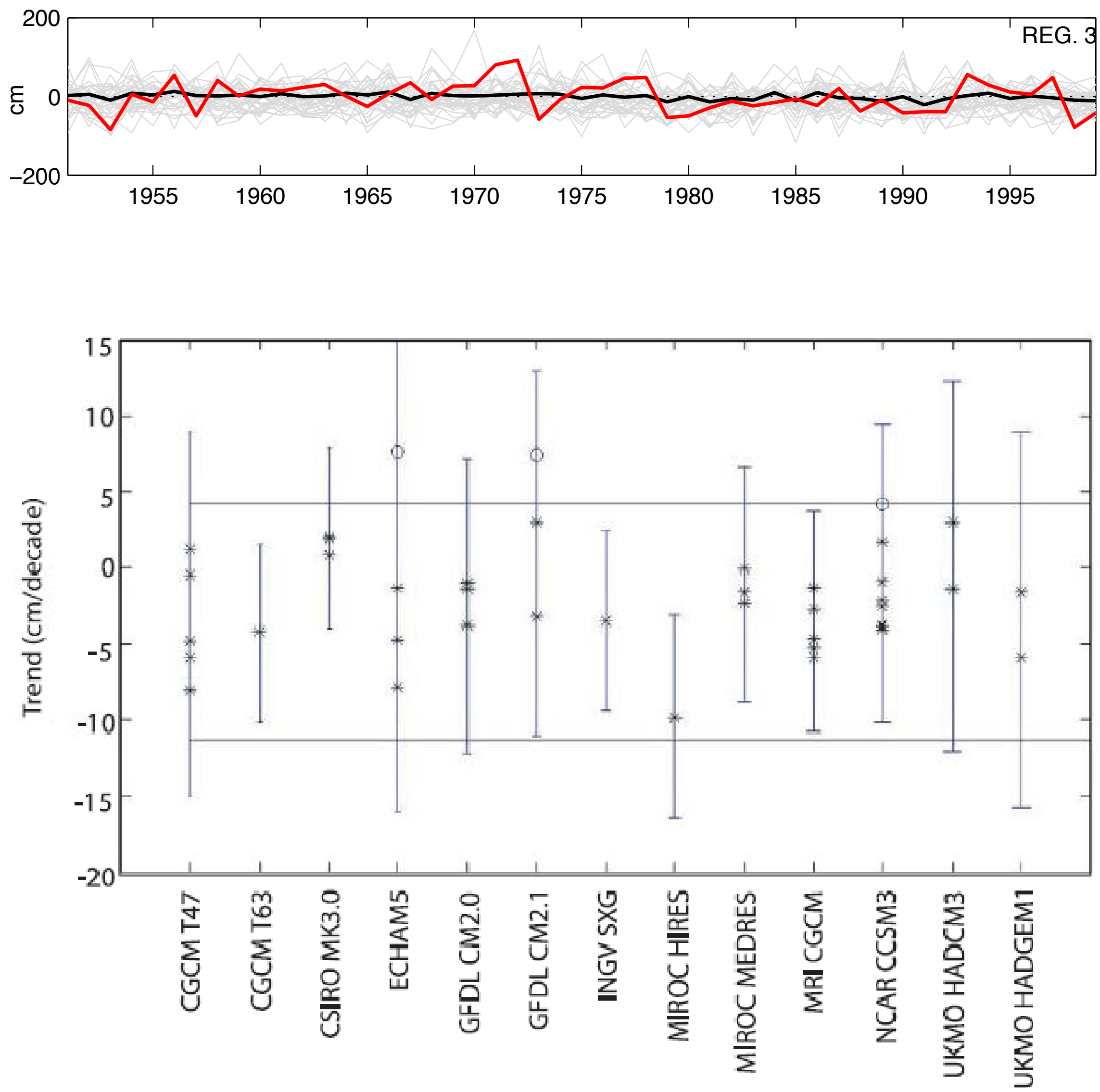


Figure 3.11. Same as Figure 3.8, except for Region 3.

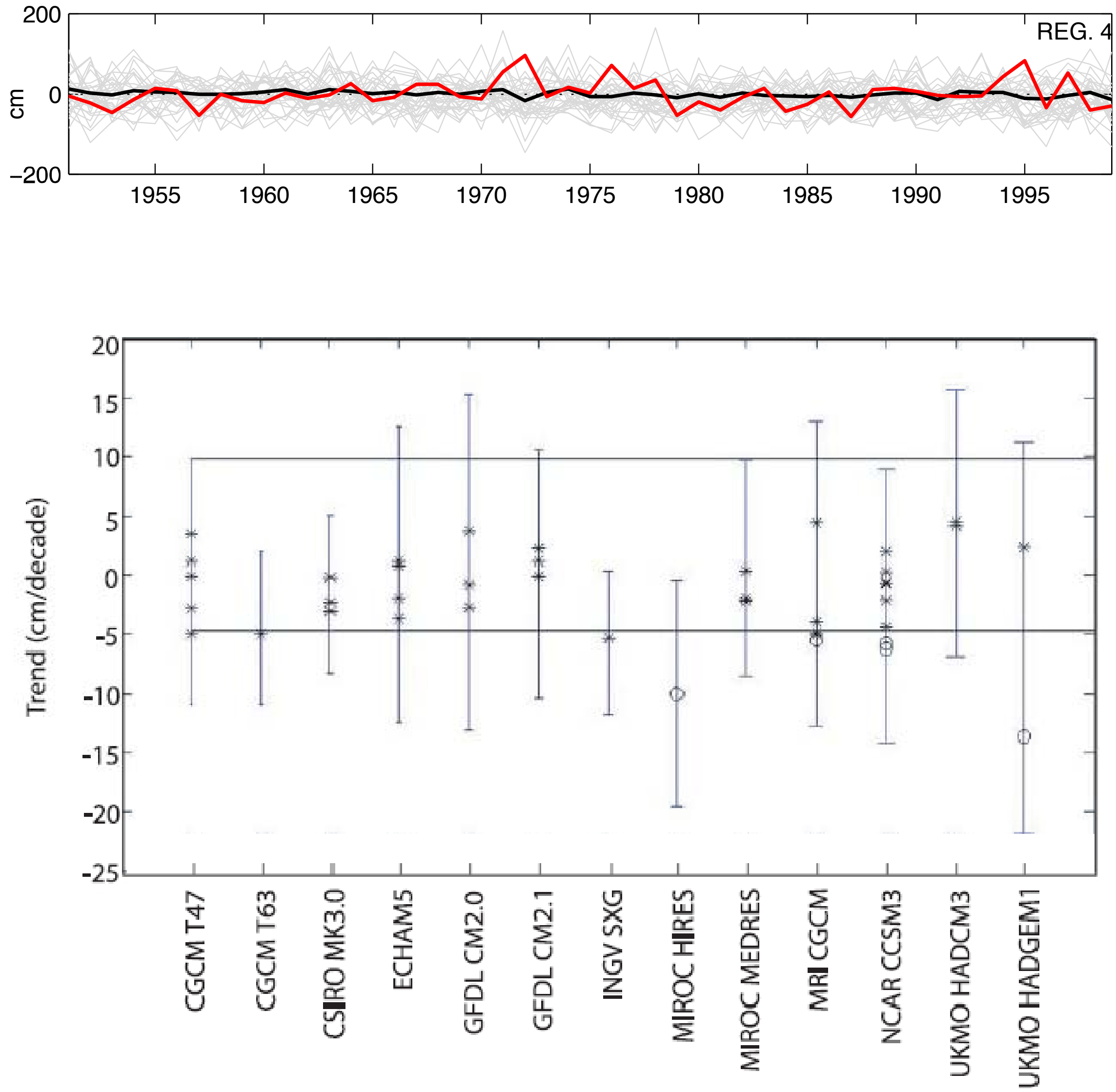


Figure 3.12. Same as Figure 3.8, except for Region 4.

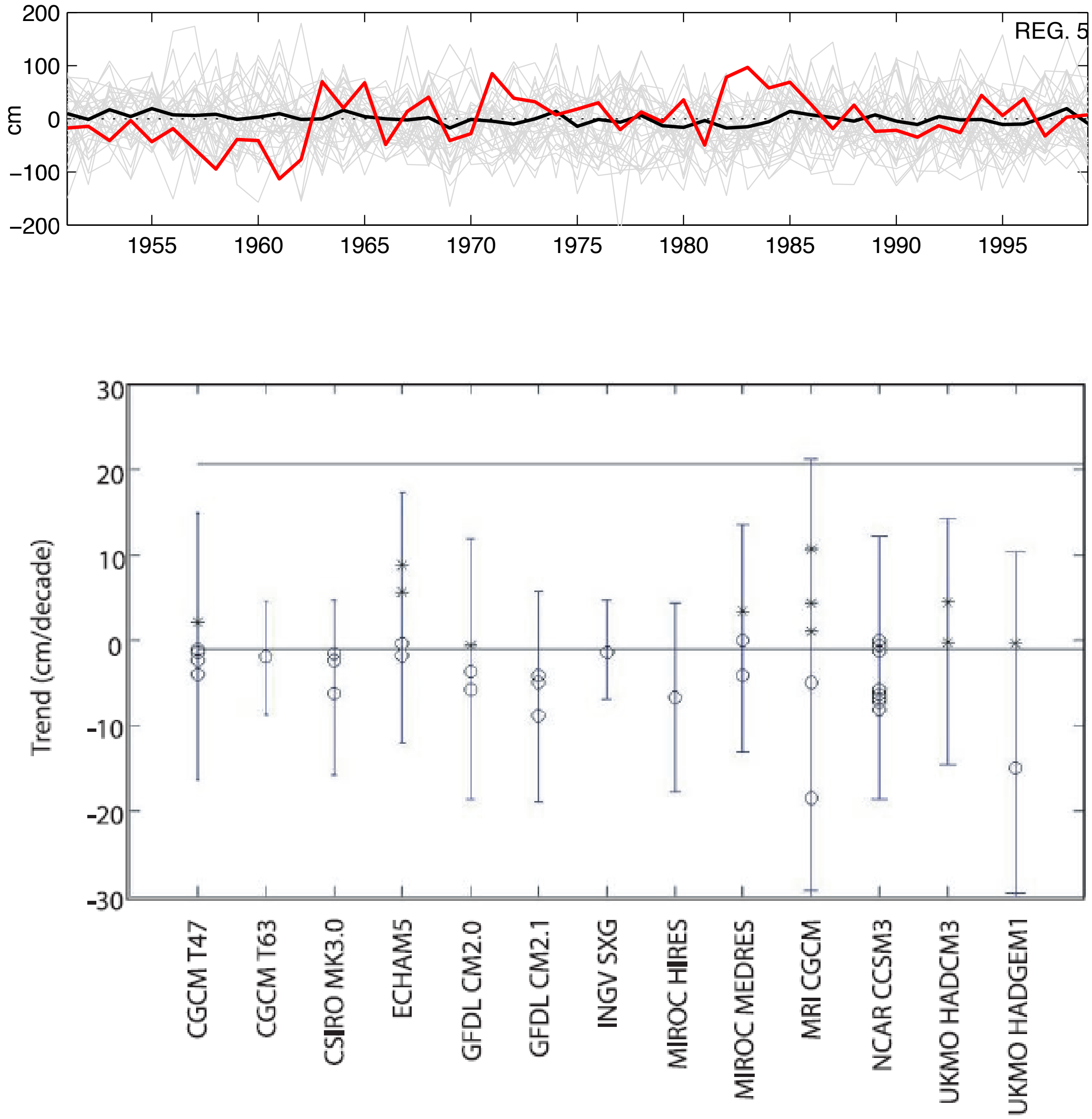


Figure 3.13. Same as Figure 3.8, except for Region 5.

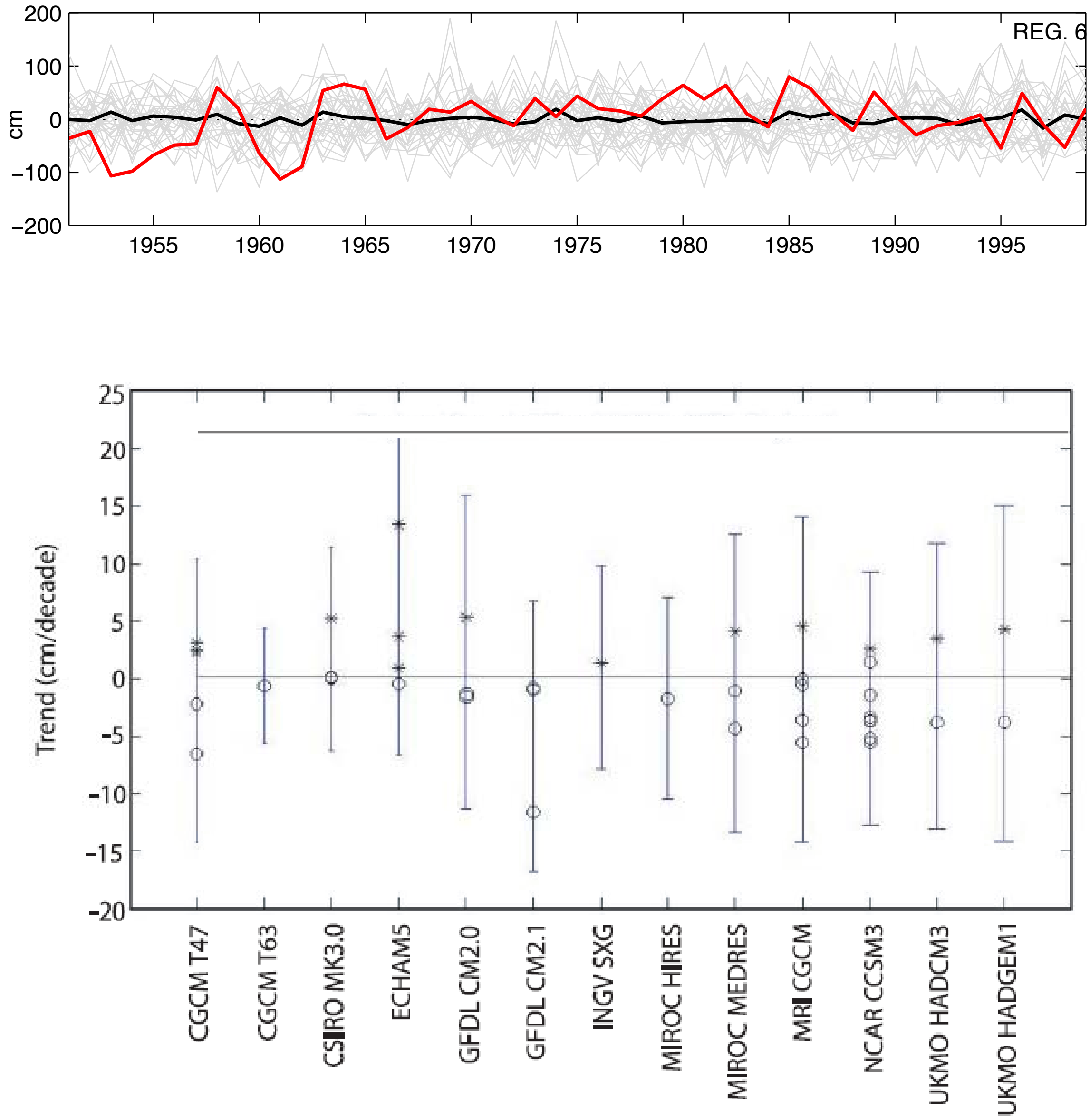
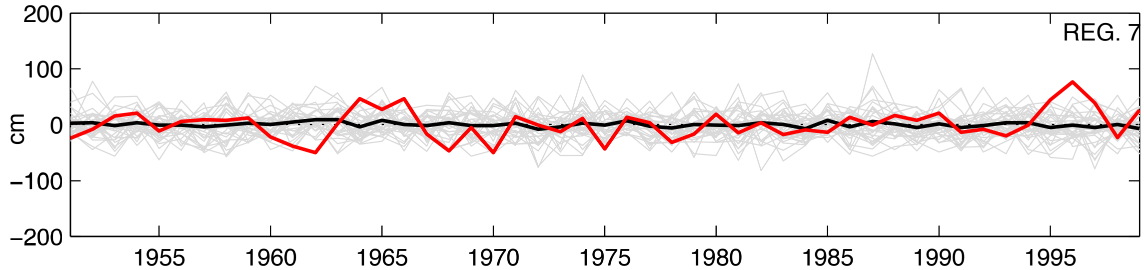


Figure 3.14. Same as Figure 3.8, except for Region 6.



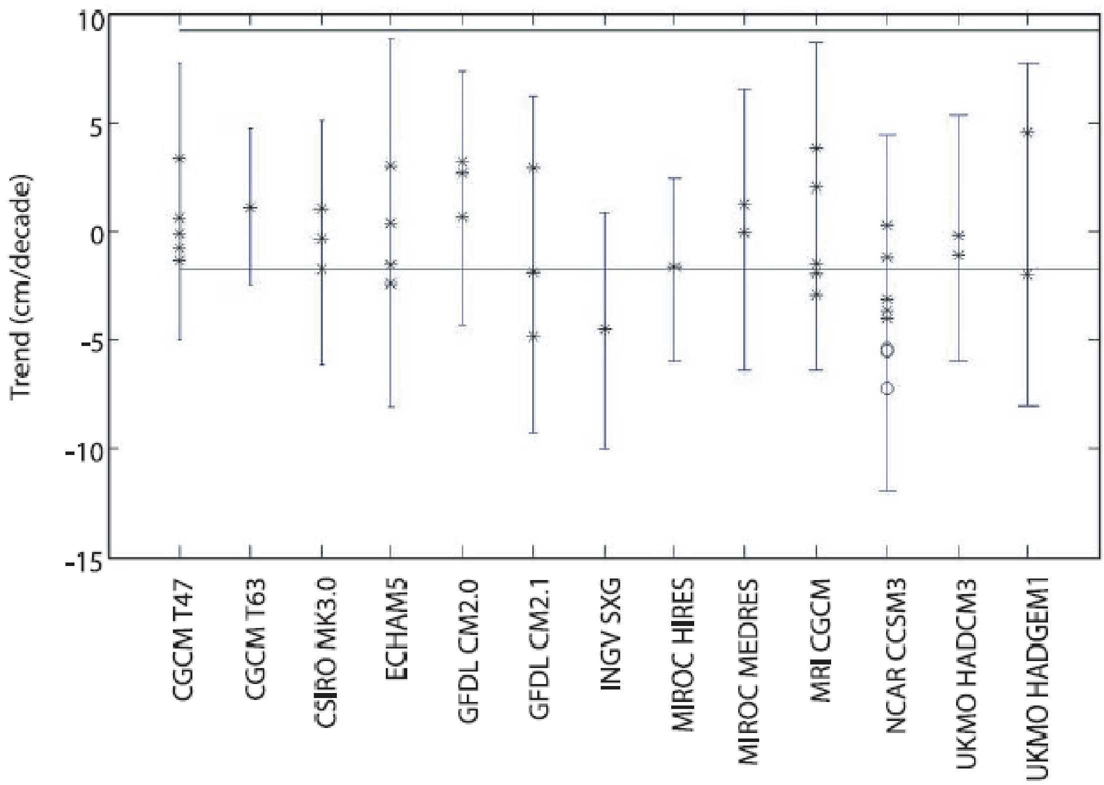
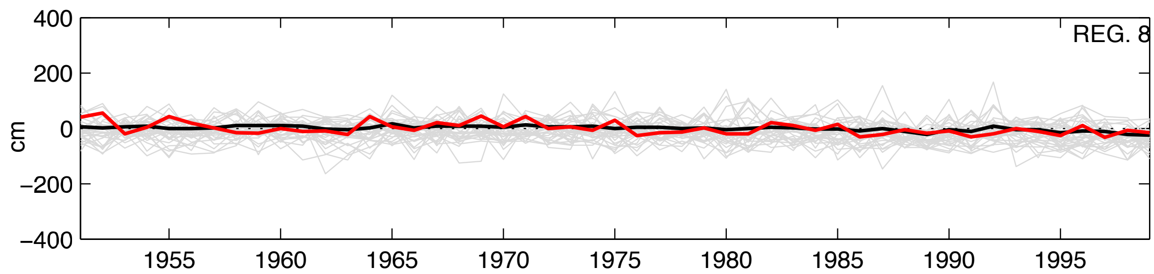


Figure 3.15. Same as Figure 3.8, except for Region 7.



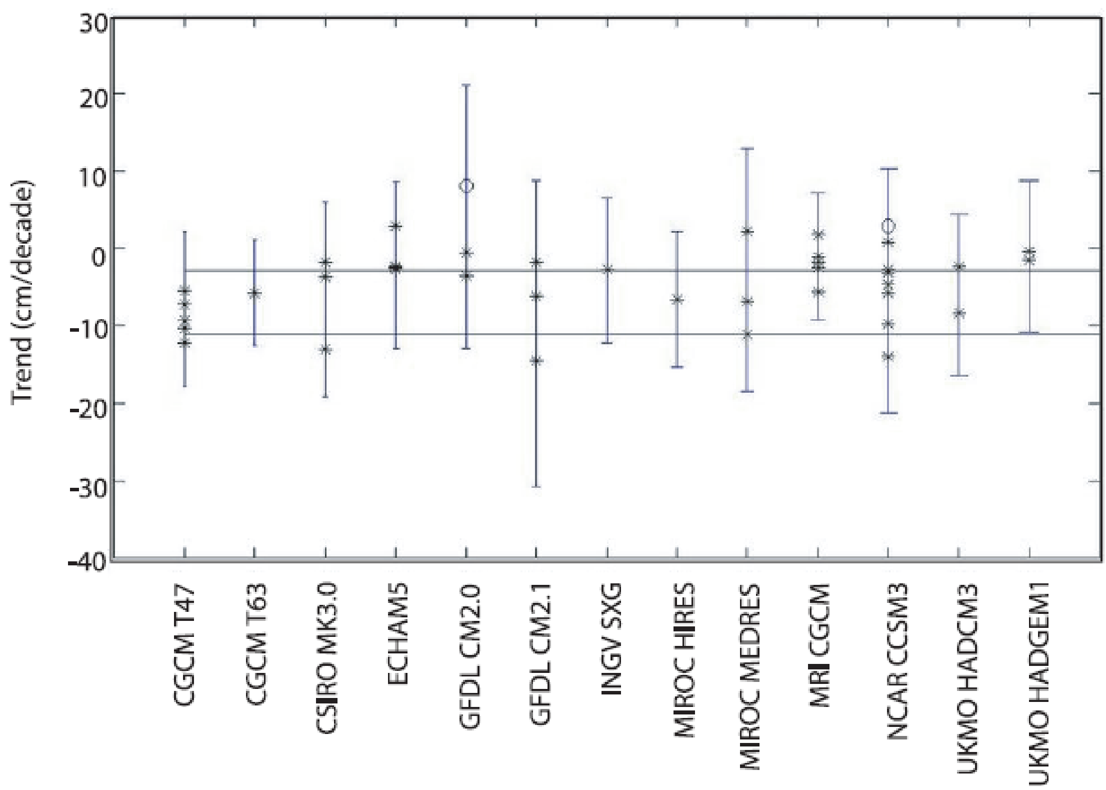


Figure 3.16. Same as Figure 3.8, except for Region 8.

