CHARACTERISTICS AND TRENDS IN NORTH AMERICAN SNOWFALL FROM A COMPREHENSIVE GRIDDED DATA SET

by

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ABSTRACT

Snowfall is an important component of the climate system because of its effects on the albedo, radiation budget, and water budget of the Earth. This study examines spatial and temporal trends in snowfall from a new gridded data set (1° by 1° lat/long) across the North American continent. The data is first quality controlled, after which several tests for temporal homogeneity are conducted. The interval from 1949 to 1999 is used for the analyses as this is the period with the most reliable data.

Results not only confirm previous literature but also expand our understanding of North American snowfall due to better spatial and temporal coverage of the data set. A detailed climatology of snowfall across the continent is constructed which highlights the Rocky Mountains, coastal Alaska, Great Lakes snowbelts and southeastern Canada as regions with the largest snowfall accumulations. A subsequent analysis of snowfall trends confirms the Pacific Northwest as a homogeneous area of decreasing snowfall to a spatial extent that, previously, has been undocumented. Several smaller regions of increased snowfall are seen in the Great Lakes, Alaska, Great Plains, Northeastern United States, and northern Canada. Univariate and multiple linear regression analyses suggest that snowfall variability across the continent is associated with several atmospheric teleconnection patterns, especially the Pacific Decadal Oscillation, Pacific/North American Index, and the Arctic Oscillation. In the Pacific Northwest teleconnection patterns are found to explain up to 70% of the variation in snowfall. Thus, a more complete understanding
of snowfall/teleconnection relationships is needed to better predict potential future changes in the snowfall climatology of the continent.
Chapter 1

INTRODUCTION AND LITERATURE REVIEW

1.1 Introduction

Snowfall is an important aspect of climate with significant impacts for human activities and wellbeing. It has a large effect on the surface albedo and therefore the radiation budget. Snowfall and its subsequent melt also contribute to the water resources of many regions. Outdoor recreational industries often depend on snowfall for their livelihood. Not the least important, large snowfall events have negative socioeconomic impacts on individuals and societies throughout North America.

While examining the winter of 1977-78 in Urbana-Champaign, IL, Changnon (1979) found that the average urban resident incurred losses and expenses of $93 on factors associated with extreme snow and cold conditions. Extrapolating from this figure, Illinois as a whole expended more than $11 billion. Personal consequences included inconvenience, anxiety, extra work, and injury. Financial losses also resulted from residential heating and repairs, vehicle maintenance and repair, snow equipment/clothing to be purchased, lost income due to work absences, extra medical costs, social costs, and extra food costs. Other effects of the extremely snowy winter were experienced but not measured, such as expenses for street and highway departments, industries, transportation systems, communication and utility companies, and loss of taxes by state and municipal governments. One economist
calculated a $4 million state income tax loss from work stoppages that resulted from just one of the 1978 storms (Changnon, 1979). Changnon also found that rural households were impacted by severe winter weather more than urban households.

In studying December 1989 conditions in the Lake Erie Snow Belt, Schmidlin (1993) found increases in costs for schools, snow and ice control on Interstate I-90 and massive costs and loss of shipments at lake ports. By contrast, ski centers saw a 50-100% increase in revenues while colleges, airports, agriculture, hospitals, urban mass transit, electric utilities, and government agencies experienced minor disruptions. (Schmidlin, 1993)

Concern over large snowfall events are not limited to the Midwest or Great Lakes regions. Branick (1997) found that of all the types of winter weather (heavy snow, ice, blizzards, thunder) heavy snow is the dominant type, occurring in 80% of winter weather events. Also, significant winter weather events can occur year-round (primarily in mountainous regions), but tend to occur almost daily somewhere in the United States from mid-November through March. The purpose of this research is to better understand the temporal and spatial variability of snowfall across the entire North American continent.

1.2 Snow Cover and Snow Water Equivalent

Beside studies on the human impact and costs of snowfall, two other topics relevant to this research have been addressed in the literature, snow cover and snow water equivalent (SWE). Snow cover is defined as the spatial extent of snow-covered ground, usually expressed as a percent of the total area covered in a given region (Glickman, 2000). Snow cover differs from snowfall in that snowfall is a measure of accumulated snow during a short period of time, usually one to two days,
while snow cover is the amount of ground covered by snow over a longer period of time before melting, usually weeks to months. Snow water equivalent is the amount of water that can be melted from a snowfall event. This measure is often identified as a liquid-to-solid ratio and can help describe the atmospheric conditions that produced the snowfall.

Several studies have examined regions in the United States and Canada with significant trends in snow cover and SWE (Brown and Goodison, 1996; Leathers and Luff, 1997; Frei and Robinson, 1999; Mote et al., 2004; Knowles et al., 2006). Brown and Goodison (1996) reconstructed snow cover over southern Canada and found no statistically significant trends over the time period 1915 to 1992. However, the data suggest that spring snow cover has decreased while winter snow cover increased. Regionally, the prairies show decreased snow cover in winter and spring since 1970. Leathers and Luff (1997) found low northeastern United States snow cover duration in the 1940s and 1950s, while from the 1950s to the end of the record in 1988, snow cover remained relatively constant varying around a mean value. Frei and Robinson (1999) found that North American snow cover is dependent on the longitudinal position of the North American ridge and the meridional oscillation in the 500mb geopotential height field. They also noted a general decrease in Northern Hemisphere snow cover extent since the early 1970s. Mote et al. (2004) found temporally decreasing SWE amounts on April 1st in the Pacific Northwest and Western United States, while snowfall water equivalent as a ratio of winter total precipitation was shown to be decreasing in the Western United States by Knowles et al. (2006), most pronounced in March due to warming. Many of these findings are...
similar to snowfall tendencies, reinforcing an obvious physical connection between snowfall, snow cover and SWE (Leathers and Luff, 1997).

Two studies that explicitly identify similar trends among these three variables are Karl et al., (1993), and Leathers and Luff (1997). Karl et al. found that mean monthly maximum temperature explained up to 78% of the variance in snow cover and snowfall. A change in temperature and humidity during snowfall also influences the liquid-to-solid ratio of the SWE (Karl et al., 1993), and temperature and snowfall amount during a year is highly correlated with snow cover duration (Leathers and Luff, 1997). Since these variables are related to snowfall, similar trends should be seen among all three.

The factors controlling each snowfall event are numerous, and sometimes last for only a few hours (Leathers et al., 1993). Small changes in meteorological factors such as temperature and moisture may create large differences in the amount of snowfall that occurs in an event. Regional locations of snowfall can vary further due to changes in the combination of synoptic-scale parameters, such as variations in storm tracks and regional temperature patterns (Bradbury et al., 2002). In contrast, snow cover studies, concerned with the presence of the snow cover over at least several days, incorporate a low temperature persistence factor (Harrington et al., 1987). Because of these differences, snowfall may be more representative of the short-term meteorological events that produce it.

Although there have recently been many snow cover, and snow water equivalent (SWE) studies, few have assessed the trends, climatological aspects, and climate change potentials of snowfall (IPCC, 2001, Leathers et al., 1993). Because of the impact snowfall has on human life and its implications for a changing climate
(Kunkel et al., 2007), an extensive investigation of the changing characteristics of snowfall in North America is useful.

1.3 Reasons for snowfall variability

As snowfall amounts change temporally, linear trend analysis may be used to approximate the magnitude and rate of change over time. The cause of these changes in snowfall is difficult to determine. Multiple meteorological factors interact to produce snowfall, which makes identifying the cause of any particular trend difficult. A tendency for increasing snowfall amounts may be the result of an increased snowfall frequency in a region, or the same number of events but higher snowfall totals. Karl and Knight (1998) developed an equation to estimate the proportion of any trend that is attributable to a change in intensity versus frequency. They found that there has been an approximately 10% increase in United States liquid precipitation since 1910, which is attributed mainly to increasing snowfall totals in the upper 10 percentiles of the precipitation distribution. Similarly, Scott and Kaiser (2004) found decreasing linear trends since 1948 in Pacific Northwest extreme snowfall events that correspond to an overall decrease in snowfall for that region. Scott and Kaiser also found a trend towards a decreasing number of snowfall events in that region.

Changnon et al. (2006) and Changnon (2006) examined characteristics of snowstorms across the United States. They found a North-South latitudinal distribution of high to low frequency values, except for higher values in the Great Lakes and mountain regions of the East and Western United States. An upward national trend in snowstorm occurrence is mainly due to increases in the frequency of snowstorms in the upper Midwest, East and Northeast United States. Decreases in
frequency were seen in the Midwest, South, and West Coast. A common feature seen in southern Canadian regions was significant increases in the frequency of snowfall events which corresponded to increases in winter snow cover (Brown and Goodison, 1996). On the other hand, reductions in spring snowfall events were associated with decreased snow cover duration and increases in maximum temperatures.