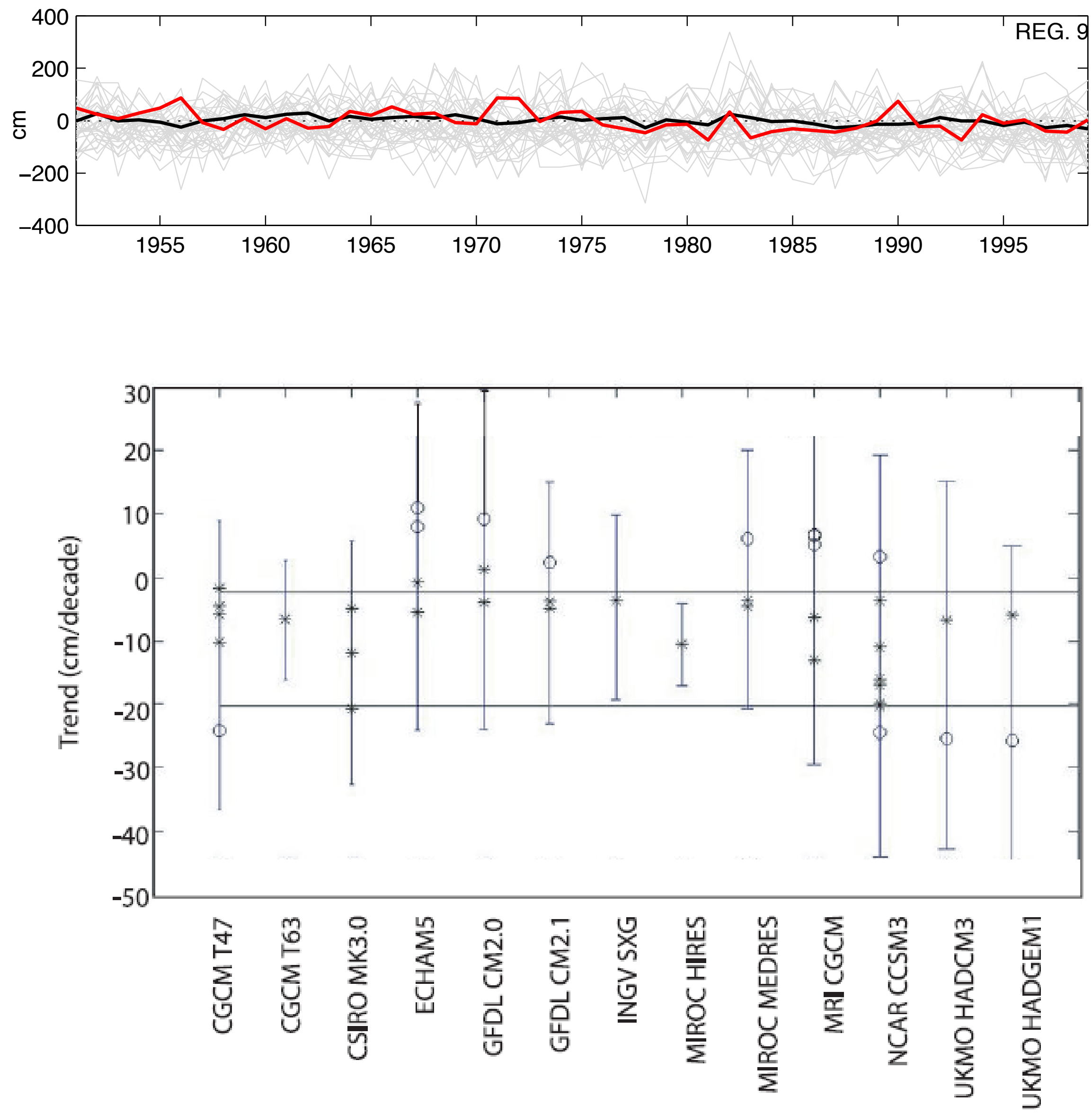
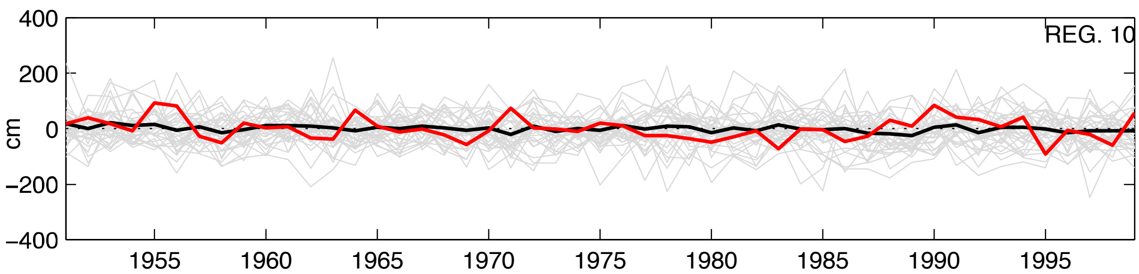


Figure 3.17. Same as Figure 3.8, except for Region 9.



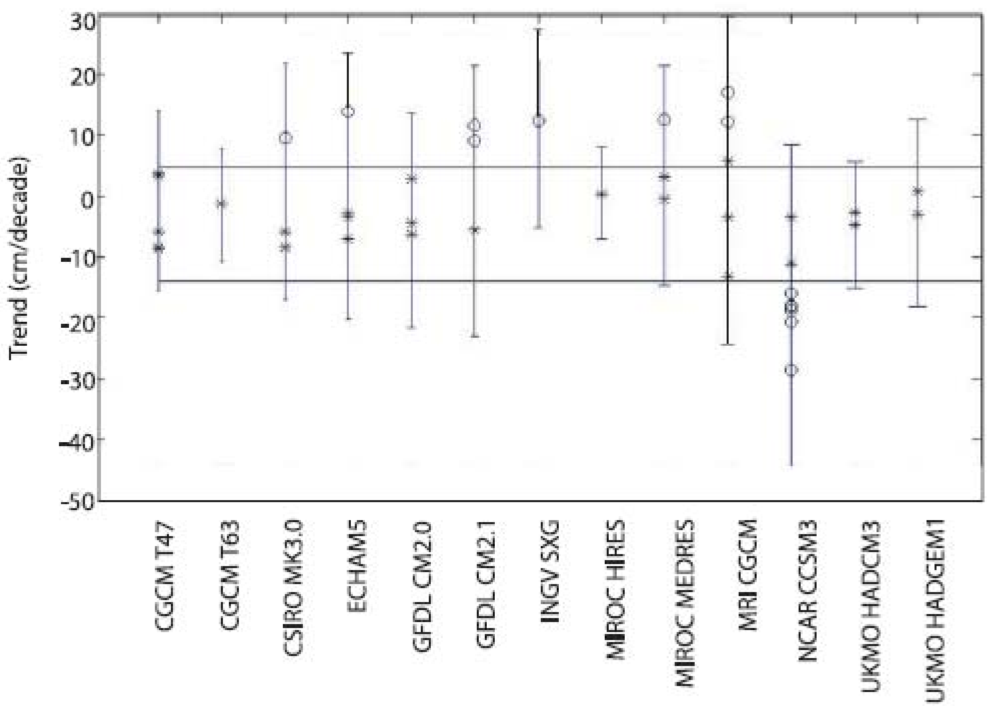
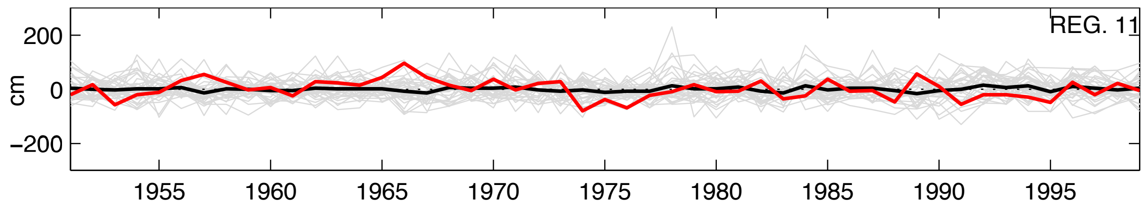


Figure 3.18. Same as Figure 3.8, except for Region 10.



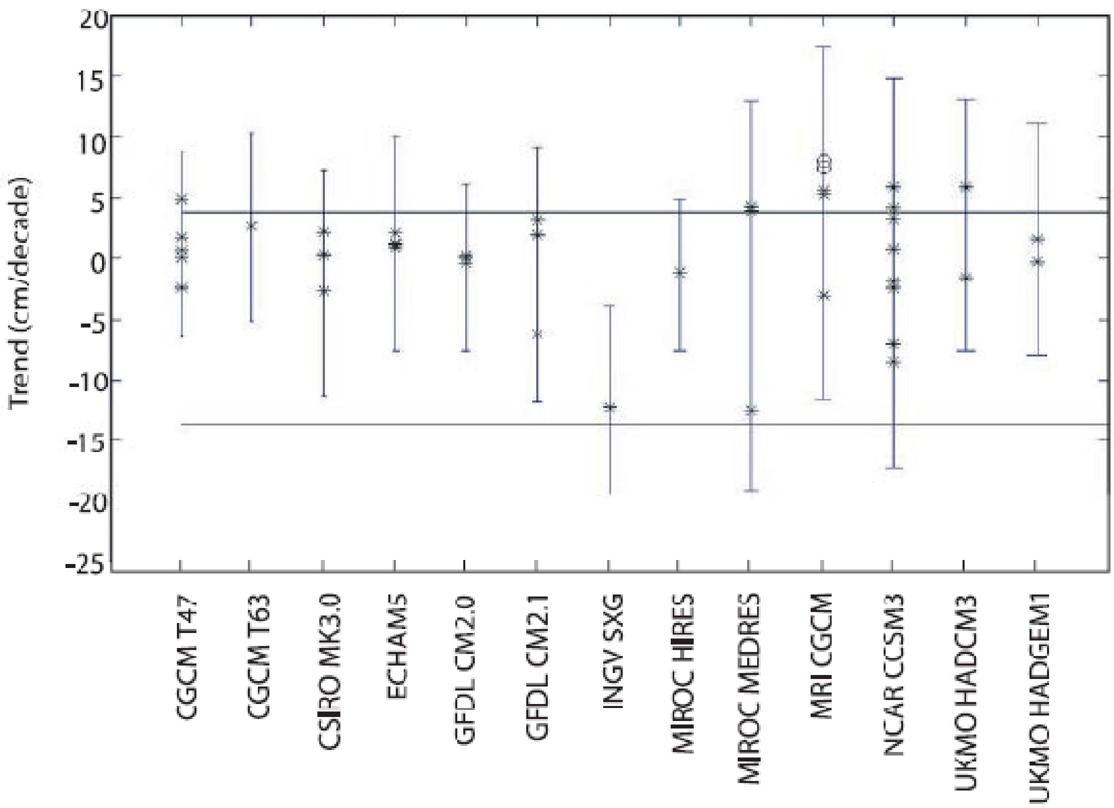
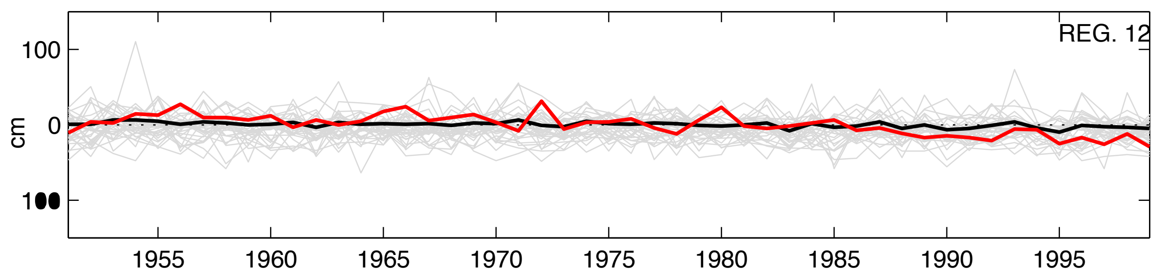


Figure 3.19. Same as Figure 3.8, except for Region 11.



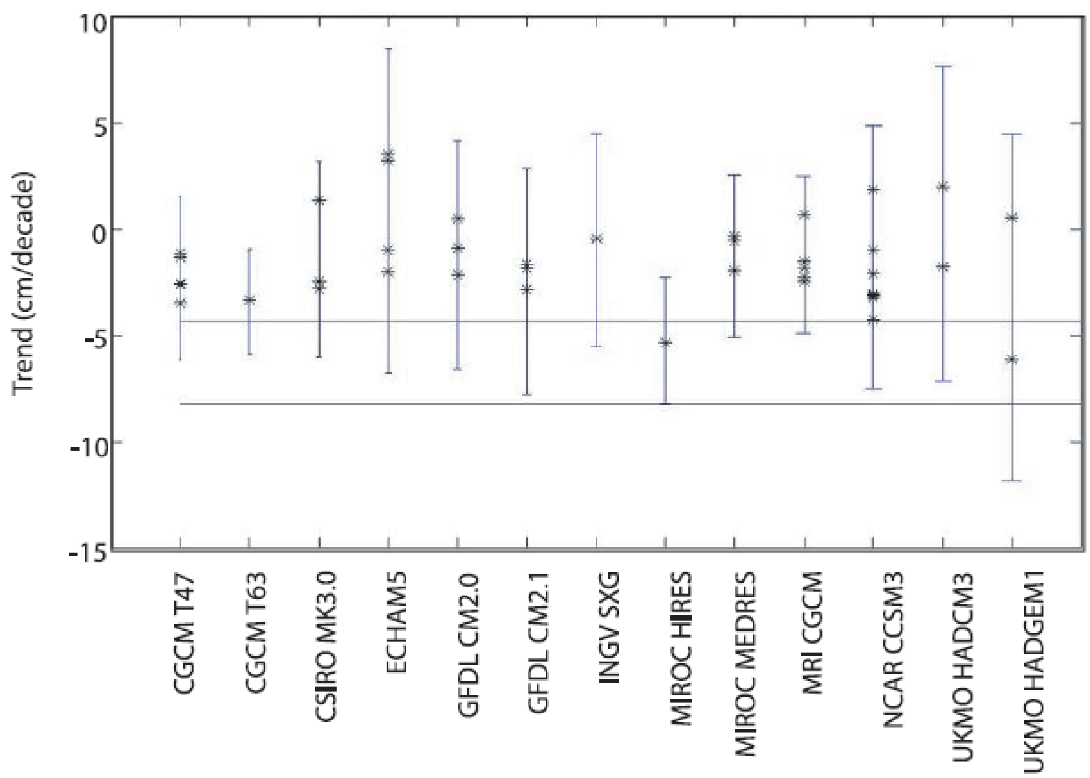
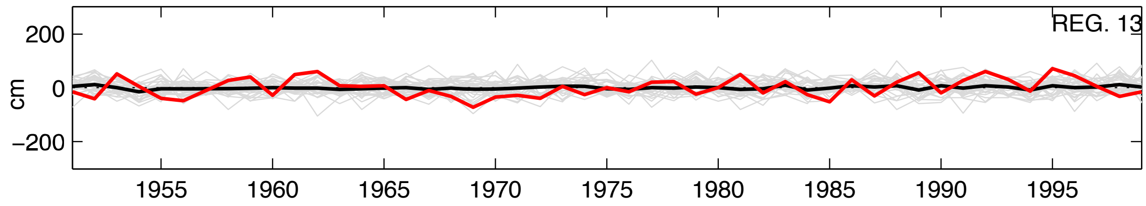


Figure 3.20. Same as Figure 3.8, except for Region 12.



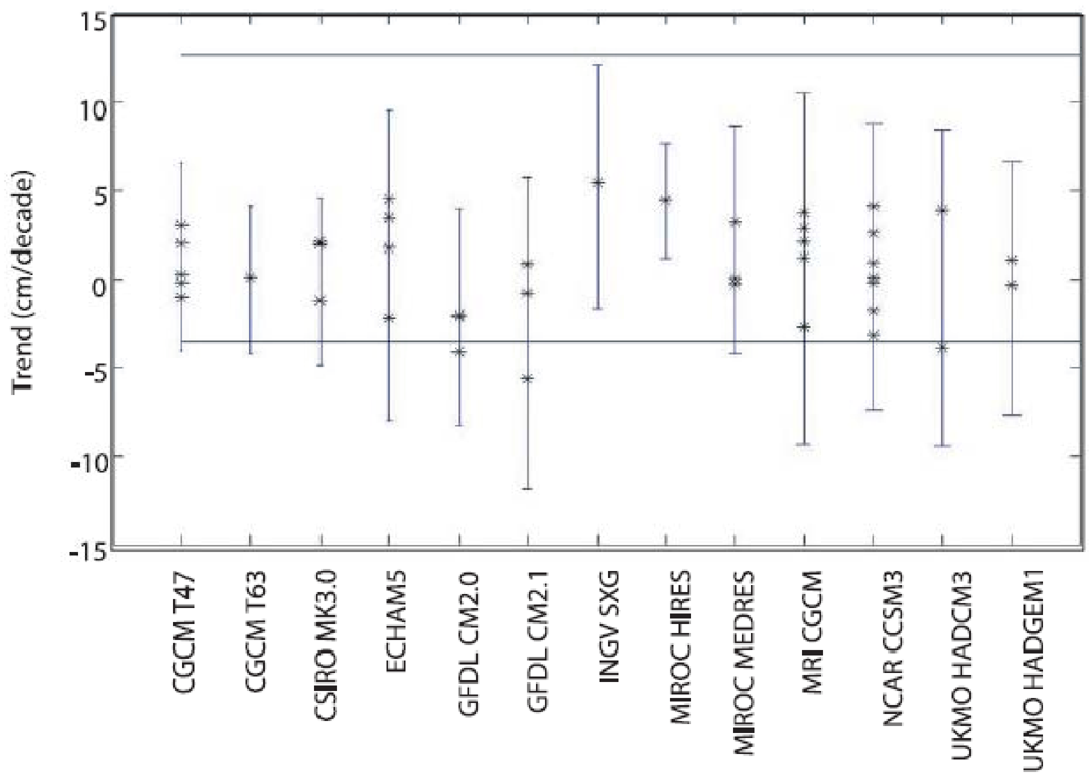
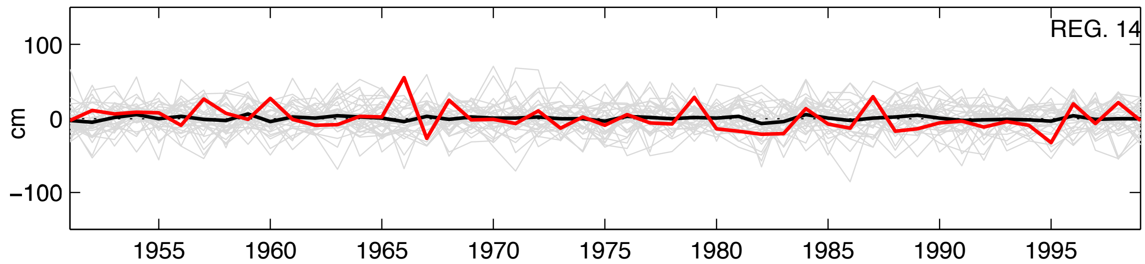


Figure 3.21. Same as Figure 3.8, except for Region 13.