1.4 Observed Change in Snowfall

Two types of snowfall studies dominate the literature. One group documents trends in snowfall nationally and regionally. The second attempts to correlate changes in snowfall with possible forcing factors to help establish cause. In a snowfall climatology of the United States, Harrington et al. (1987) conducted a harmonic analysis that mapped the monthly percentage contribution of a station’s annual snowfall total. They found that latitude, elevation, and continentality combined to influence the length of the snowfall season at a given station. It was also shown that most of the United States experiences a peak in snowfall in February. Earlier peaks occur in the Pacific Northwest and around the Great Lakes, while the western high plains experience a late February and March maxima in snowfall.

Leathers et al. (1993) studied the temporal characteristics of United States snowfall to the east of the Cascade and Sierra Nevada mountain ranges from 1945 to 1985. While no region experienced decreases in monthly snowfall, two regions were identified with significant increases, the Great Lakes/Upper Midwest area and the High Plains. The Great Lakes/Upper Midwest region had positive trends in snowfall for December, January and February, the only region displaying any long-term tendencies in snowfall over the 40-year record. The High Plains region also experienced increased snowfall amounts; however, they only occurred in the month of
December. One possible cause of this increased snowfall may be the number and intensity of ‘Alberta Clipper’-type cyclones.

Leathers et al. (1993) used principal components analysis to identify other regions in which snowfall varied in a coherent manner. In addition to the Great Lakes/Upper Midwest region, the Central Plains and Southern Rockies, the Eastern Mid-Atlantic region through Southern New England, the Southern Mid-Atlantic to Central Plains, and the Northern Mid-Atlantic region and New England were identified as areas with spatially correlated trends in snowfall amount. The changes in each region were attributed to the movement of storm tracks and frequency of synoptic patterns.

Scott and Kaiser (2003, 2004) examined trends in snowfall across the United States over the last half-century. They looked at variables such as the annual total snowfall amount, number of snow days, length of snowfall season, mean temperature on snow days, and the number of snow days within the top ten-percentile. Several regions experienced coherent trends, such as the Pacific Northwest, the Ohio River Valley, and the lee of Lakes Ontario and Erie. In particular, Scott and Kaiser (2003, 2004) found that the Pacific Northwest has had significant decreases in the number of snow days, percent of normal snowfall, length of snow season, and number of extreme events over the period of record from 1948-2001. These observations were corroborated by Mote et al. (2004). In the Ohio River Valley there have been decreases in the snow day mean temperature and the length of snow season, however, annual total snowfall amounts have increased. These trends suggest that the physical characteristics of snowfall may have been changing over the past half-century to lighter snow with a higher liquid-to-solid ratio. To the lee of Lake Ontario, Scott and
Kaiser found a significant trend toward increasing amounts of snowfall and number of snow days. This finding is consistent with other studies of lake-effect snowfall (Leathers and Ellis, 1996; Smith and O’Brien, 2001; Burnett et al. 2003).

Kunkel et al. (2007) investigated trends in two subsets of the Cooperative Observer Program (COOP) network data. One subset was from 1930-2004 (1119 stations) and the other from 1900-2004 (233 stations). The stations grouped by high annual average in snowfall had upward trends over both time periods, and the stations grouped by low annual averages experienced significant downward trends. The high snowfall stations are located in the Great Lakes region and in some high elevation regions. Despite increasing trends for the high snowfall stations and no trends for the moderate stations, both exhibited decreasing trends from 1990-2004. The decreasing trends in low snowfall stations are likely associated with northward movement of the ephemeral snow line. However, the authors caution against the reliability of trend analysis due to inhomogeneities in the station records.

1.4.1 Temperature versus Precipitation

Changes in snowfall can be attributed to many factors, but on the most fundamental level, snowfall varies with temperature and precipitation. Serreze et al. (1998) identified two snowfall regimes based on the influence of temperature and precipitation. Snowfall is predominantly a function of total precipitation amount in the upper Midwest and Kansas-Nebraska where it is relatively cold and dry. In the warmer and moister Midwest, southern United States and the Northeast, the mean maximum temperature on precipitation days is the important variable. Hartley (1996) also found that sea surface temperature anomalies are associated with changes in
southern New England snowfall, mainly due to the influence of temperature rather than precipitation on snowfall total in this region.

This gives a significant amount of value to studies related to winter temperature and precipitation as possible indicators of changes in snowfall. Rodionov and Assel (2003) identified the Pacific Decadal Oscillation as having a negative association with winter temperature severity, where positive PDO events lead to a deep trough in the east and more cold air out breaks in the Great Lakes. While the authors do not give any implications for changes in winter precipitation characteristics, it is clear that this study highlights locations with likely variability in snowfall due to changes in temperature and moisture.

1.4.2 Great Lakes snowfall

Several studies have examined changes in the characteristics of lake-effect snowfall downwind of the Great Lakes. These studies are important because the large snowfall events in this region are potentially dangerous and detrimental to the local economy. A lake-effect snow event depends on a significant air/lake temperature difference, the extent of ice cover, wind direction, and wind speed (Eichenlaub, 1970). Because so many broad-scale factors contribute to these events, a changing trend in lake-effect snowfall may result from a regional-scale manifestation of larger-scale climate phenomenon, such as global warming (Burnett et al., 2003).

Ellis and Leathers (1996) and Leathers and Ellis (1996) found that “large snowfall increases have occurred across north-western Pennsylvania and western New York over the period [1950/51-1981/82]. The largest increases in snowfall have occurred during the mid-winter months (December-February), near the eastern shores of Lakes Erie and Ontario. Trends of greater than 1.7 cm year^{-1} are common in this
region, leading to linearly extrapolated snowfall increases of over 100 cm (85 per cent increases) over the 60-year period of record.” (Leathers and Ellis, 1996:1130), however, there are small and insignificant trends in all stations that are not located in lake effect regions.

To determine the cause of these changes, Ellis and Leathers (1996) and Leathers and Ellis (1996) identified nine main synoptic patterns that were associated with producing snowfall (at least 2.0 cm). Five were identified as lake-effect snowfall patterns and the remaining four were cyclonic patterns. These synoptic types were then used to model the changes in snowfall in order to account for the increases. The authors found that “Taken together, snowfall increases, due to frequency and intensity changes of lake-effect snowfall synoptic types, account for the majority of observed snowfall increases across the region. Frequency and intensity changes of the cyclonic snowfall types account for very little of the observed snowfall changes.” (Leathers and Ellis, 1996: 1132).

Similarly, Burnett et al. (2003) found that stations in the region affected by lake-effect snowfall show statistically significant increases since 1931, and that non-lake-effect stations show no significant linear trend. Burnett’s findings are supported with d$^{18}$O (CaCO$_3$) data for the eastern Finger Lakes of New York that have exhibited statistically significant trends toward lower values since 1915. The authors agree with Leathers and Ellis that increases in snowfall in the lake-effect region are not likely the result of changes in traveling cyclonic disturbances. However, Burnett et al. suggest that changes in the Great Lakes thermal characteristics are responsible for increased lake-effect snowfall. They argue that increased lake surface water temperature and a decrease in lake ice cover may lead to greater enhancement of
snowfall. Resolving these contrasting positions requires examining lake surface temperatures and lake ice cover to determine whether the cause of increases in lake-effect snowfall is due to changes in synoptic patterns alone or to Great Lakes thermal characteristics.

1.4.3 Teleconnections

A teleconnection is a linkage between changes in weather that occur in widely separated regions of the globe. It is seen as a correlation of field fluctuations among widely separated points. Teleconnections are most commonly observed in variability over monthly or longer timescales (Glickman, 2000). A common example of a teleconnection pattern is the El Niño Southern Oscillation (ENSO), where the ocean and atmosphere collaborate together. In determining the cause of changes in North American snowfall, these influences must be taken into consideration.

The Arctic Oscillation (AO) is defined as the leading empirical orthogonal function (EOF) of wintertime monthly mean Northern Hemisphere sea level pressure (Thompson and Wallace, 2000). A positive AO corresponds to low pressure over the polar region and high pressure at the midlatitudes (the negative phase is opposite). During the positive phase, oceanic storms in the Pacific are pushed to the North, meaning that the Western United States is dryer, while Alaska experiences wetter weather. East of the Rocky Mountains cold weather outbreaks are not as severe resulting in warmer temperatures. The AO varies between these two phases, but has tended to stay in the positive phase since the 1970s.

Similarly, the North Atlantic Oscillation (NAO) is a fluctuation in sea level pressure (van Loon and Rogers, 1978), with an index calculated from the time-averaged difference between the pressure at two centers of action (namely the Azores
high and Icelandic low). It is the leading wintertime mode of variability in the Atlantic basin and possibly a regional manifestation of the AO (Wallace, 2000; Thompson and Wallace, 2000). Positive NAO years have a stronger than normal subtropical high pressure center and a deeper than normal Icelandic low. This creates a larger pressure difference, which leads to stronger storms crossing the Atlantic Ocean, with a more northerly track. This is associated with warm and wet winters in both Europe and the eastern United States.