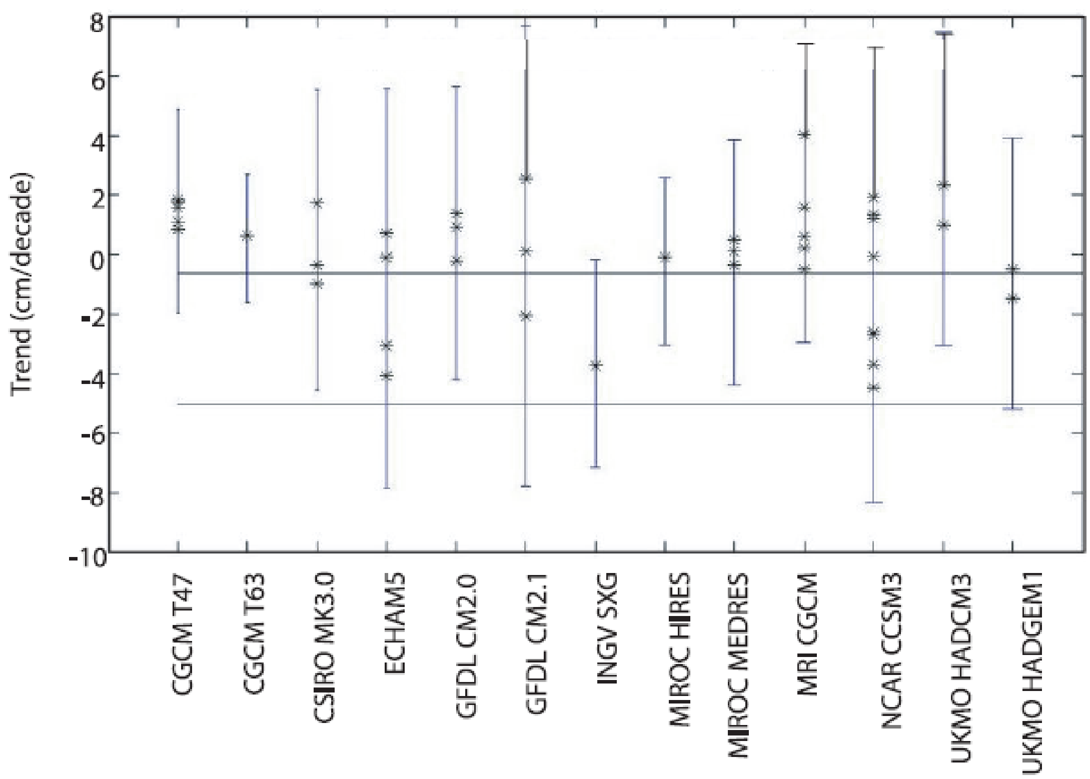
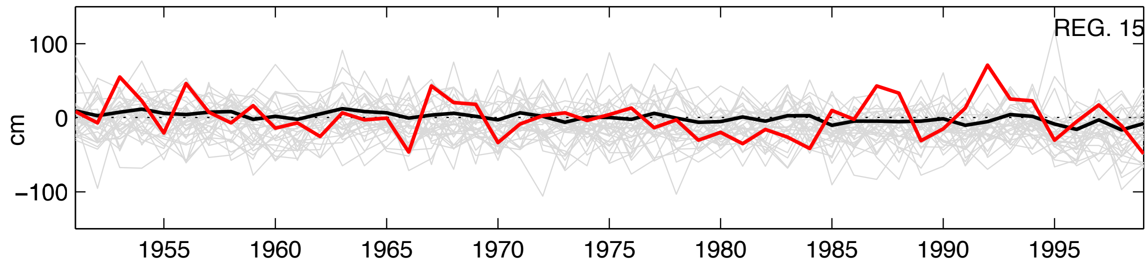


Figure 3.22. Same as Figure 3.8, except for Region 14.



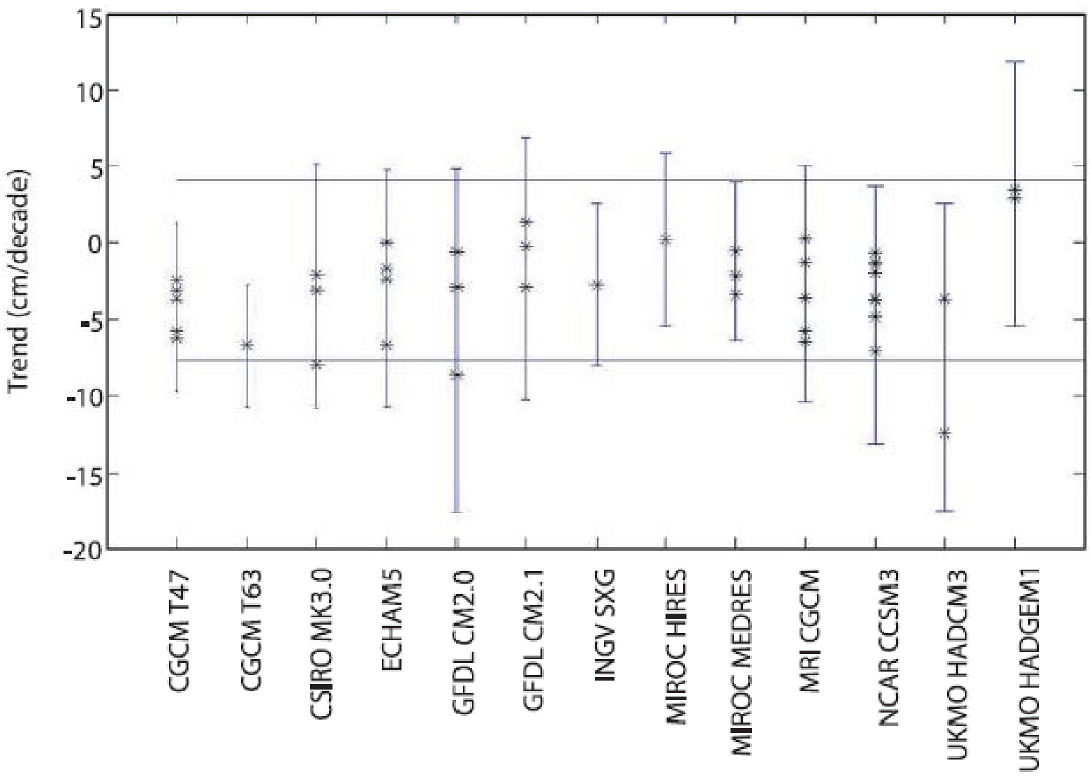


Figure 3.23. Same as Figure 3.8, except for Region 15.

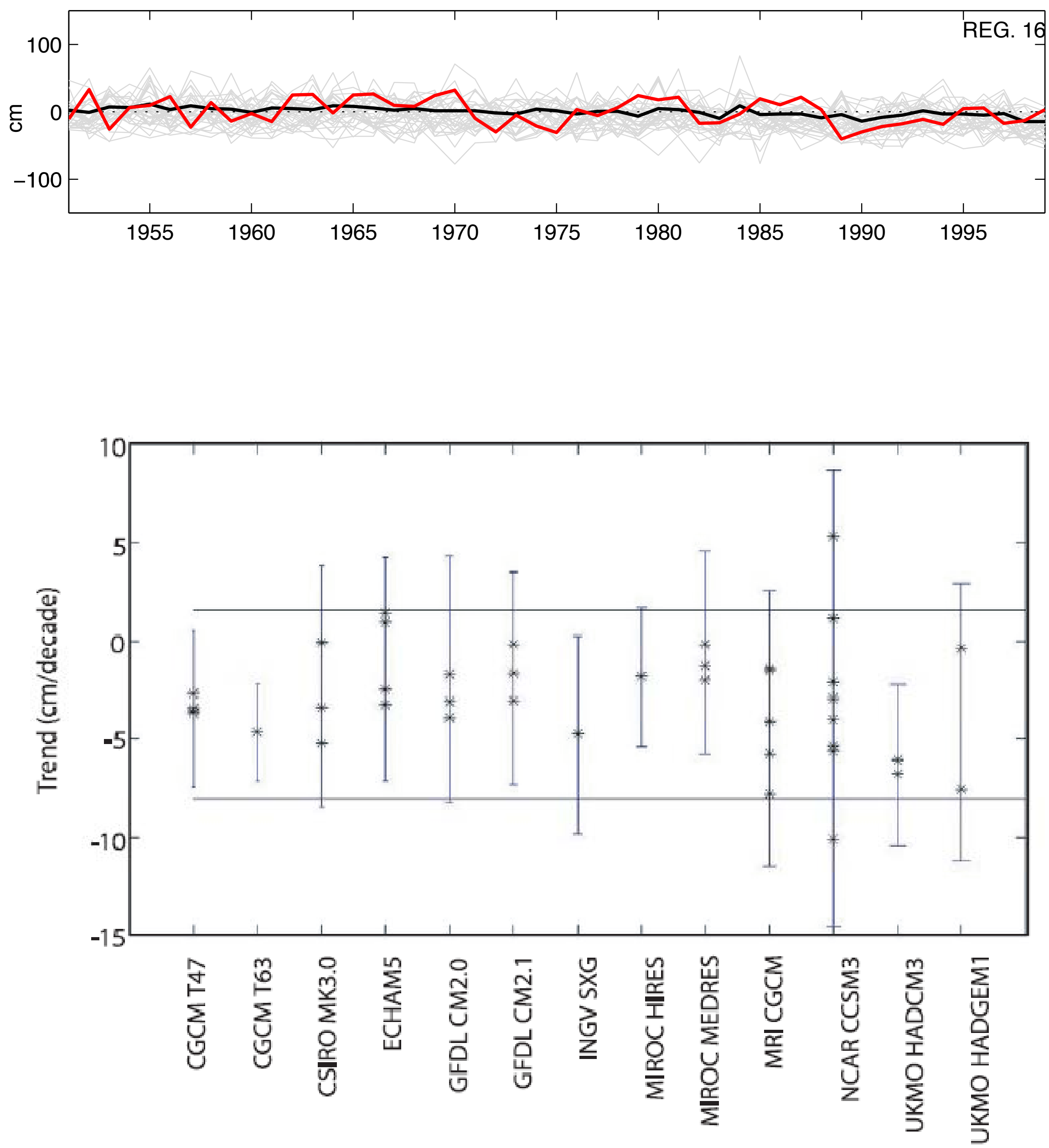
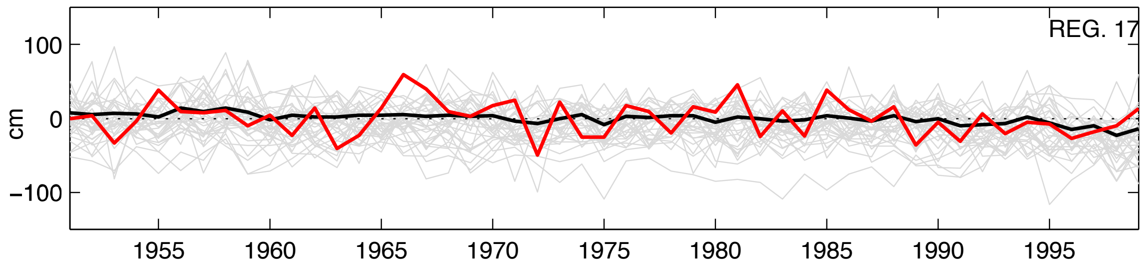


Figure 3.24. Same as Figure 3.8, except for Region 16.



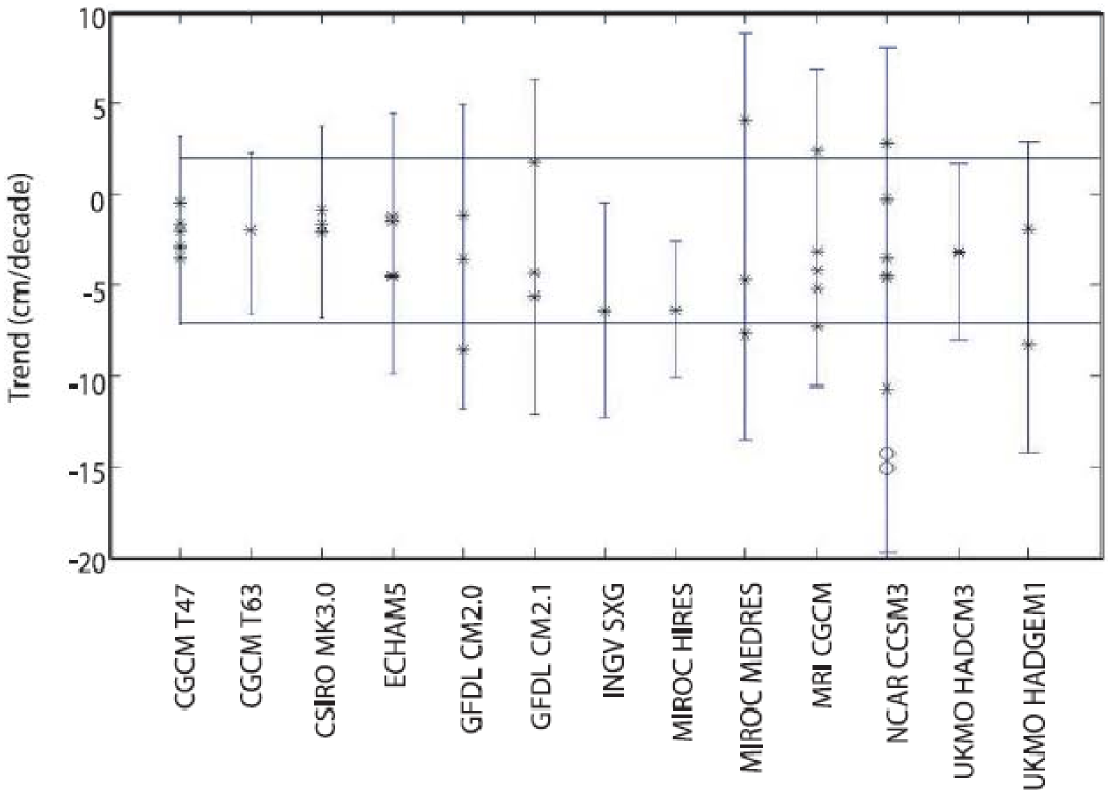
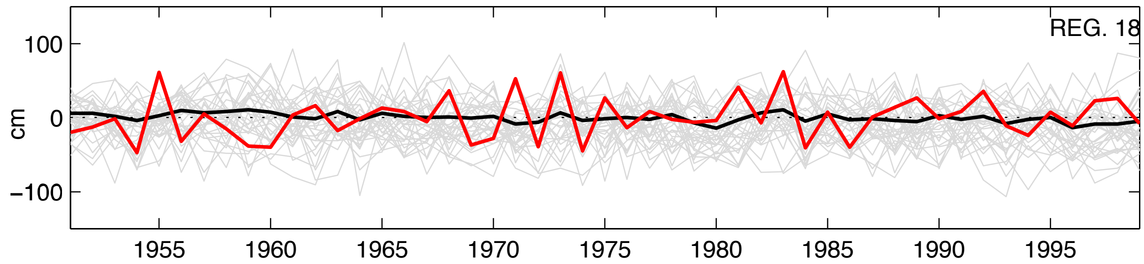


Figure 3.25. Same as Figure 3.8, except for Region 17.



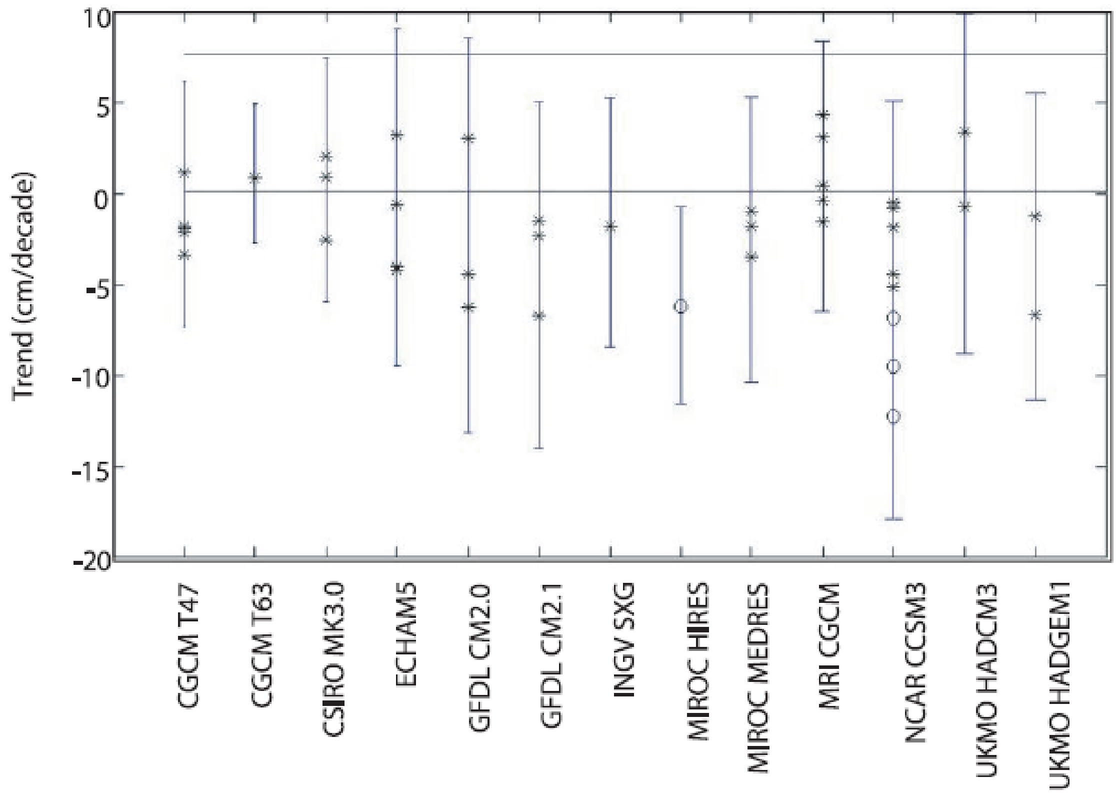
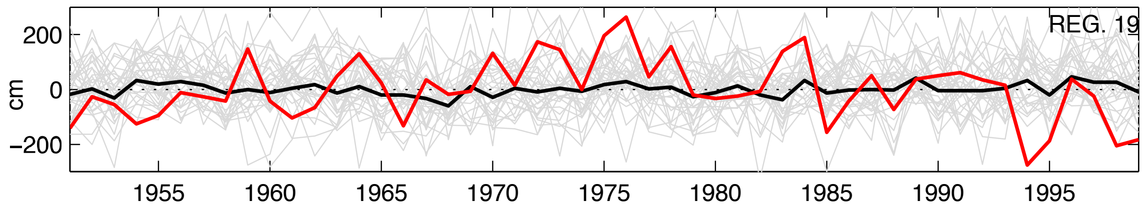


Figure 3.26. Same as Figure 3.8, except for Region 18.



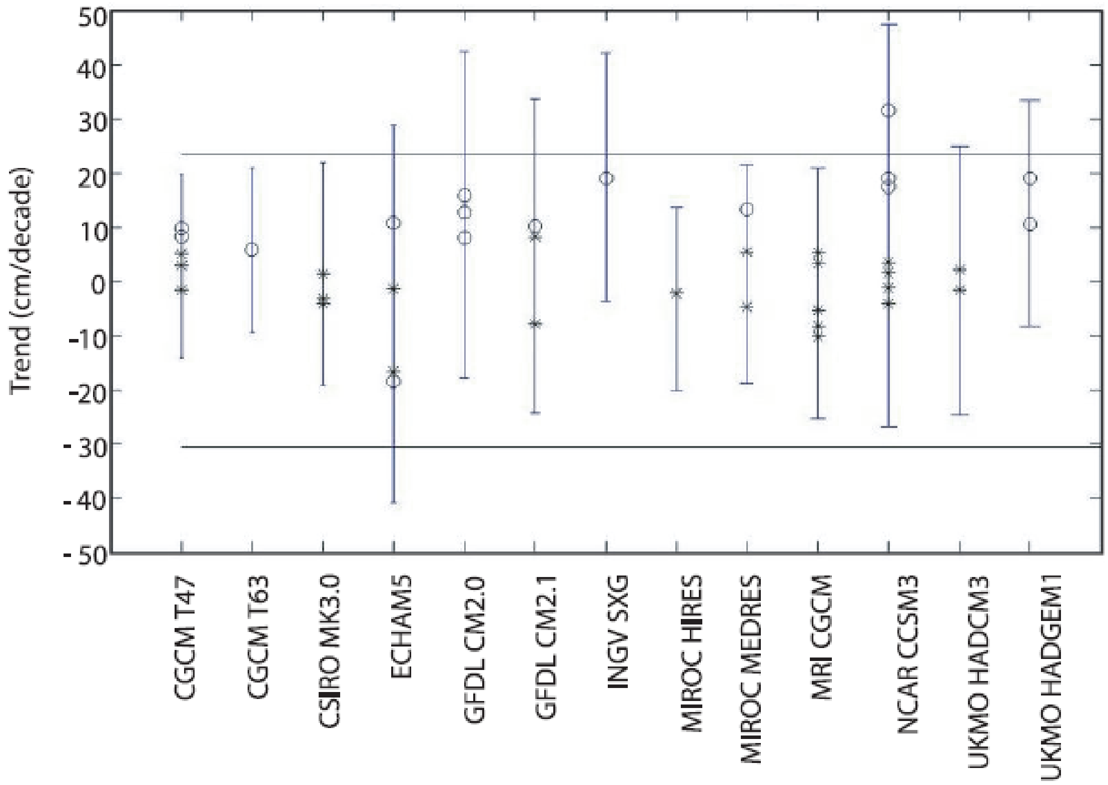
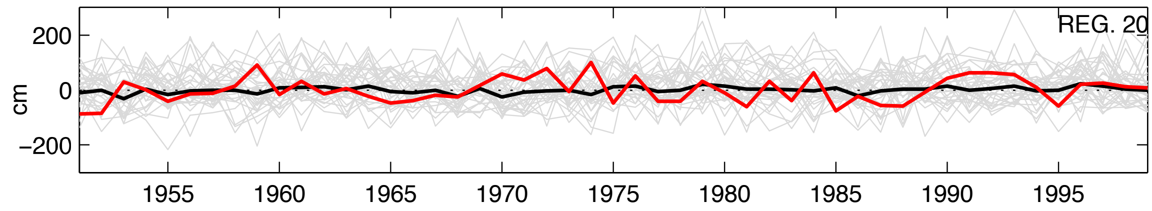


Figure 3.27. Same as Figure 3.8, except for Region 19.