Bradbury et al. (2002) examined the position and amplitude of the northeastern United States trough to find relationships with regional sea surface temperatures, the NAO, and Pacific North American (PNA) teleconnections. They found that the NAO has a positive correlation with northeastern United States temperatures and a negative correlation with regional snowfall at some locations in the northeast. They also found that the negative NAO is associated with an eastward displacement of the Eastern United States trough. The authors concluded that the NAO is an important factor in northeastern climate, although the relationship is complex.

In a study of 27 New England stations from 1950 to 1992, Hartley and Keables (1998) examined winter climate and its associations with synoptic climate conditions. Previously, New England snowfall trends have been characterized by high snowfall in the 1950s and 1960s and high variability and lower snowfall in the 1970s and 1980s (Leathers et al., 1993), which is corroborated by Hartley and Keables. Low-snowfall composites show that these winters have larger variability and below-normal snowfall, due to some combination of less precipitation and higher temperatures. In high-snowfall composites they found higher amounts of cyclonic
activity along the eastern seaboard, in conjunction with more meridional flow in the mid-troposphere. These sea level pressure anomalies resemble the negative phase of the NAO, which the authors assess as the most important mechanism associated with the New England climate variability. As evidence, the high snowfall seen in the 1950s and 1960s coincides with the negative phase of the NAO and a positive PNA. The decreases in snowfall since the 1970s correspond to a positive phase in the NAO.

The El Niño/Southern Oscillation (ENSO) is a natural phenomenon in the equatorial Pacific that affects precipitation, pressure, and wind patterns in the tropics. It also affects the jet stream positions in the mid-latitudes, which in turn influences United States storm tracks. Several different indices of ENSO are in use to monitor the phenomenon. Some of the most common monitoring indices are the Tahiti/Darwin Southern Oscillation Index (SOI), the equatorial SOI, the Oceanic Niño Index (ONI) and the Japanese Meteorological Agency sea surface temperature index (JMA SST). Smith and O’Brien (2001) used the JMA SST index which is defined as a 5-month running mean of sea surface temperature in a region of the Pacific Ocean from 5°N-5°S, and 120°-170°W. If the mean exceeds 0.4°C above normal for 6 months or more, it is considered to be an El Niño event (Trenberth, 1997).

One study that examined the relationship between ENSO and snowfall found several regions across the United States that had significant correlations between ENSO phases and snowfall (Smith and O’Brien, 2001). During a cold phase of ENSO or a La Niña event, the Pacific Northwest experiences early and midwinter increases in all quartiles. In this region there is also a greater variability in snowfall amount during La Niña phases. Increases in early winter snowfall are also seen in the Great Basin. Accounting for these conditions, Smith and O’Brien (2001: 1188) state:
“Hypothetically, the increased dynamics and more frequent/stronger Pacific storms associated with the jet over these regions during a cold phase will combine with slightly cooler temperatures and result in increased snow.” Cold phases also show increases in late winter snowfall in the northern Great Lakes region.

During a warm phase of ENSO, or an El Niño event, the Northeast and northern Texas experience increased snowfall in midwinter. The Northeast also evidences increased variability in midwinter snowfall. However, the Pacific Northwest, Great Lakes region and northern Rockies show a pattern of decreased snowfall during warm phases. The Ohio Valley and Midwest experience a reduction of snowfall in both the warm and cold phases for the early and midwinter periods. Smith and O’Brien suggest that this is due to changes in the jet streams. A weak polar jet and limited amounts of moisture from the Gulf of Mexico can lead to decreases in snowfall.

These findings of Smith and O’Brien are corroborated by related SWE literature. While investigating the influence of interannual and interdecadal teleconnection patterns on April 1st SWE, Hunter et al. (2006) found that La Niña events as well as cold PDO phases (and their combination) are associated with increases in SWE, particularly in the Pacific Northwest.