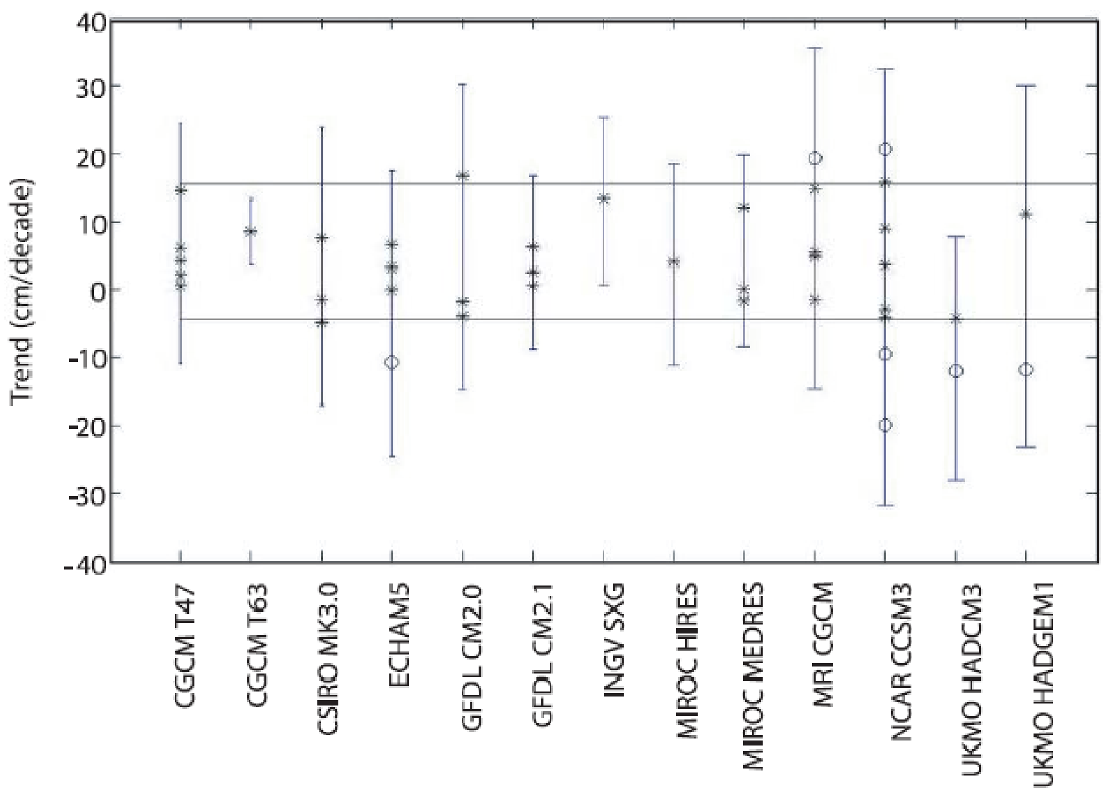


Figure 3.28. Same as Figure 3.8, except for Region 20.

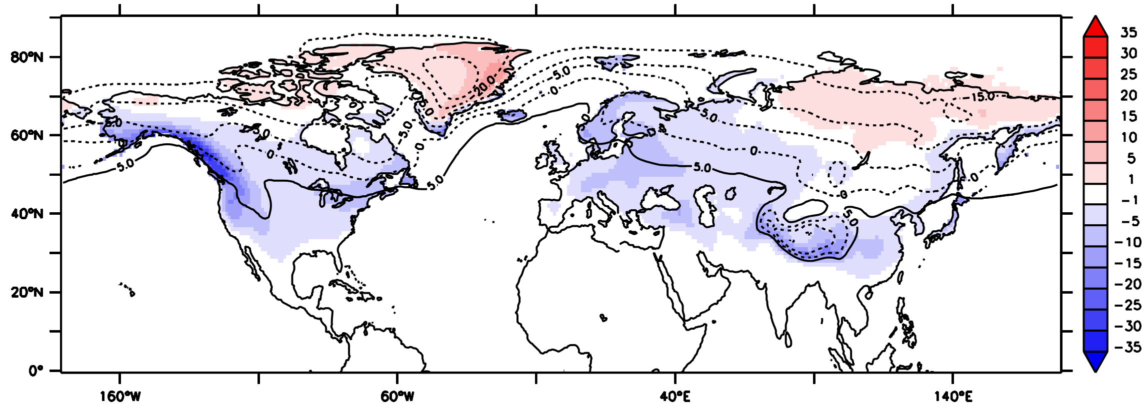


Figure 4.1. Multi-model ensemble mean trend in annual snowfall for the 21st century (2001 to 2099, shaded). Trends were computed using a least-squares regression and are in units of cm of accumulated annual snowfall per decade. Only trends that are statistically significant (95% level) are shaded. Isotherms are contoured for the multi-model ensemble mean temperatures from 1970 to 1999.

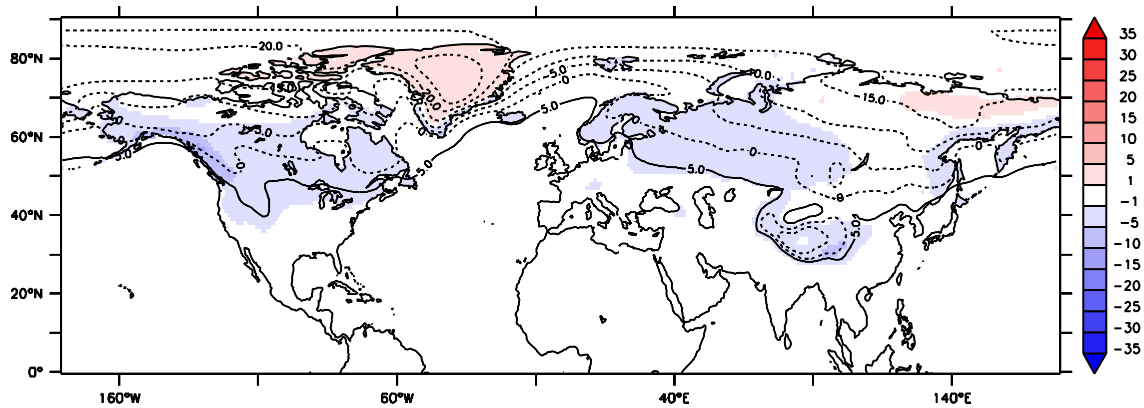


Figure 4.2. Same as Figure 4.1, only for September, October, and November.

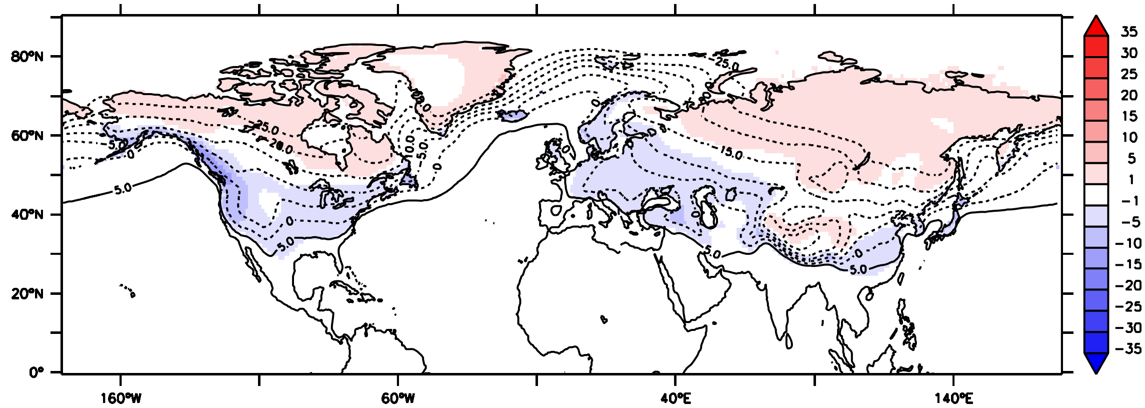


Figure 4.3. Same as Figure 4.1, only for December, January, and February.

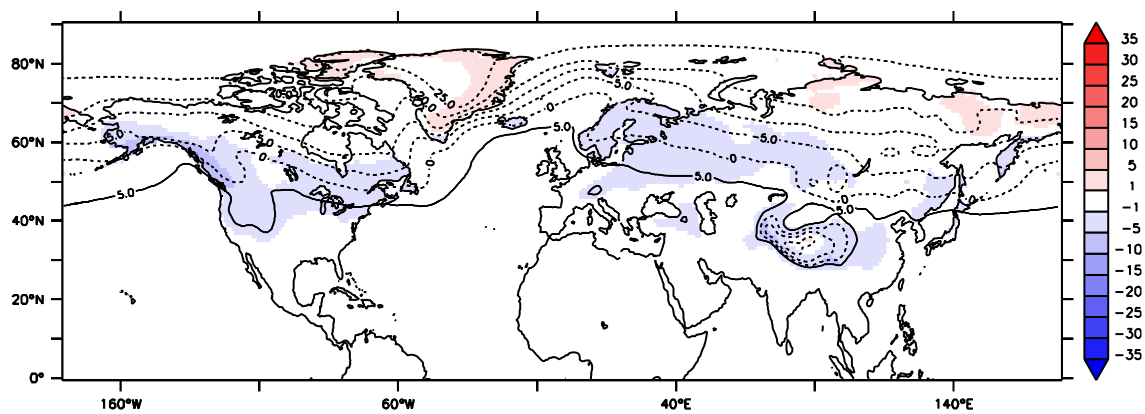


Figure 4.4. Same as Figure 4.1, only for March, April, and May.

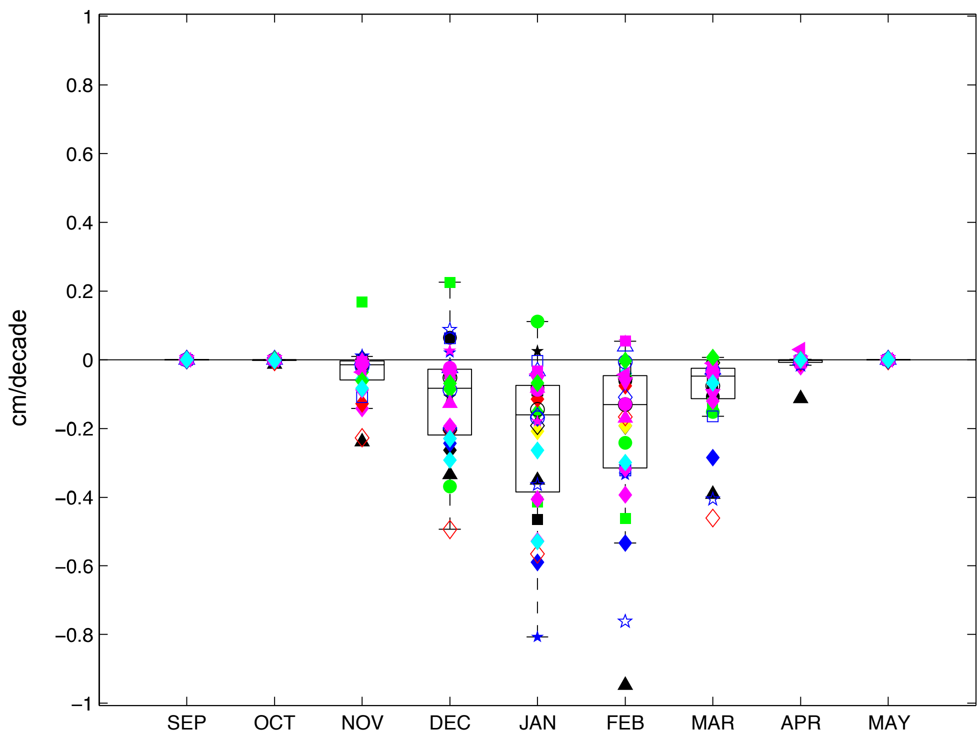


Figure 4.5. Monthly changes in snowfall for Region 1. Changes are presented as a rate units of cm/decade. Color markers represent model-simulated trends: CGCM T47, black filled markers; CGCM T63, cyan filled markers; CSIRO Mk3.0, yellow filled markers; ECHAM5, green filled markers; GFDL CM2.0, magenta filled markers; GFDL CM2.1, blue filled markers; INGV SXG red filled markers; MIROC 3.2 Hi-Res., red open markers; MIROC 3.2 Med-Res., black open markers; MRI-CGCM, blue open markers; NCAR CCSM3, magenta open markers; UKMO HadCM3, green open markers; UKMO HadGEM1, cyan open markers. Different marker shapes of the same color denote individual simulations that were performed with the same model.

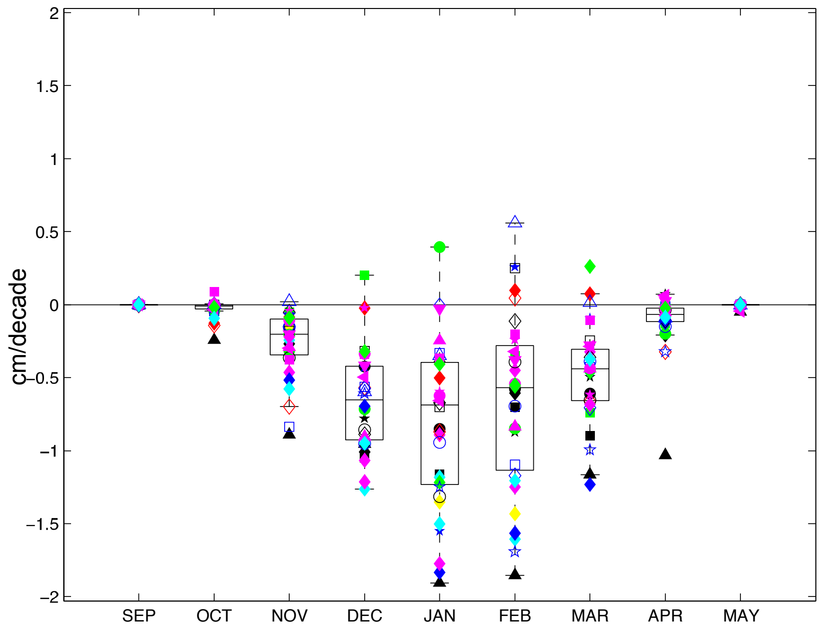


Figure 4.6. Same as Figure 4.5, except for Region 2.

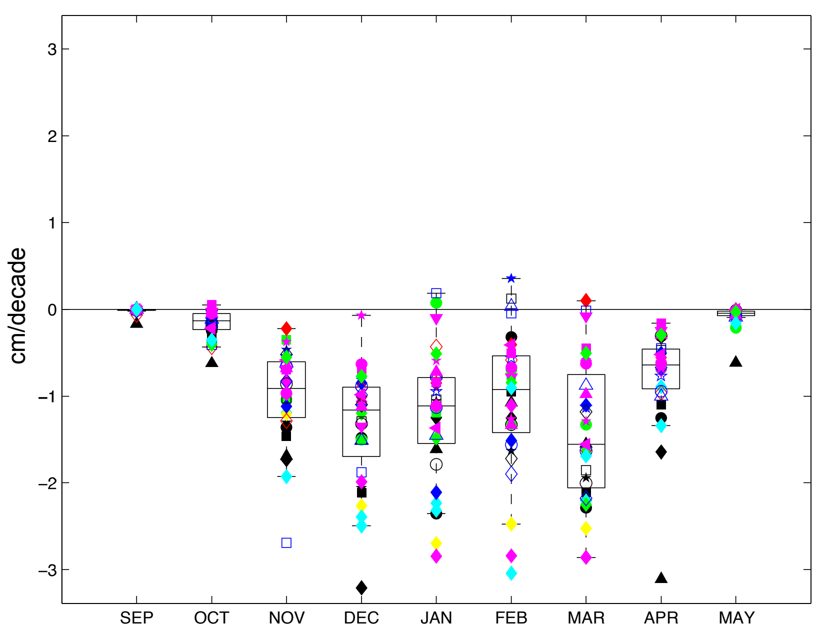


Figure 4.7. Same as Figure 4.5, except for Region 3.

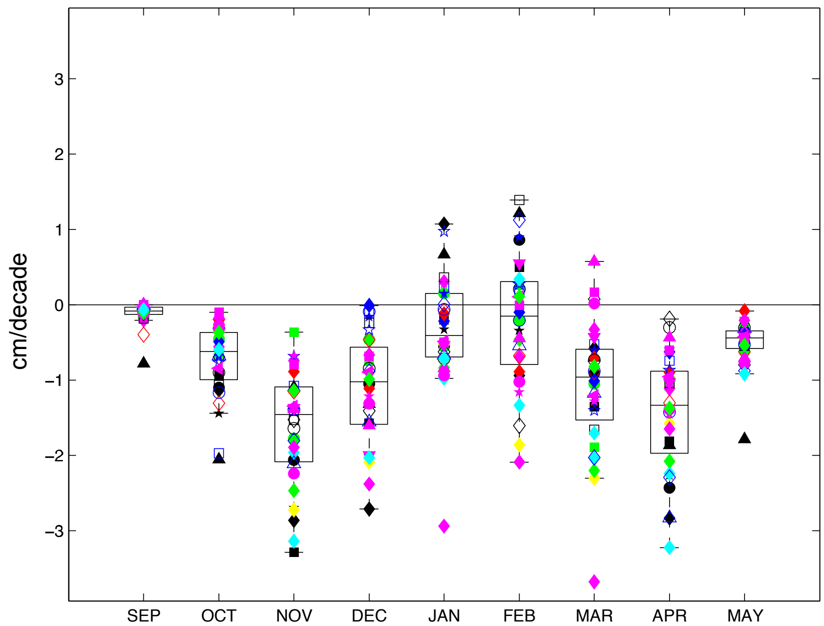


Figure 4.8. Same as Figure 4.5, except for Region 4.

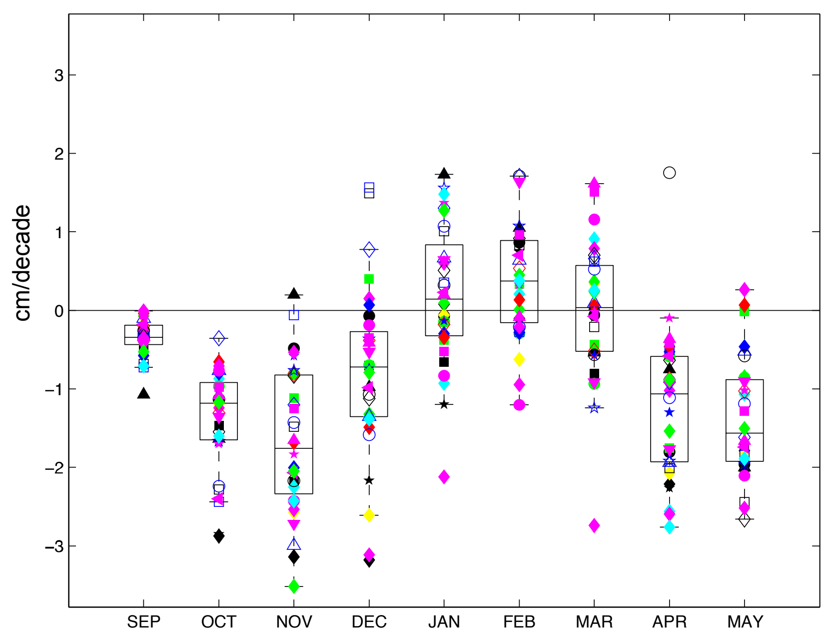


Figure 4.9. Same as Figure 4.5, except for Region 5.

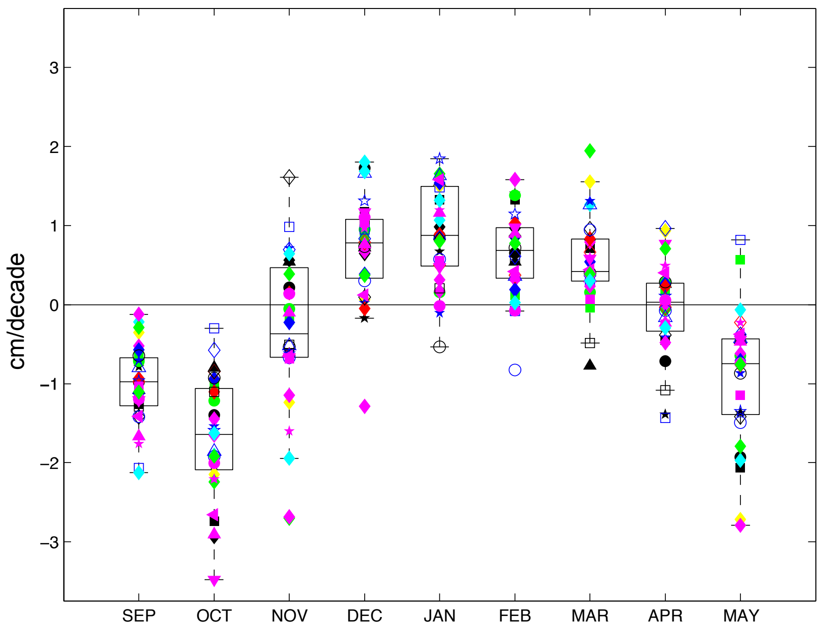


Figure 4.10. Same as Figure 4.5, except for Region 6.

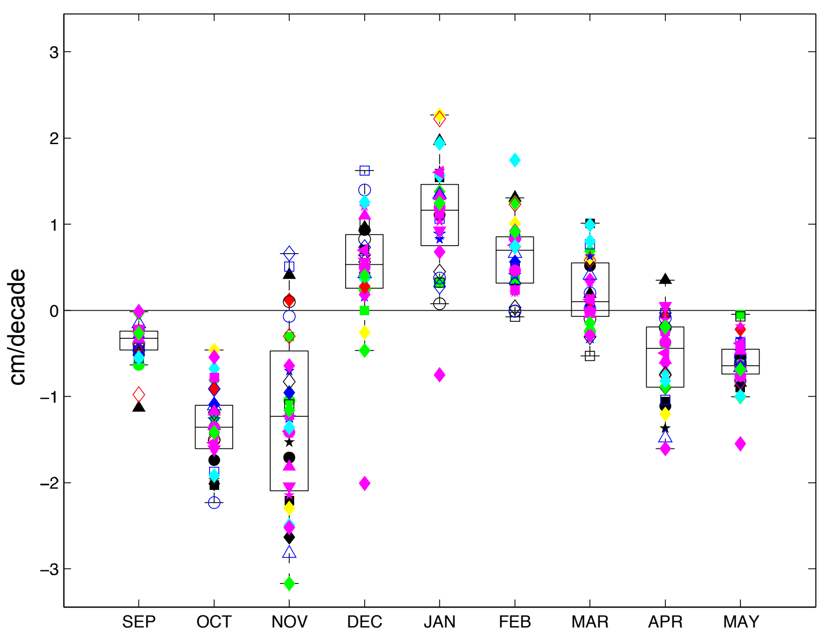


Figure 4.11. Same as Figure 4.5, except for Region 7.

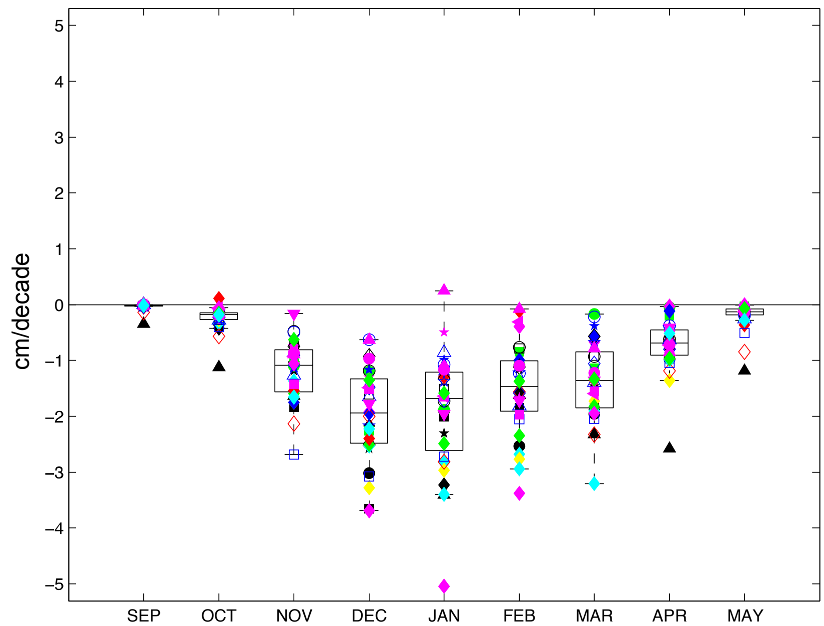


Figure 4.12. Same as Figure 4.5, except for Region 8.

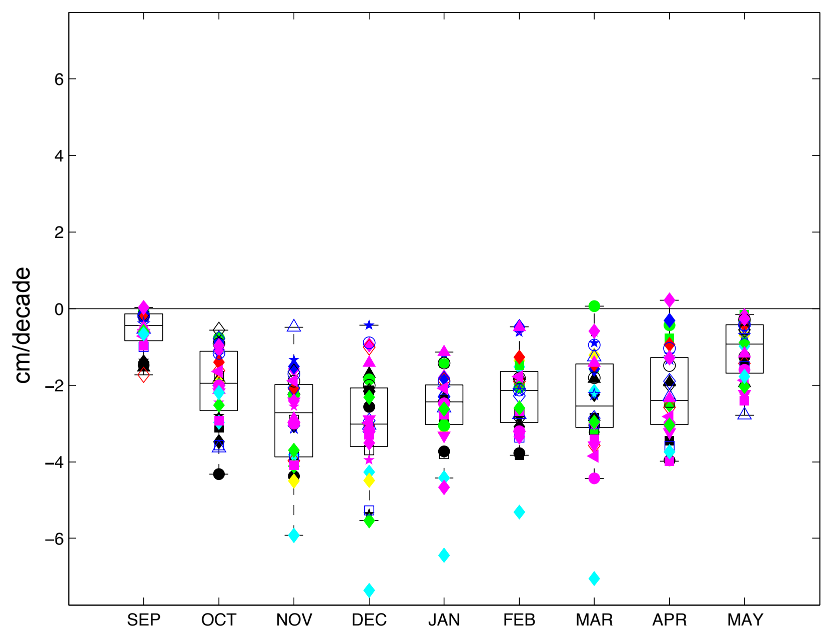


Figure 4.13. Same as Figure 4.5, except for Region 9.

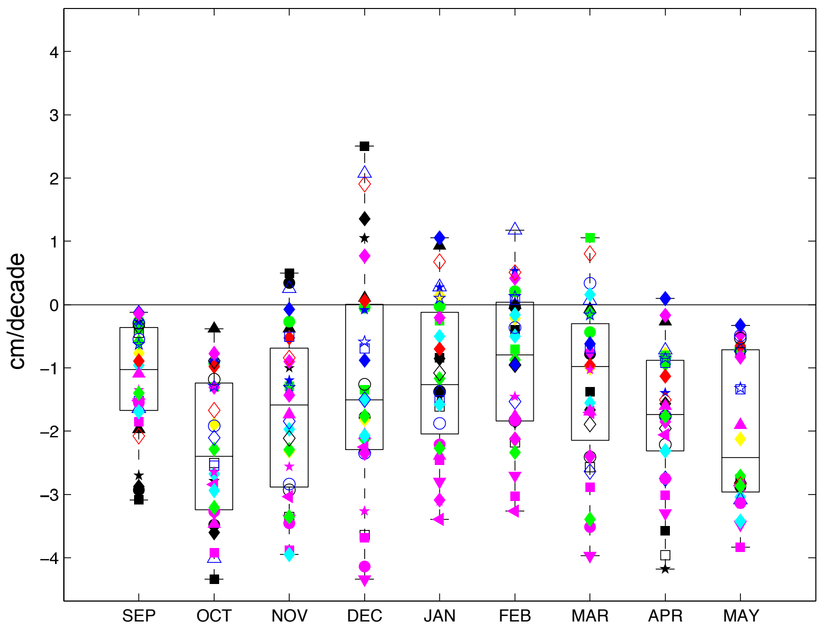


Figure 4.14. Same as Figure 4.5, except for Region 10.

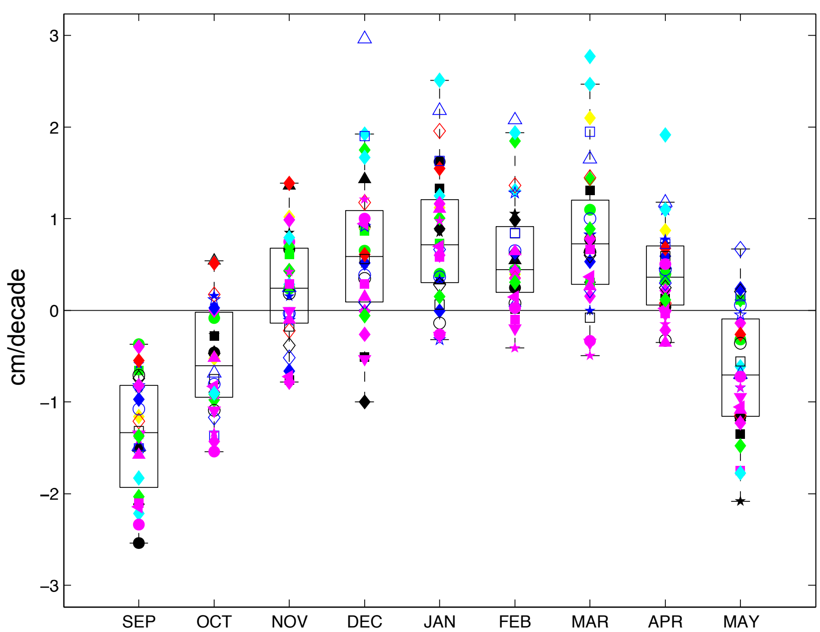


Figure 4.15. Same as Figure 4.5, except for Region 11.

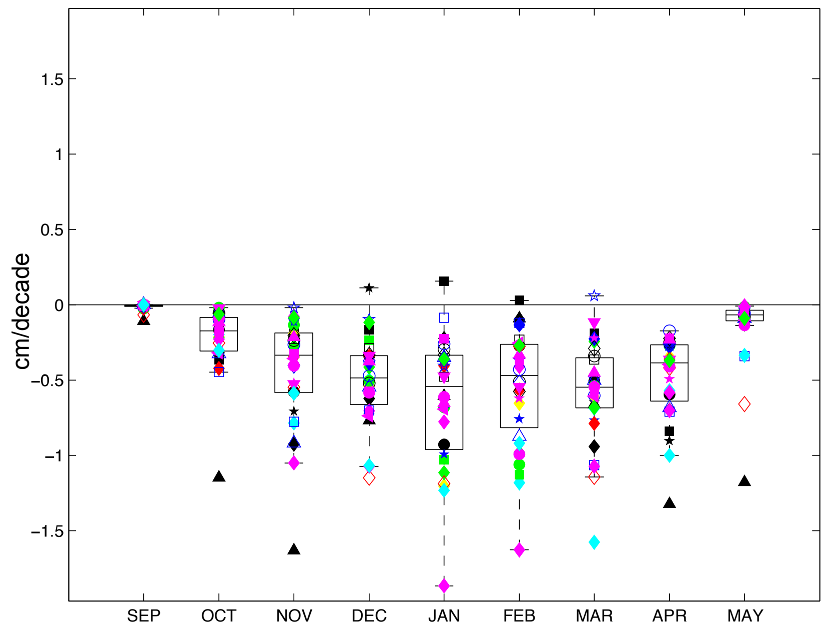


Figure 4.16. Same as Figure 4.5, except for Region 12.

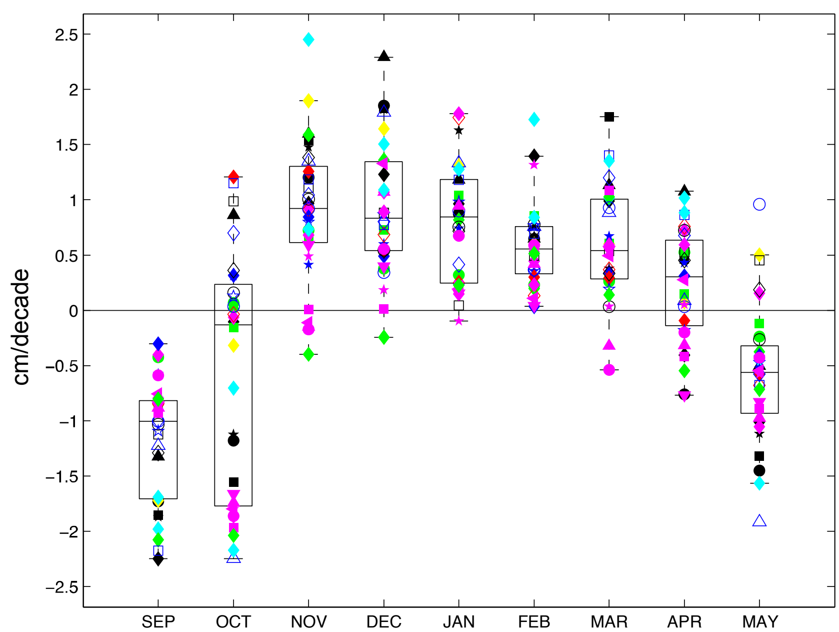


Figure 4.17. Same as Figure 4.5, except for Region 13.

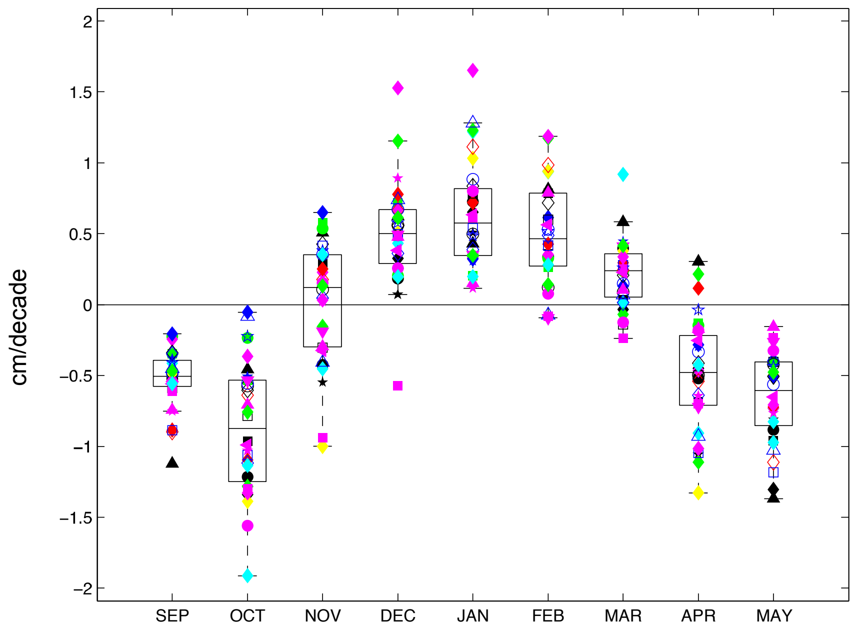


Figure 4.18. Same as Figure 4.5, except for Region 14.

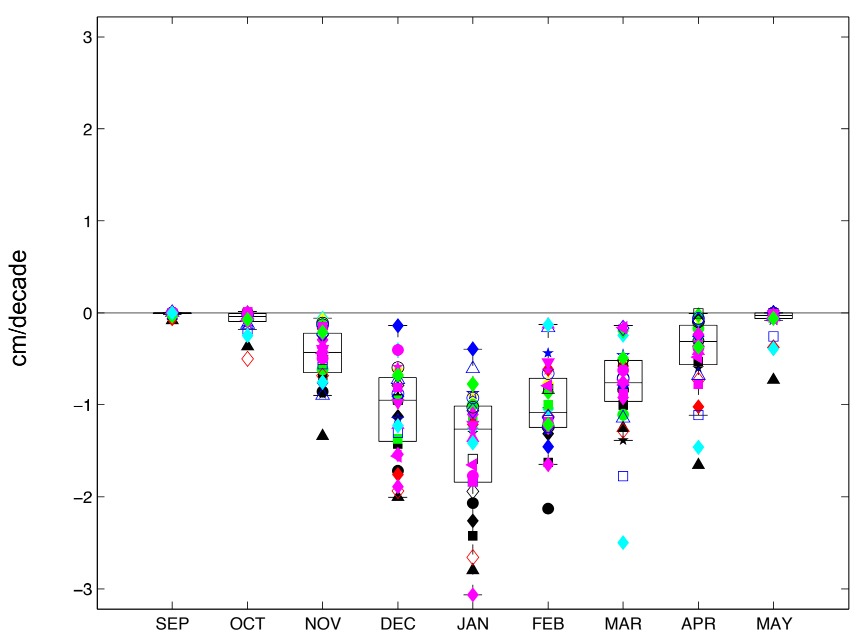


Figure 4.19. Same as Figure 4.5, except for Region 15.

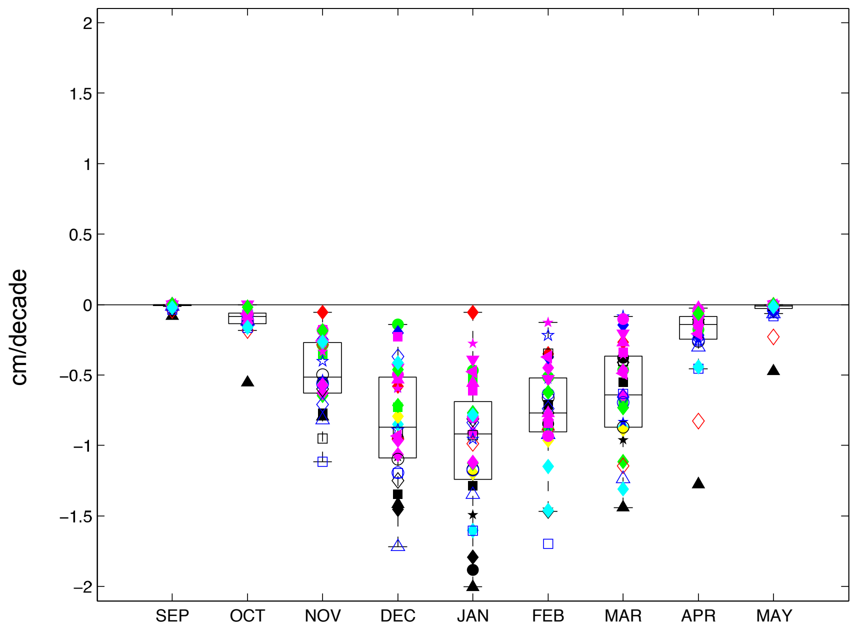


Figure 4.20. Same as Figure 4.5, except for Region 16.

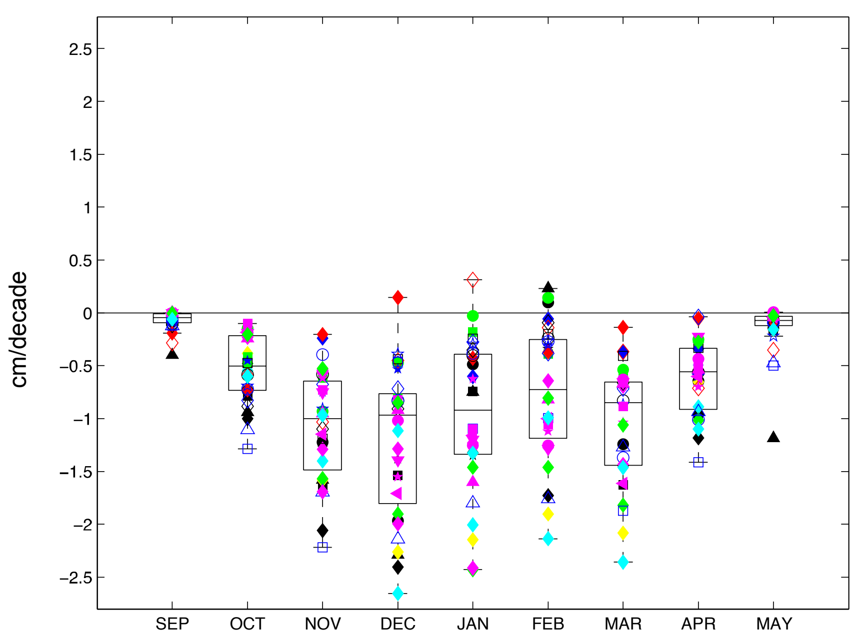


Figure 4.21. Same as Figure 4.5, except for Region 17.

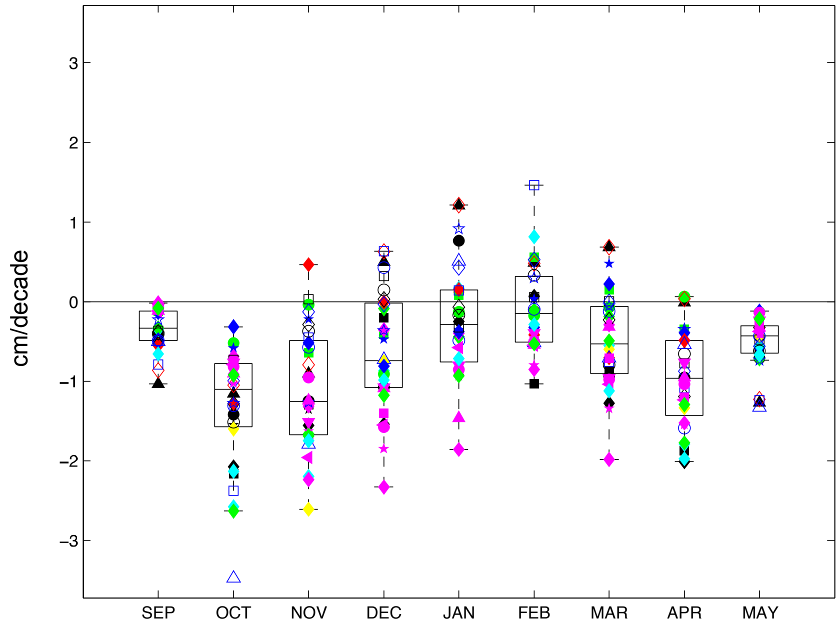


Figure 4.22. Same as Figure 4.5, except for Region 18.

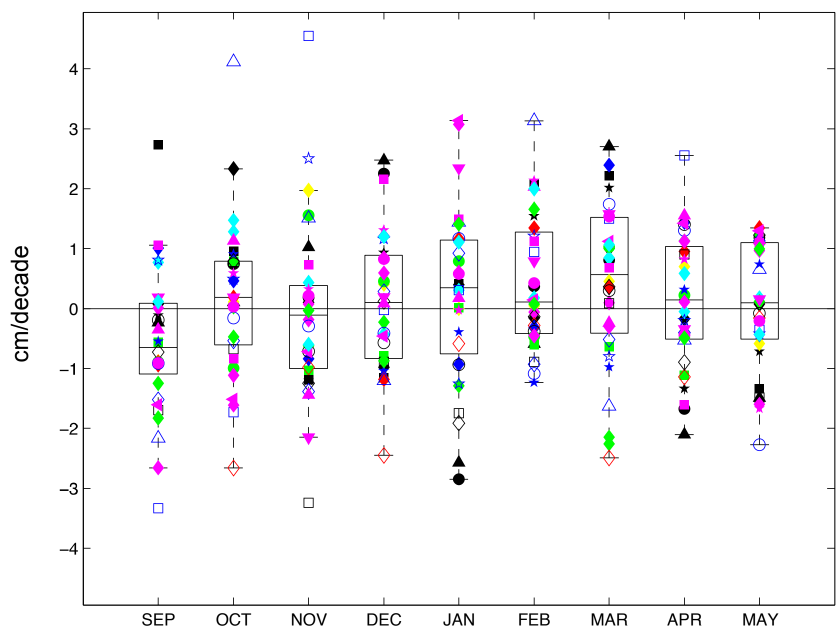


Figure 4.23. Same as Figure 4.5, except for Region 19.

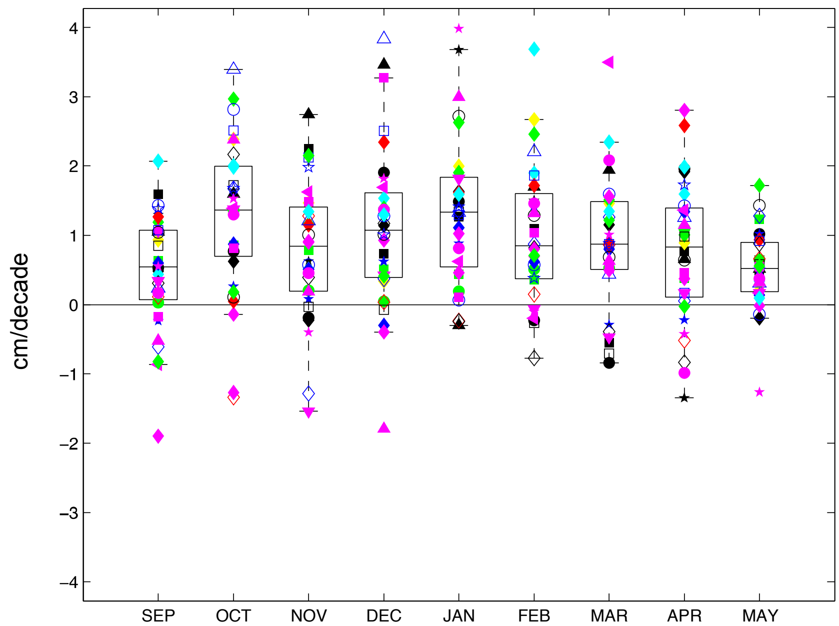


Figure 4.24. Same as Figure 4.5, except for Region 20.