Patten et al. (2003) examined the frequency of United States snowfall events (light, moderate, and heavy) with respect to ENSO phase. During a neutral ENSO year they found the highest occurrences of light snowfall events in the Great Lakes and Eastern United States, moderate events along the Great Lakes and mountain west, and heavy events occur in the west. The Northwest had the most consistent relationship with ENSO, where a negative ENSO (la Niña) results in increased
snowfall at all event sizes. It is also a symmetric relationship, where positive ENSO (el Niño) causes decreasing snowfall. Other areas with significant changes in snowfall frequency associated with ENSO (but more varying in event size) are the Northern and Eastern Great Lakes, Northeast Corridor, and New England.

The Pacific Decadal Oscillation (PDO) is a Pacific Ocean phenomenon that is identified by sea surface temperature (SST) anomalies. It is defined as a leading mode of multi-decadal variability in SSTs in the extratropical North Pacific [Hare, 1996; Zhang, 1996; Mantua et al., 1997; Minobe, 2000; Mantua and Hare, 2002]. It is computed as the time series scores associated with the leading principal component of SST in the Pacific, poleward of 20° N. The warm (positive) phase is characterized by cool SSTs in the central North Pacific, and warm SSTs along the West Coast of North America. It is also correlated with low sea level pressures in the northern subtropical Pacific and circulation anomalies that extend through the depth of the troposphere. In the United States there is decreased winter precipitation in the Southwest and from the Pacific Northwest to the Great Lakes region. These anomalies may result in changes in snowfall and snowpack. There are also wet periods in the coastal Gulf of Alaska and temperature changes in the spring. The cool phase of the PDO leads to the opposite anomalies of the warm phase.

While there is a lack of studies examining PDO and its relationship to snowfall, the relationship between PDO and the Pacific Northwest climate is seen in April 1st snowpack (McCabe and Dettinger, 2002; Hunter et al., 2006). In a principal components analysis (PCA), McCabe and Dettinger found that the leading mode of variability in snowpack is associated with the PDO and explains 45% of the variance. They identified an increase in storminess in the central North Pacific, associated with
the positive phase of the PDO. Hunter et al. (2006) coupled inter-decadal forcing mechanisms and found that PDO warm phase plus NAO negative phase is associated with decreasing April 1st SWE in the Washington/Oregon Cascades and in northwestern Montana/Idaho.

The Pacific North American pattern is characterized by a 500 mb trough in the North Pacific, a ridge over the Rocky Mountains and a trough in eastern North America. The PNA teleconnection is observed in North America as a fluctuation of mid-tropospheric mean flow (Wallace and Gutzler, 1981) that results in an intensification or damping of the typical longwave pattern. During positive years, flow is more meridional, while a negative index represents zonal flow.

Leathers et al. (1991) showed that strong relationships between the PNA and US temperature and precipitation exist. The southeast and northwest parts of the United States are areas of particularly strong correlations, as the meridionality or zonality of the 700 mb flow alters the location of the polar front jet and regional storm tracks.

The PNA is examined temporally by Leathers and Palecki (1992) who ascertained that over time the PNA shifts between modes of predominantly zonal and meridional mid-tropospheric flow. During the meridional flow phase a quasi periodicity was found on the order of ~40 months. They also found that SSTs in the Pacific and Asian landmass variables explain a large portion of the PNA variance.

Serreze et al. (1998) examined Eastern United States monthly snowfall data for 206 stations to determine linkages between snowfall, precipitation, maximum temperature on precipitation days, and low-frequency circulation patterns. They found that in the upper Midwest, snowfall is related to variations in precipitation while
in the Midwest, Southeast, and Northeast snowfall is associated with the maximum temperature on precipitation days. In a rotated PCA analysis, the PNA is identified as the 1st principal component. The PNA is most noticeable in January and February snowfall, where composite differences are significant over the Midwest, Southeast, and mid-Atlantic regions of the United States. For example, in the Southeast, snowfall increases during the positive phase of the PNA, as this produces a trough over the eastern United States, that leads to lower temperatures on precipitation days. Concurrently, the upper-Midwest would experience less snowfall during a positive PNA, related to changes in the frequency of precipitation.

In an observational/regional modeling study, Notaro et al. (2006) examined December climate in the Northeast United States and its connections with the PNA and NAO. They found that while a positive PNA index occurs during more frequent cyclone activity along a less dispersed track, total winter time precipitation is less than average. However, they found that Northeast snowfall was more related to the NAO than PNA and higher than normal lake-effect snowfall was produced during a positive PNA and negative NAO.