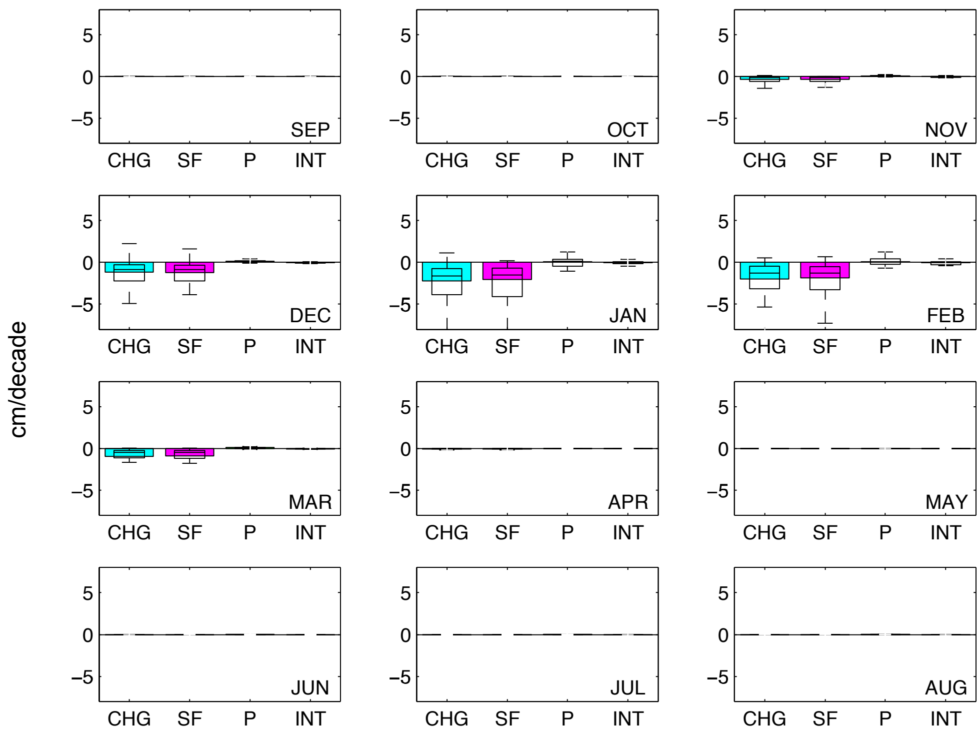


Figure 4.25. Decomposition of monthly snowfall changes into parts attributable to changes in temperature and precipitation for Region 1 (units are cm/decade). CHG represents the change in snowfall (ΔS), SF represents the effect of a change in snow fraction (ΔSF), P represents the effect of a change in precipitation (ΔP), and INT represents the interaction of ΔSF and ΔP. Shading represents the multi-model ensemble while the box plots illustrate the results from all individual climate simulations.

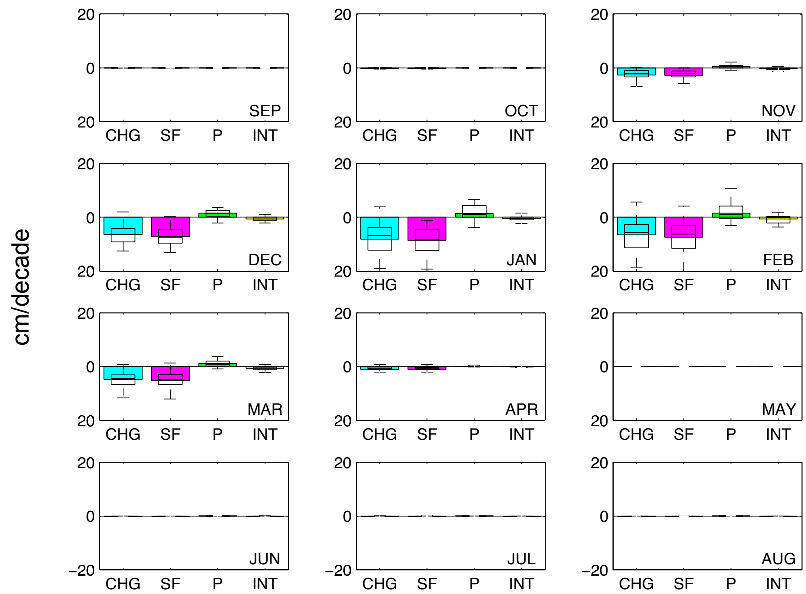


Figure 4.26. Same as Figure 4.25, except for Region 2.

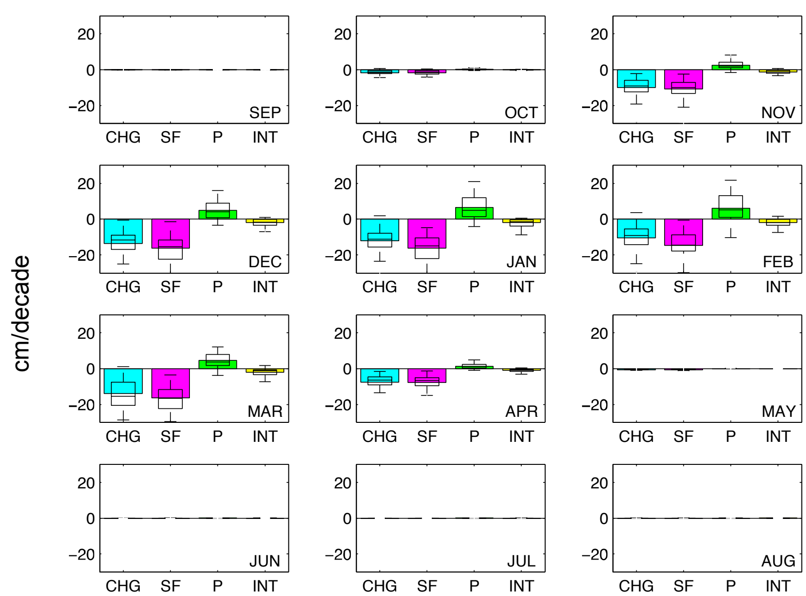


Figure 4.27. Same as Figure 4.25, except for Region 3.

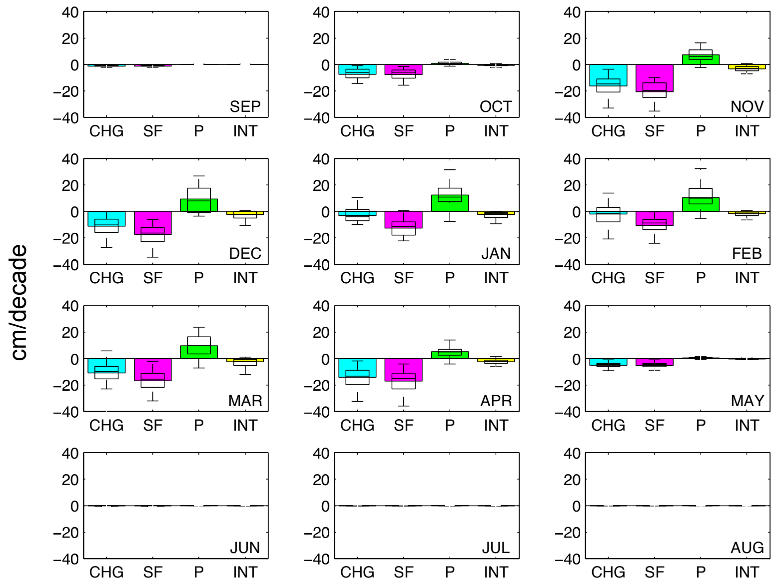


Figure 4.28. Same as Figure 4.25, except for Region 4.

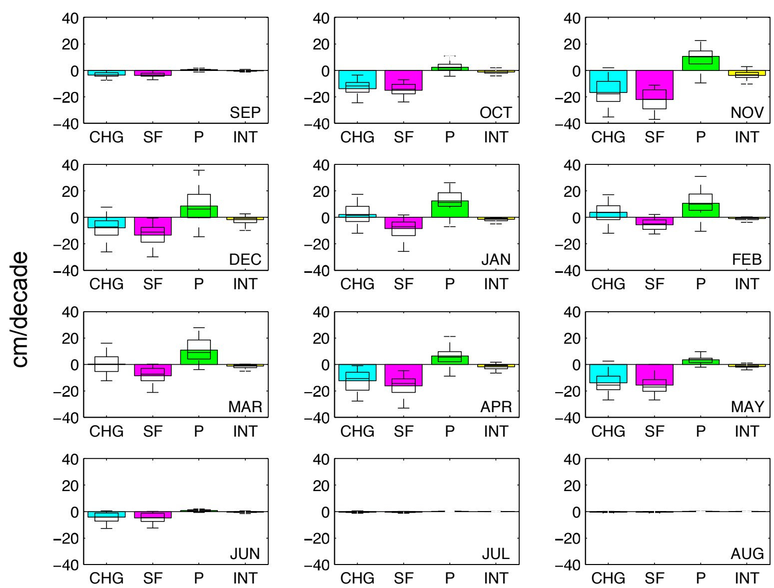


Figure 4.29. Same as Figure 4.25, except for Region 5

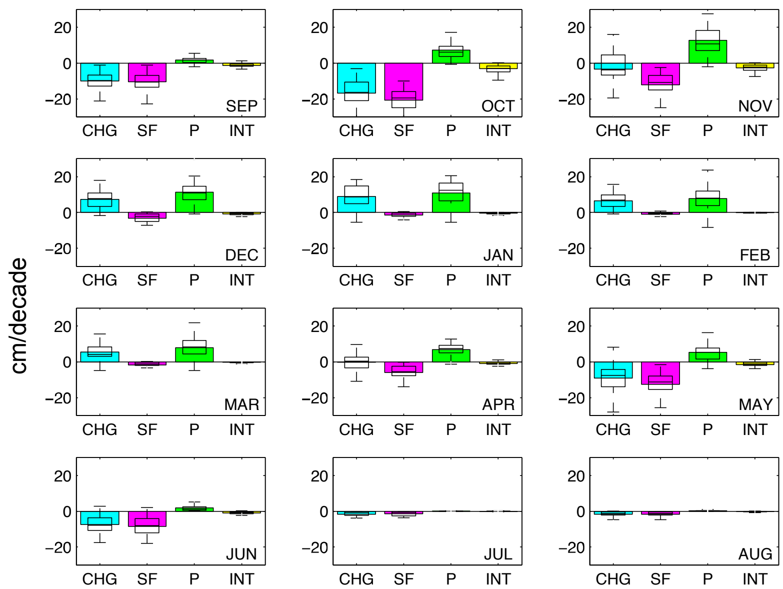


Figure 4.30. Same as Figure 4.25, except for Region 6.

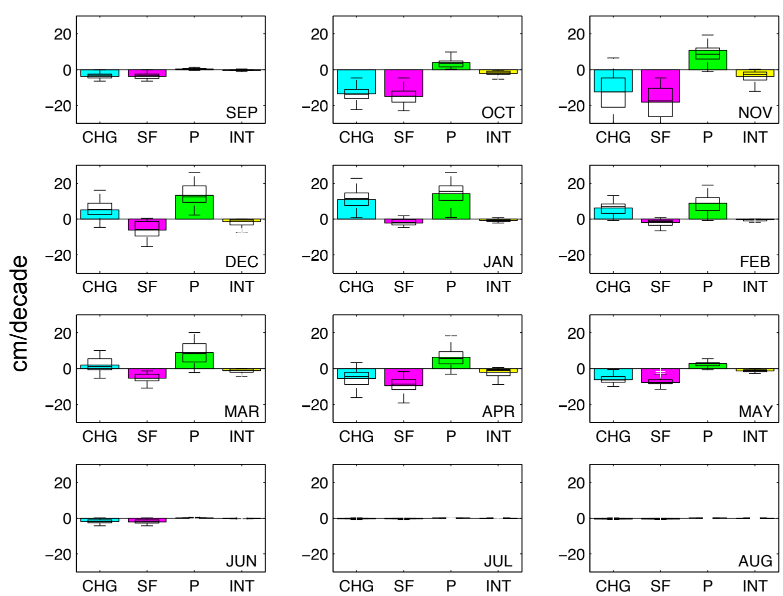


Figure 4.31. Same as Figure 4.25, except for Region 7.

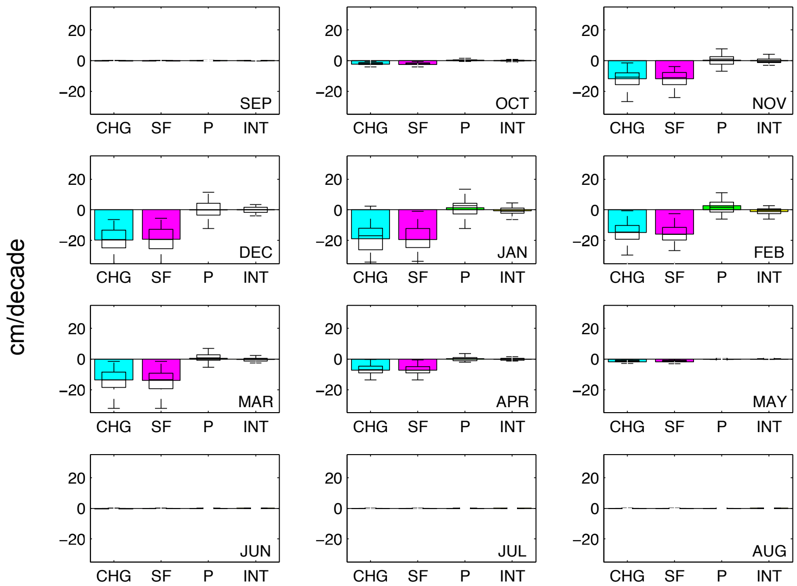


Figure 4.32. Same as Figure 4.25, except for Region 8.

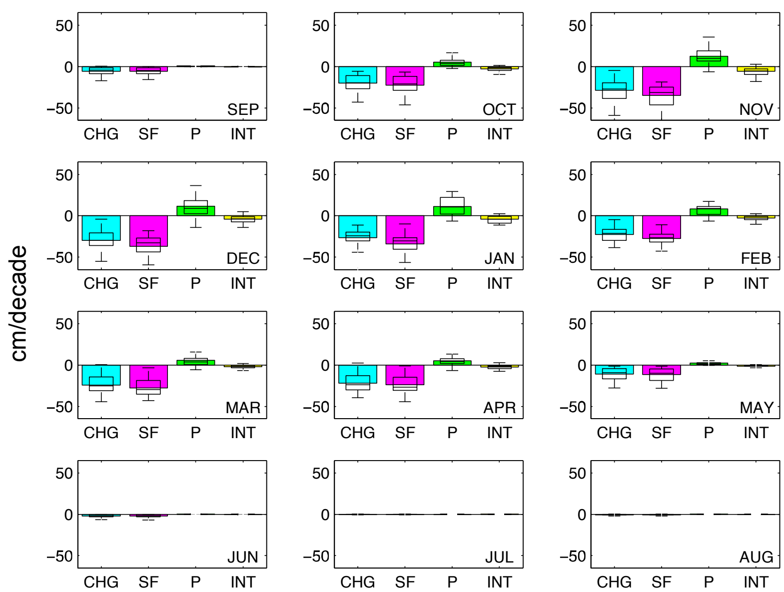


Figure 4.33. Same as Figure 4.25, except for Region 9.

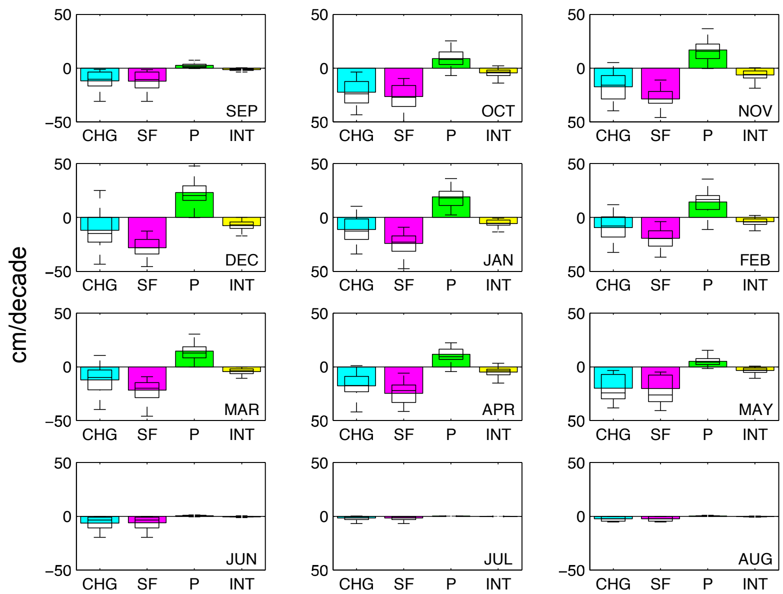


Figure 4.34. Same as Figure 4.25, except for Region 10.

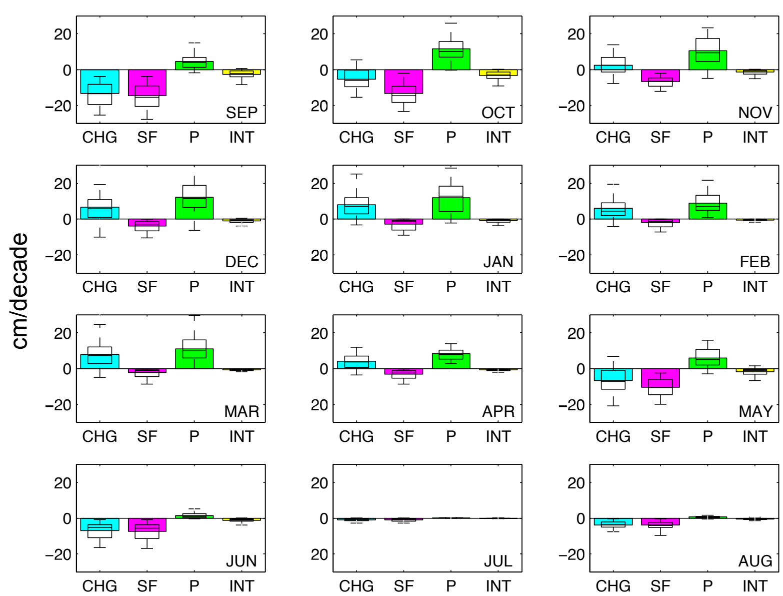


Figure 4.35. Same as Figure 4.25, except for Region 11.

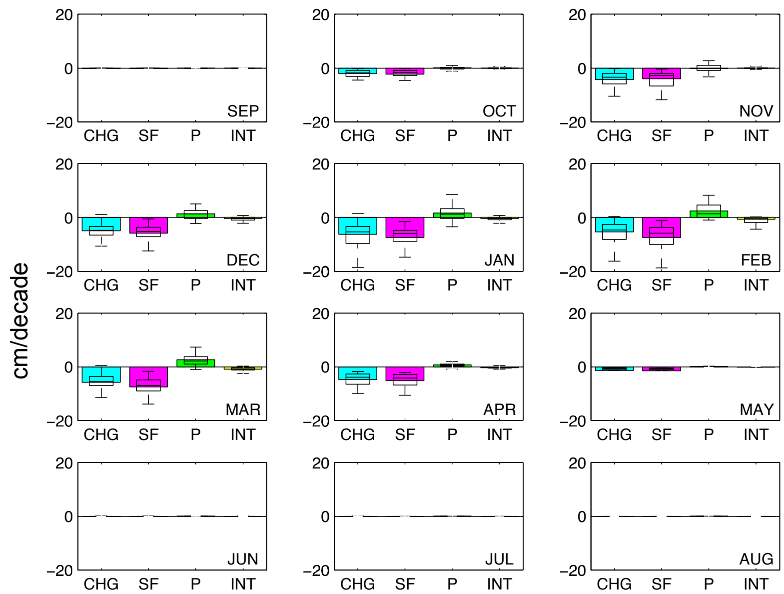


Figure 4.36. Same as Figure 4.25, except for Region 12.

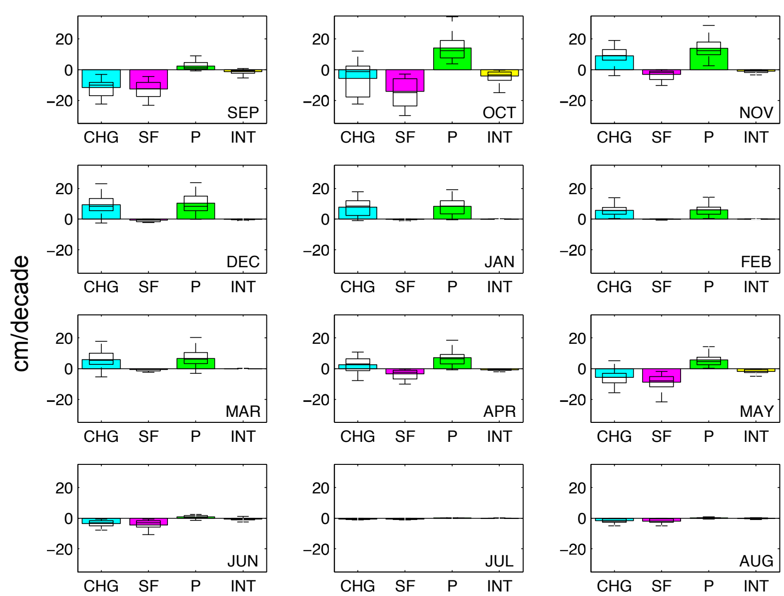


Figure 4.37. Same as Figure 4.25, except for Region 13.