The results of Notaro et al. (2006) are similar to those found by Bradbury et al (2002). In this study (previously described) they found significant correlations between the PNA and the intensity of the eastern trough of the Pacific North American pattern, but found no significant relationship between PNA and climate variables in the Northeast. Hartley and Keables (1998) and Leathers and Ellis (1996) did not find any consistent correlations between the PNA index and snowfall.
1.5 Snowfall data

Snowfall is inherently difficult to measure. Mountain regions are typically biased by a preponderance of valley stations because of the location of most mountain cities (Karl et al., 1993). Few United States snow gauges have wind shields, preventing additional snow from blowing into, or being removed from, the measuring area (Groisman and Easterling, 1994). Point precipitation measurements that are used to calculate area-averaged values do not consider vertical gradients and are sensitive to large-scale inhomogeneities because their noise component has been decreased by averaging. This condition can result in network-induced large-scale decadal variations (Groisman and Easterling, 1994). Several human factors involved in the data collection may contribute to spurious trends (Kunkel et al., 2007). These include changes in observer, station location, observer practices, instructions to observer, observer’s adherence to instructions, time of observation, and the use of snowboards. Finally significant problems may be related to the data set used. Karl and Knight (1998: 234) point out that “for seasonal trends, even when the trends are statistically significant, differences among the datasets can be up to 4% per century.” These differences and the previously mentioned errors make it very important to choose an appropriate data set for the intended application.

Among the few commonly used data sets for snowfall analysis, limitations often encountered involve the number of reliable stations that are available and the length of record. Since all studies conduct quality control procedures to insure the best possible selection of stations, the number of useful stations is often significantly reduced. Of the previously examined studies, Kunkel et al. (2007) use the largest number of United States stations, 1119. Because they are looking at the conterminous United States as a whole, spatial coverage would require a larger number of stations.
The largest number of Canadian stations, 1107, is used by Groisman and Easterling (1994). However, during the early part of the record, it is not uncommon for there to be large gaps in the spatial coverage, despite large numbers of reporting stations. The least number of stations used is by Burnett et al. (2003) who selected 25 stations near the region of lake-effect snow enhancement. The remaining studies fall within this range.

The period of record is another factor affecting station selection since many data sets lack a long-term spatially coherent record. The longest period used in any of the studies already discussed is Kunkel et al. (2007), who used data from 1900-2004. This long period of record could be used when the number of stations in their study was reduced from 1119 to 233 and large gaps in spatial coverage were accepted. The shortest period of record used in those reviewed was from 1950/51 to 1989/90 by Ellis and Leathers (1996).

The final factor affecting the selection of an appropriate data set is temporal resolution. Snowfall data typically have daily to seasonal or annual temporal coverage. Many studies use monthly snowfall data. This resolution level is useful when identifying general trends in snowfall amounts, but a higher resolution data set is preferred when examining the frequency of snowfall events, changes in the number of extreme snowfall events or synoptic patterns associated with snowfall. This higher resolution is ideal to identify changes in the short-term weather conditions that influence and produce snowfall.

1.6 Present Study

A review of the literature shows there are no snowfall trend studies that focus on entire North American continent. Of the studies reviewed, some were further
constrained by only focusing on particular regions, over limited time periods. There is also a lack of studies examining the connection between continent-wide snowfall and teleconnection patterns. The present study seeks to address these limitations using a relatively new data set with better United States/Canadian coverage over a longer time period that may reveal greater insight into the causes of temporal fluctuations in snowfall.

The data set used in this study is a 1° x 1° North American snowfall gridd product with a daily resolution and period of record that extends from 1900 to 2001 (Dyer and Mote, 2006). Compared to data from previous studies, this set is ideal. Its spatial coverage enables examination of the entire North American continent over the last century, and the higher daily resolution can facilitate the identification of shorter-term changes in snowfall. The study will use this new data set to establish the existence of long-term trends in snowfall, and test the correlation between snowfall trends and ocean/atmosphere teleconnection patterns. More generally, the study will determine the utility of the new data set in comparison with existing sets.