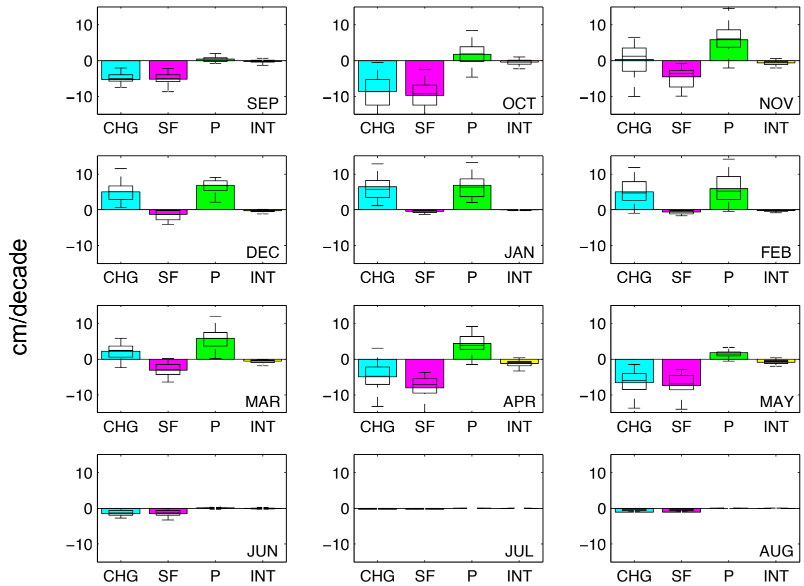


Figure 4.38. Same as Figure 4.25, except for Region 14.

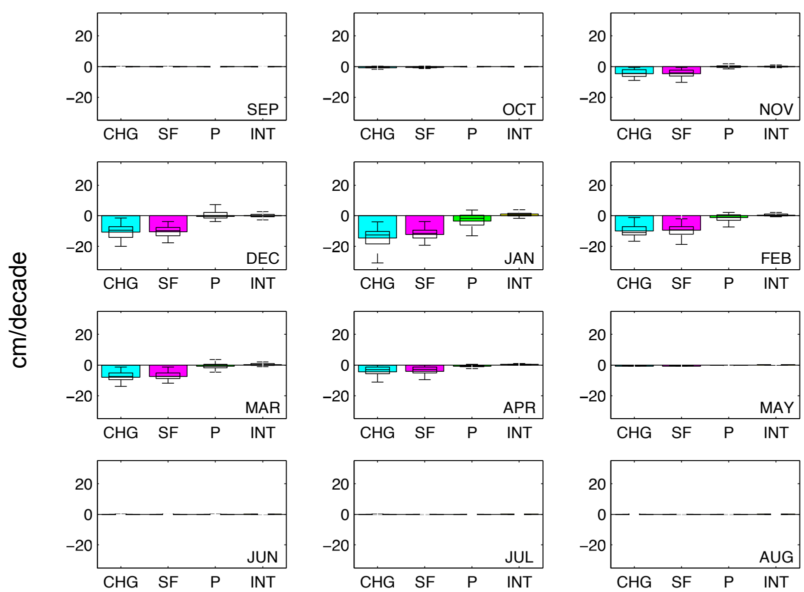


Figure 4.39. Same as Figure 4.25, except for Region 15.

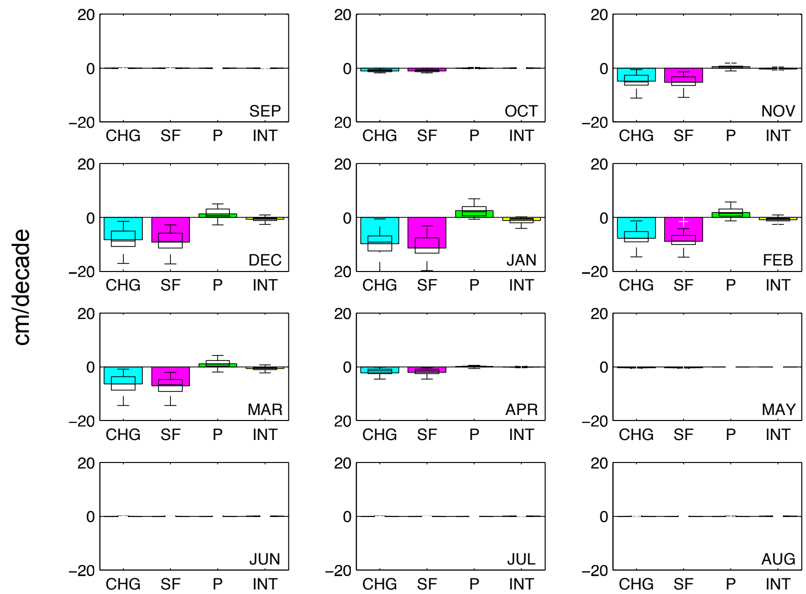


Figure 4.40. Same as Figure 4.25, except for Region 16.

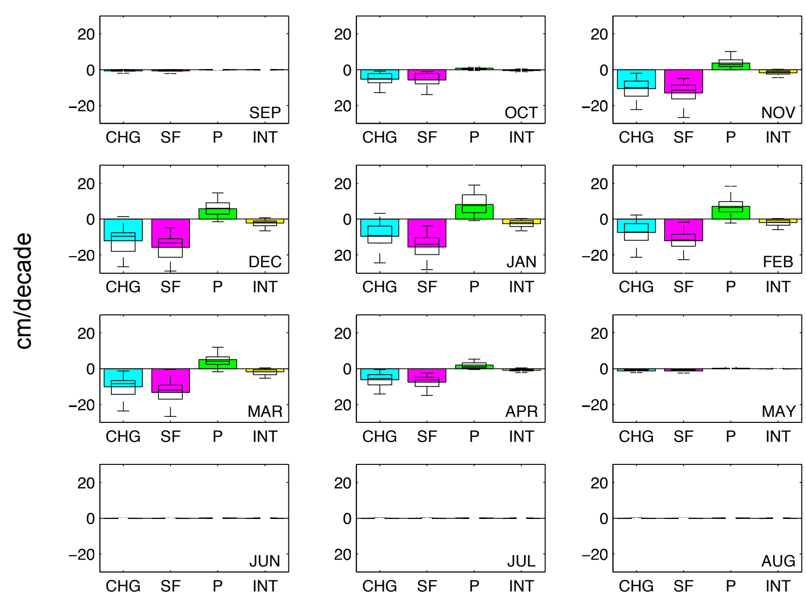


Figure 4.41. Same as Figure 4.25, except for Region 17.

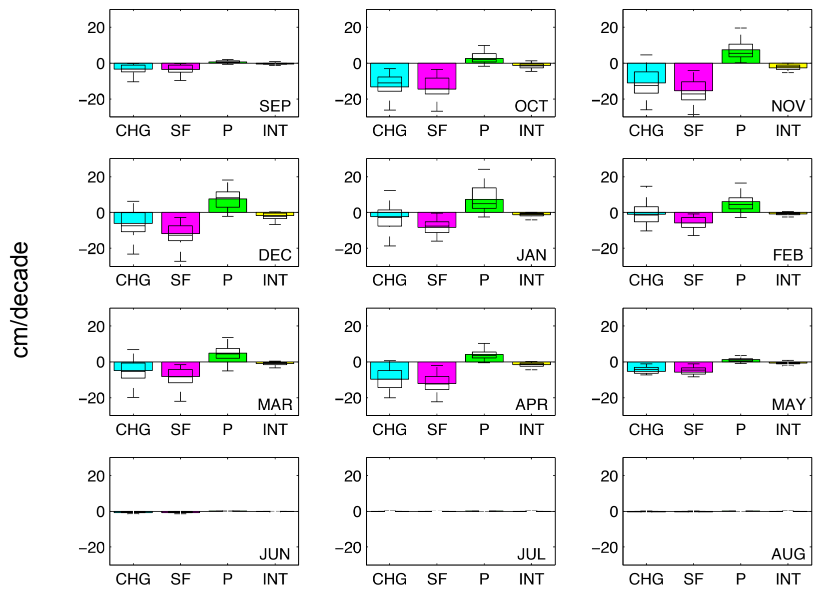


Figure 4.42. Same as Figure 4.25, except for Region 18.

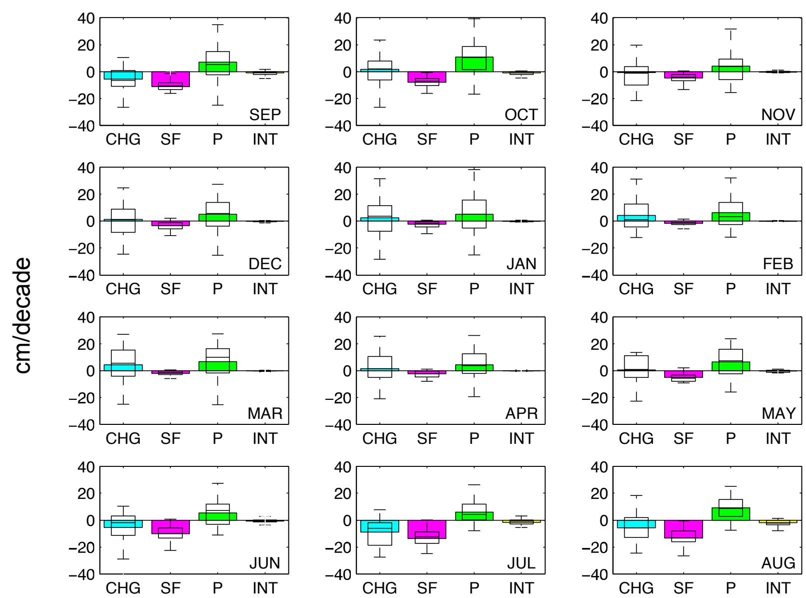


Figure 4.43. Same as Figure 4.25, except for Region 19.

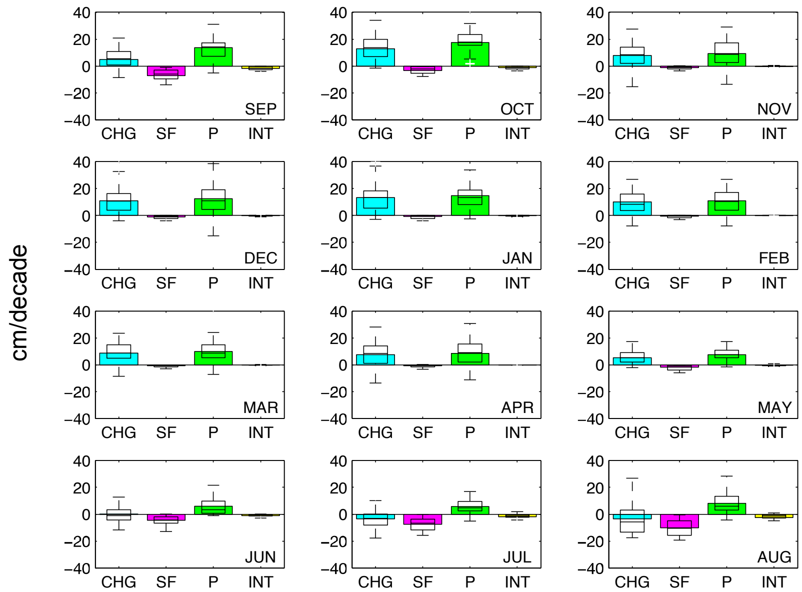


Figure 4.44. Same as Figure 4.25, except for Region 20.

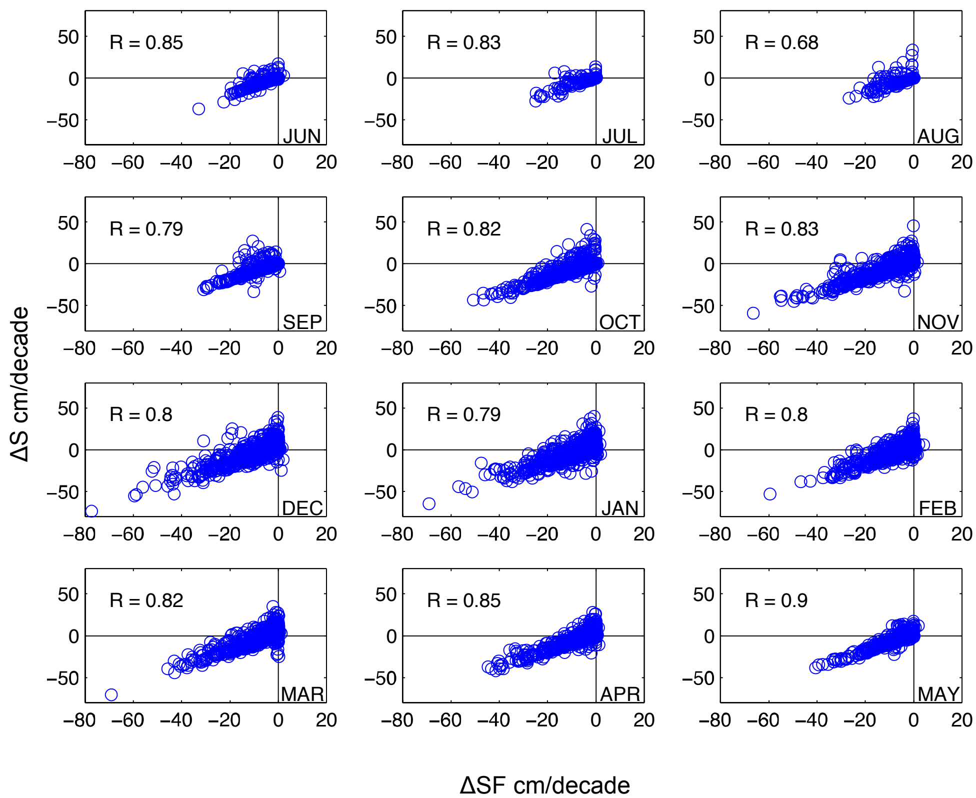


Figure 4.45 Relationship between the change in snow fraction (ΔSF , x-axis) and the change in snowfall (ΔS , y-axis) for all months (units are in cm/decade). In each subplot, results from the decomposition analysis are shown for every region and every individual model simulation. The correlation coefficient is also given.

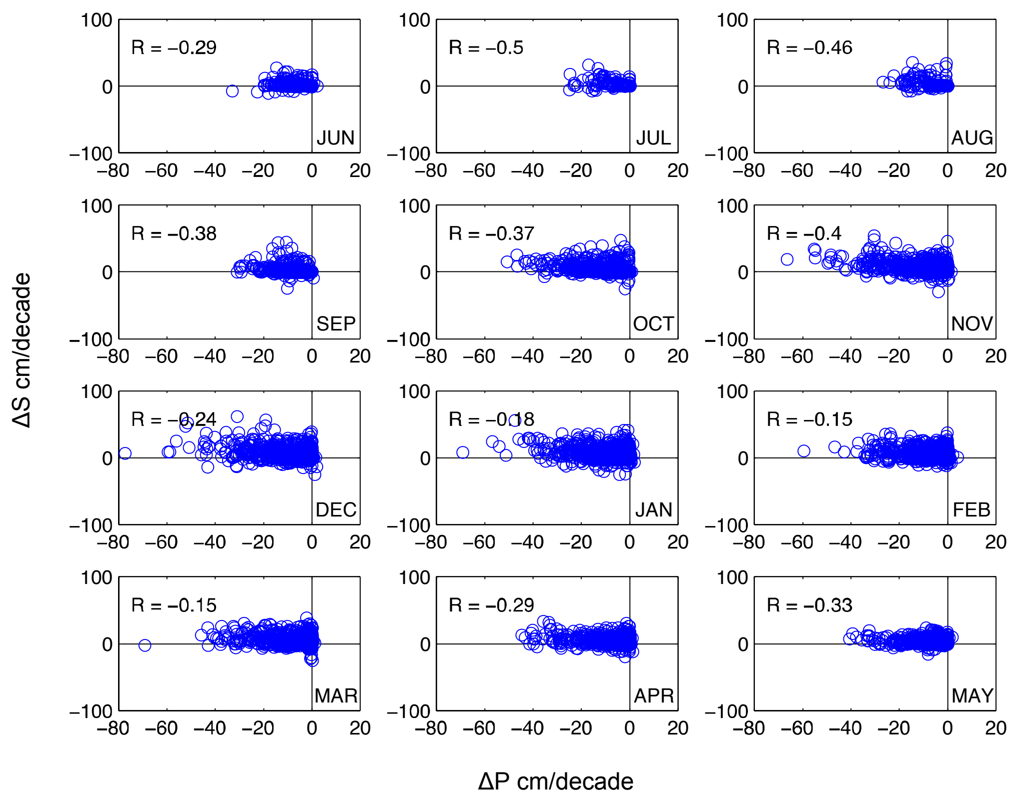


Figure 4.46. Same as Figure 4.45, for the relationship between the change in precipitation (ΔP , x-axis) and the change in snowfall (ΔS , y-axis).

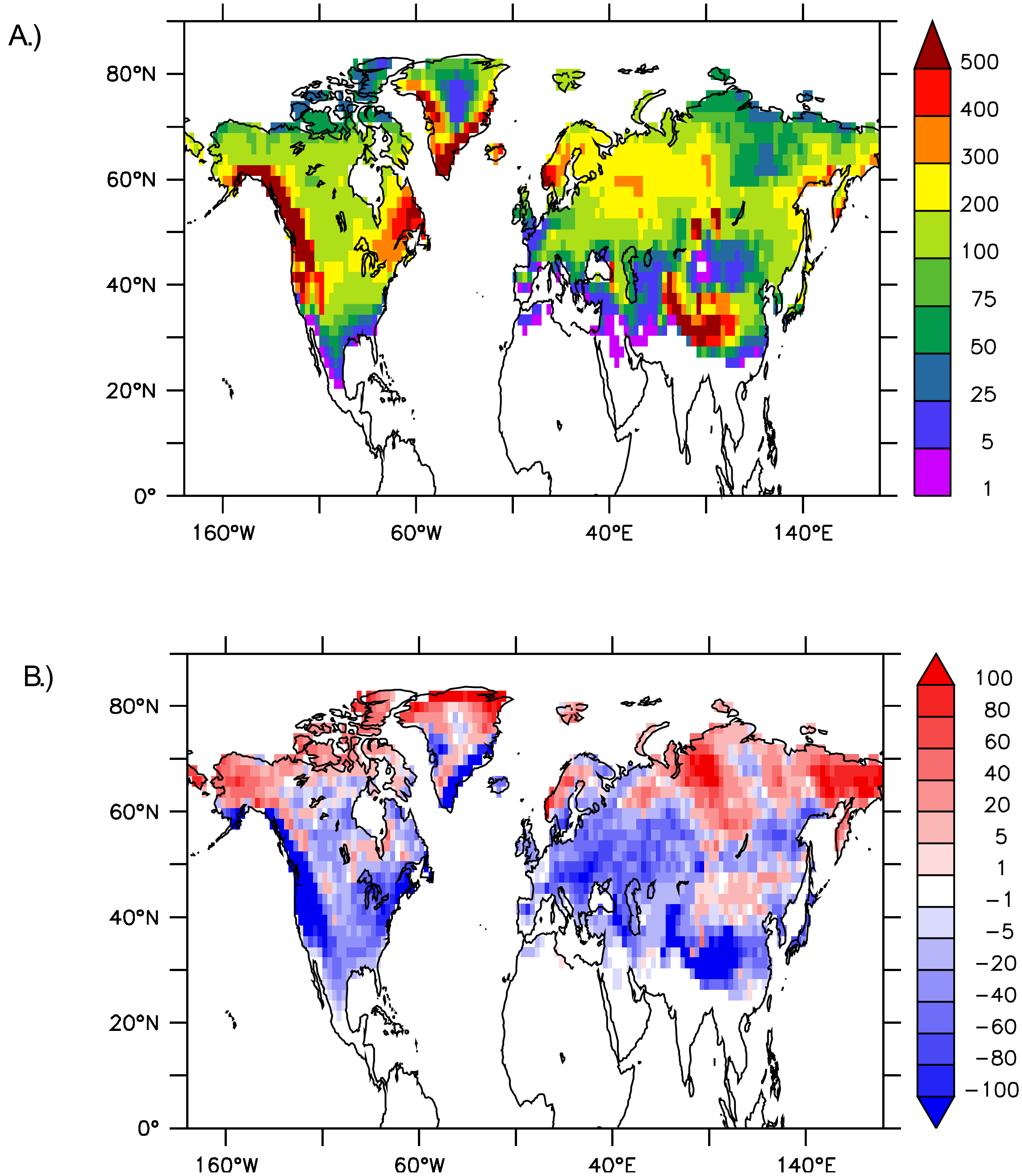


Figure 4.47. (A) Frequency of daily snowfall events exceeding 5 cm during the period 2001 to 2020 from the GFDL CM2.1 model, and (B), the difference in the frequency between (2081-2100) and (2001-2020).

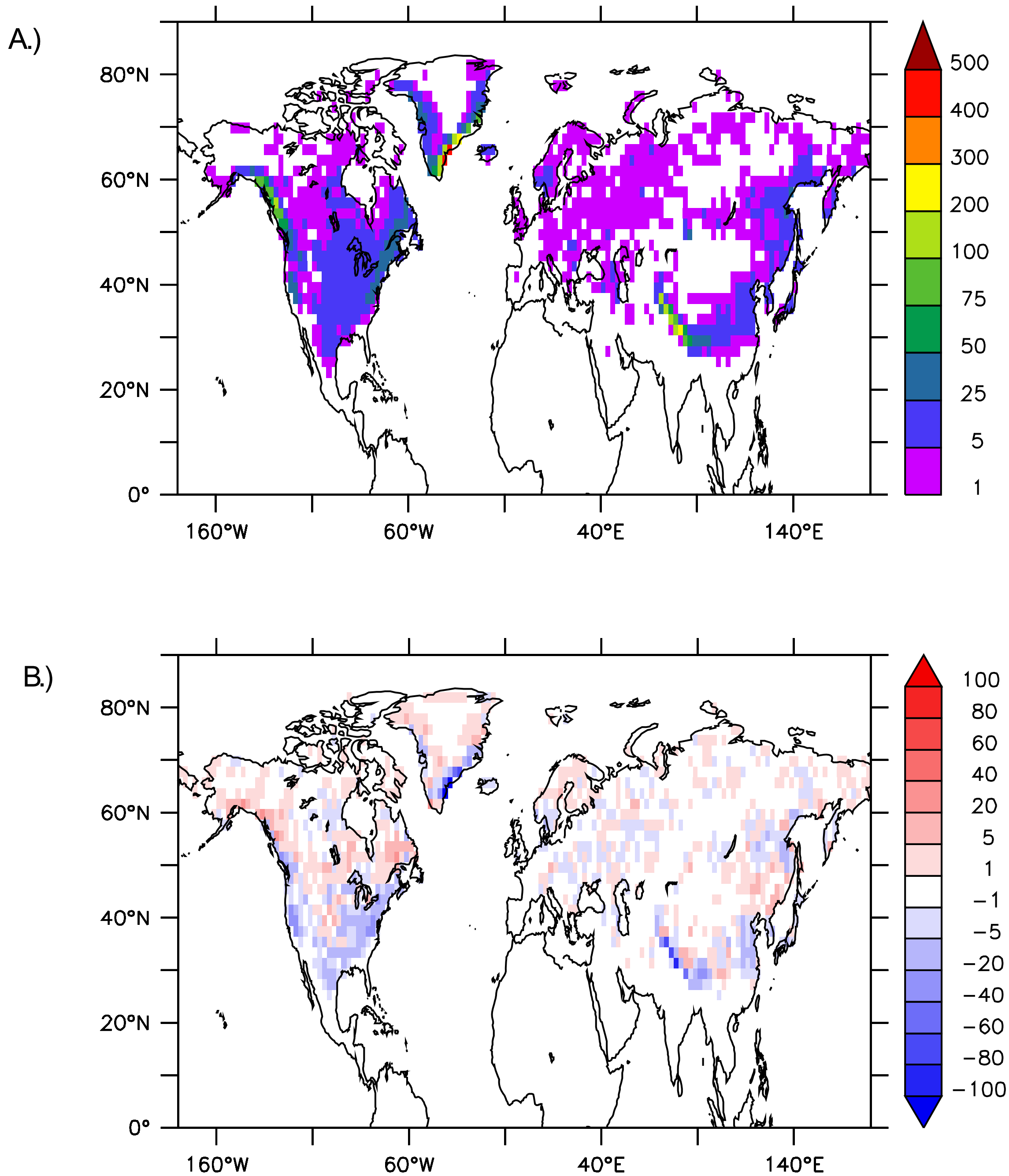


Figure 4.48. Same as figure 4.48, except for daily snowfall events exceeding 20 cm.

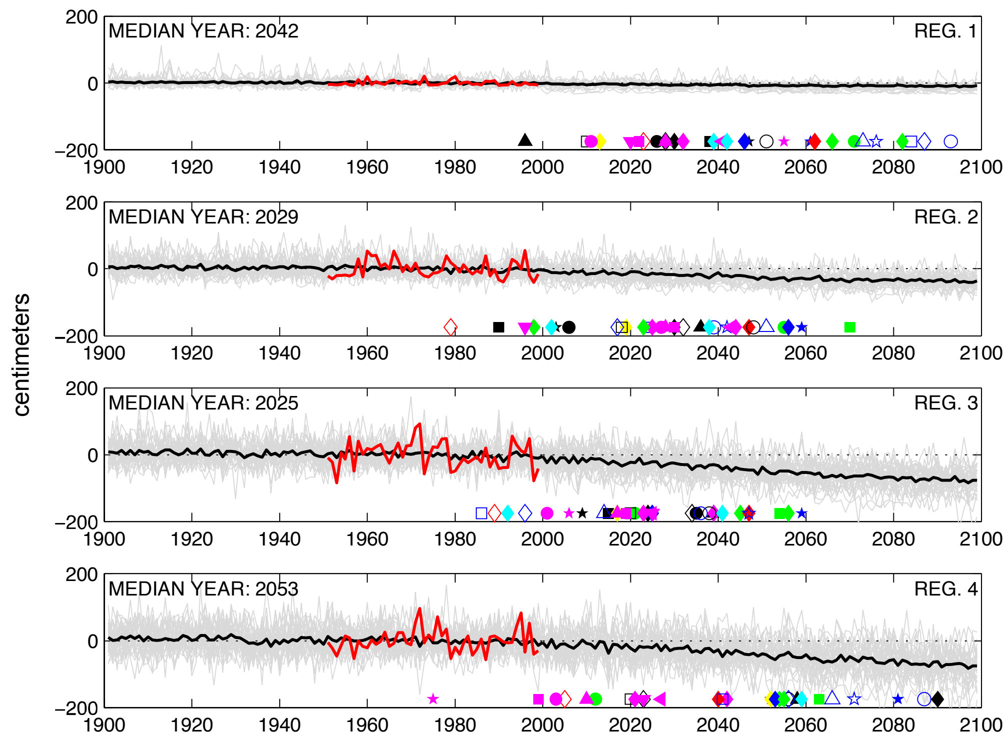


Figure 4.49. Time series of annual snowfall anomalies, relative to the 1951-1999 mean. The eastern coast of the United States (Regions 1, 2, 3, and 4) are shown. Individual model simulations are shown in gray, the multi-model ensemble mean is shown in black, and the observed data are shown in red. The detection years are shown along the x-axis and follow the same key as figure 4.5. “ND” denotes “no date” of emergence.

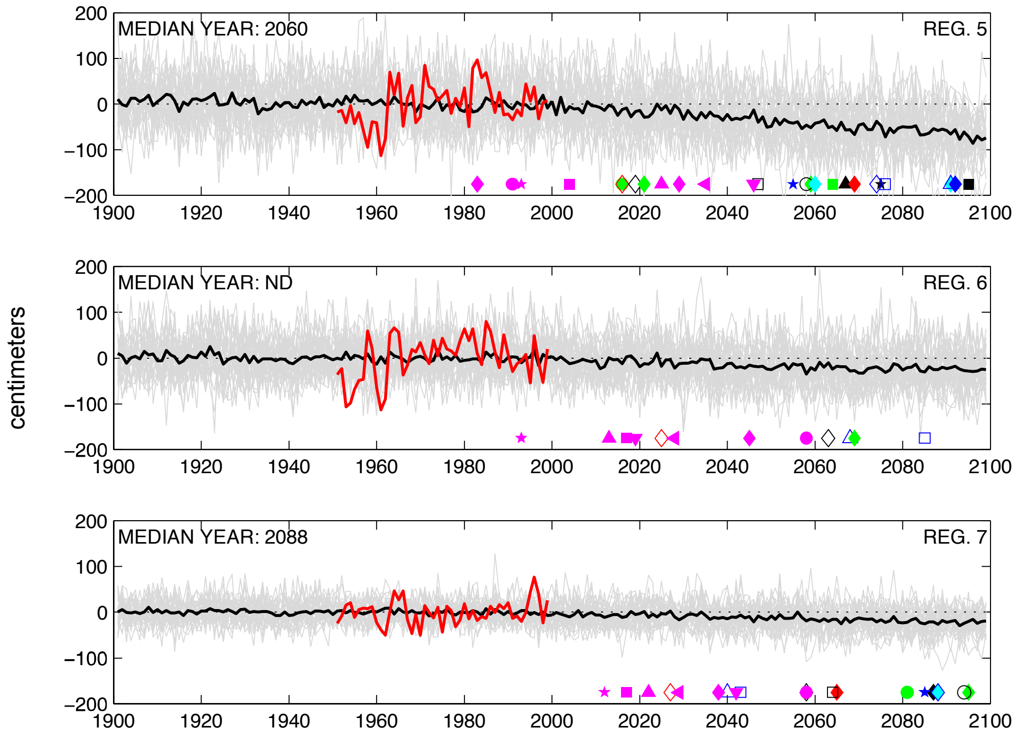


Figure 4.50. Same as figure 4.49, except for eastern Canada (Regions 5, 6, and 7).

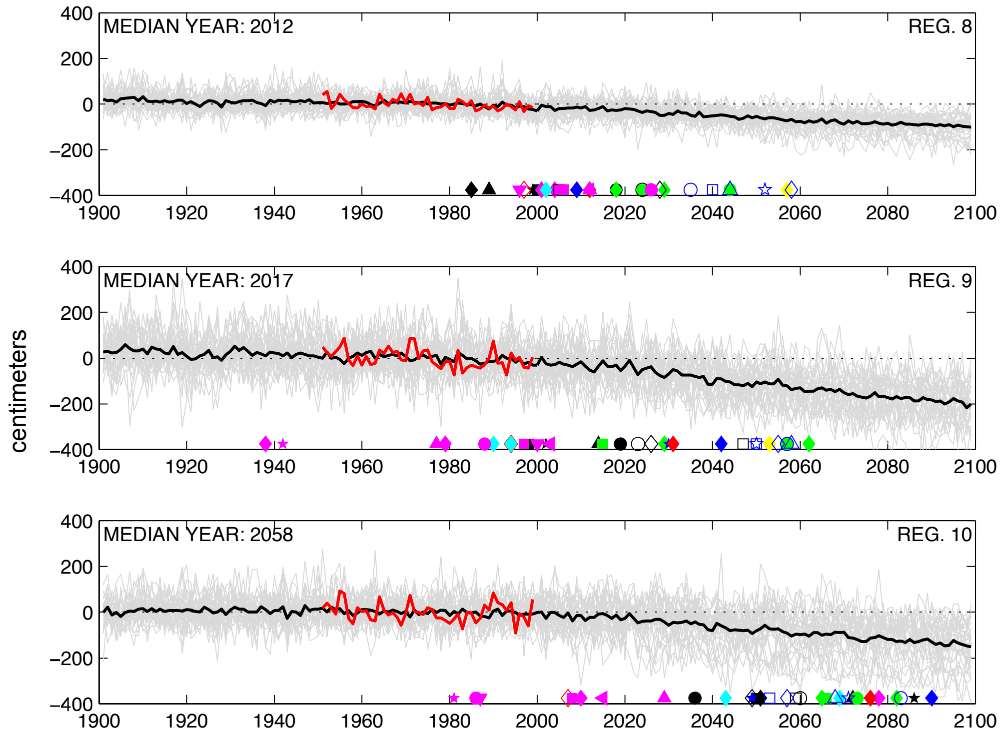


Figure 4.51. Same as figure 4.49, except for western North America (Regions 8, 9, and 10).

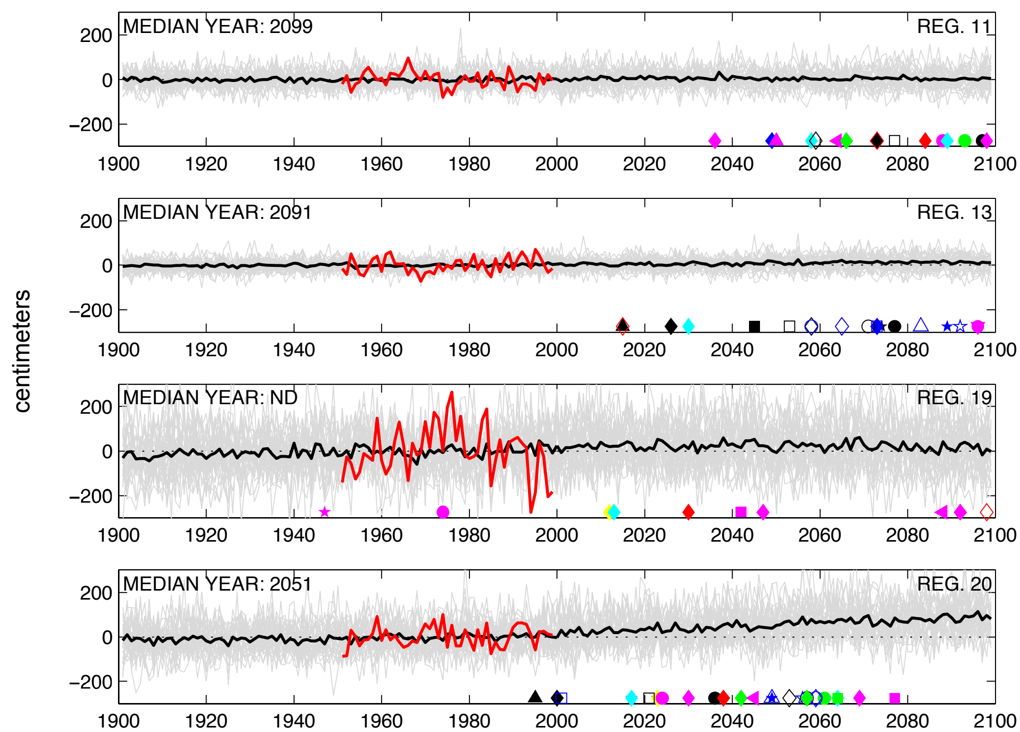


Figure 4.52. Same as figure 4.49, except for the Arctic regions of the Northern Hemisphere (Regions 11, 13, 19, and 20).

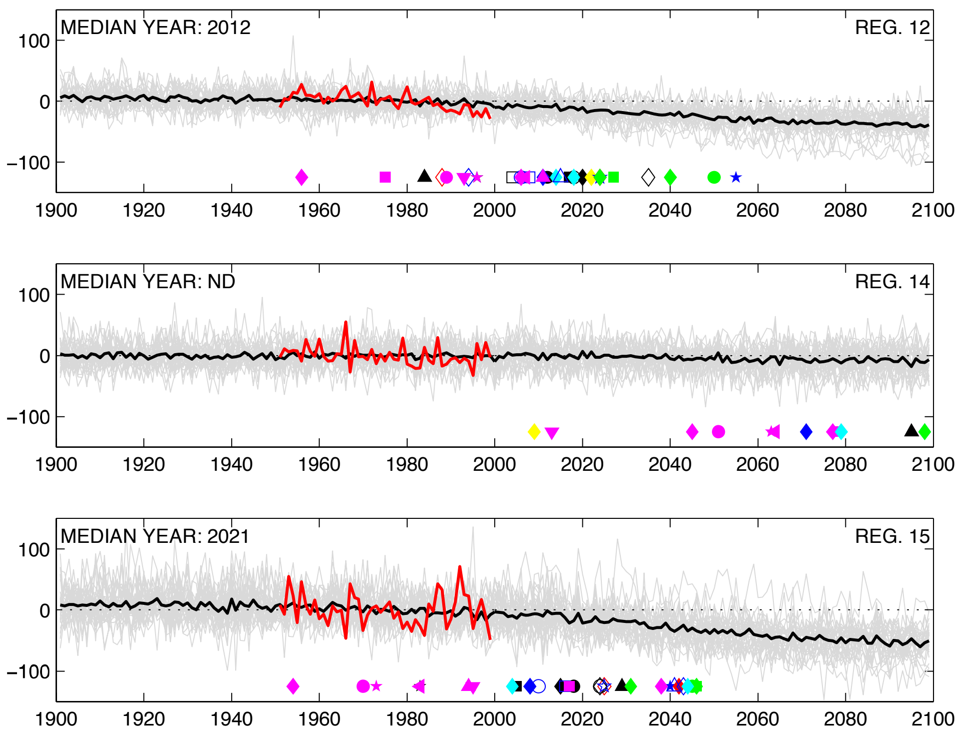


Figure 4.53. Same as figure 4.49, except for the mid-latitude regions of Eurasia (Regions 12, 14, and 15).

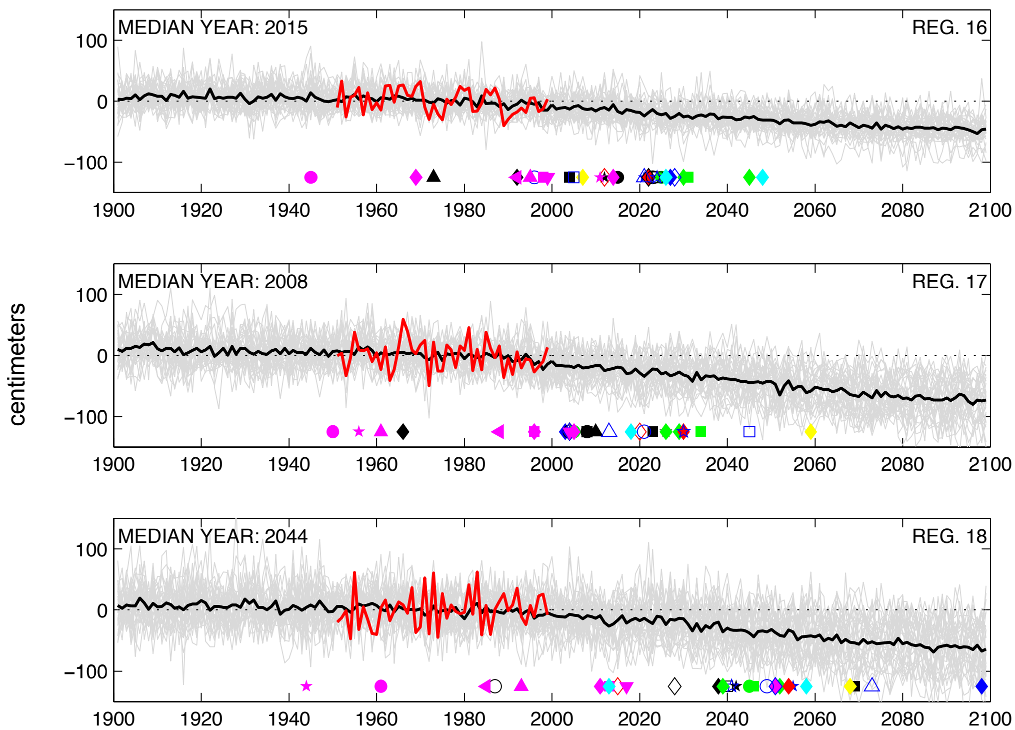


Figure 4.54. Same as figure 4.49, except for Europe (Regions 16, 17, and 18).

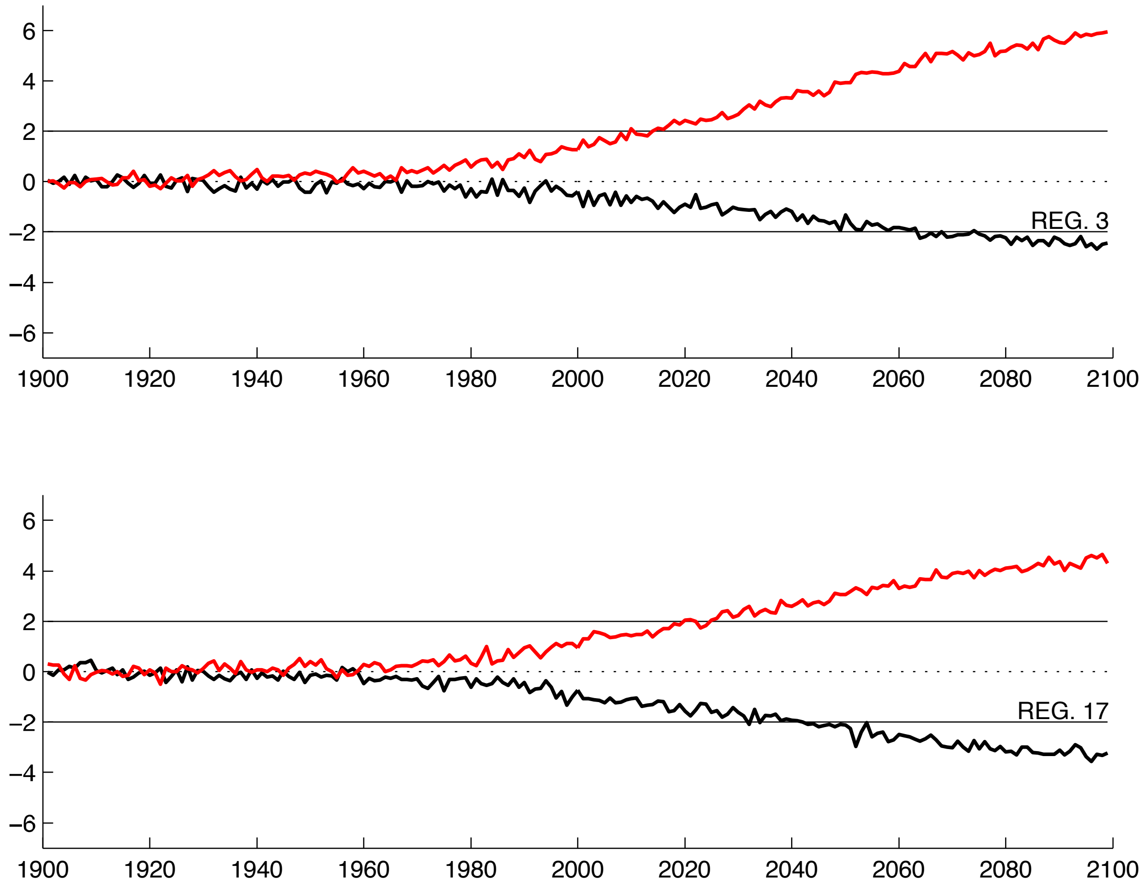


Figure 4.55. Mean standardized anomalies of temperature (red) and snowfall (black) relative to the 1901 to 1930 mean values from all of the individual model simulations.

Curriculum Vita

John P. Krasting

EDUCATION:
B.S. in Meteorology, Rutgers University, May 2003
Ph.D. in Atmospheric Science, Rutgers University, October 2008

EMPLOYMENT:

GAANN Fellow Rutgers University Department of Environmental Science	2007 – 2008
Teaching Assistant Rutgers University Department of Environmental Science	2006 – 2007
Graduate Assistant Rutgers University Department of Environmental Science	2005 – 2006
GAANN Fellow Rutgers University Department of Environmental Science	2003 – 2005