Using this new data set, the study will attempt to answer specific questions raised by the literature review: are there trends in the frequency and intensity of North American snowfall, and are these trends both linearly and spatially coherent? Finally, can any regional snowfall trends be correlated with large scale teleconnection patterns?
Chapter 2
DATA VERIFICATION RESULTS

2.1 Data Description

The data used in this study are from a $1^\circ \times 1^\circ$ interpolated snowfall data set (Dyer and Mote, 2006). These data were first interpolated onto a $0.25^\circ \times 0.25^\circ$ grid from United States National Weather Service (NWS) cooperative stations and the Canadian daily surface observations which were quality controlled using criteria from Robinson (1989). A map of the input station locations and the grid locations is shown in figure 2.1. The interpolation was completed using the Spheremap spatial interpolation program, developed by Willmott et. al (1984). The grids extend from $25^\circ$ to $71^\circ$ North latitude and from $53^\circ$ to $168^\circ$ West longitude. The period of record is 1900-2000 with a daily temporal resolution. Values calculated from the reporting stations in a grid for each day are maximum snowfall, minimum snowfall, median snowfall, mean snowfall, standard deviation, number of valid points over land, and number of invalid points over land. In this study, mean snowfall values were chosen to approximate the daily value at each grid point.
Figure 2.1  A) United States and Canada station locations for stations used in the interpolation, and B) Grid point locations, where each point represents the bottom left corner of a grid cell.
2.2 Station Density

It is important to examine the density of reporting stations used when interpolating station data to a grid. If there are no reporting stations in a grid cell, the interpolation algorithm must use data that can be a significant distance from that cell, making its reliability for analysis questionable.

Station density is examined by inspecting the number of stations contributing to each grid cell, which is given as one of the variables in the snowfall grid files. Figure 2.2 shows a time series of the total number of stations per season over the entire study area, where a snow season is July 1 to June 30. There is a steady increase in the number of stations from 1900 to about 1945 and then a large jump from a little over 2000 stations to 8500 stations around 1950. This large jump is not due to a large increase in the true number of stations, but in the availability of digitized data for the existing stations. There is a small increase until the early 1960s and then a decline in the number of reporting stations until the end of the record. The maximum number of reporting stations (approximately 9000) was reached in the early 1960s. This time series clearly shows that any interpolation before the mid 1940s is based on less than one-fourth of the data that are used for the post 1940s interpolation.
Figure 2.2  Total number of reporting stations per snow season
The percentage of total possible grid cells with data each season is shown in figure 2.3. This figure is produced by dividing the number of grid cells with any reporting stations by the total number of grid cells in the data set. This figure shows similar trends as figure 2.2. From 1900 to approximately 1977 there is an increase in the percentage of grid cells with reporting stations. An interesting difference in this figure is that even at the highest percentage only 50% of the grids have reporting stations located within them. Thus, at least, half of the data are based upon interpolation. This is a potential problem for daily precipitation data, where increasing distances from the closest stations greatly increase interpolation error (Bussieres and Hogg, 1989). This is a fact that must be kept in mind during analysis of the data.
In order to examine the distribution of reporting stations through time, maps were created that display the number of reporting stations per grid cell every 10 years (Figures 2.4 to 2.14). During the snow season of 1900, the Great Lakes region of Canada has the most reporting stations. There are several grid cells with as many as 7 reporting stations per grid in this region. There is a swath of grids with reporting stations in the United States central plains with 1 to 3 reporting stations per grid. The majority of Canada, Alaska, and the Western United States is without reporting stations.

During the 1909 snow season there are more stations in the central plains, south into Texas as well as in the Ohio river valley. There are still no grids with more than 7 reporting stations, and the majority of Canada, Alaska, and the western United States have few grid cells, if any, with reporting stations.

The 1919 snow season map has an increase in the maximum number of reporting stations in several of the grids. At this time, the maximum values are as high as 12 stations, appearing in a grid cell near Lake Ontario and a grid near Seattle Washington. The Ohio River Valley region and the Great Plains have an increased number of grids with reporting stations, as well as a swath into Southern Canada (specifically British Colombia, Alberta, Saskatchewan and Manitoba).

The mean number of stations in each grid cell for the 1929 snow season has a much more homogeneous distribution across the United States. The maximum value is 16 stations in one grid, which occurs in Washington. The Canadian station distribution also grows and extends to the North.
The station distribution continues to improve into the 1939 snow season. The maximum number of reporting stations found per grid is 15 and there are no large gaps without reporting stations in the United States or Southern Canada.

A large jump in the number of reporting stations occurs in approximately 1945 (as described in figures 2.2 and 2.3) which accounts for the increase in grids with reporting stations in the 1949 map. There is almost a solid coverage of grids with reporting stations from Southern Canada south. The grids with the maximum number of reporting stations are in the Eastern United States with a large area of 10 reporting stations or greater in each grid and a few peaks of grids with greater than 30 reporting stations. There are also some areas in California with peaks of greater than 20 reporting stations. There are scattered grids with reporting stations in the Northwest Territories and Nunavut. The maximum values are from 60 to 70 reporting stations.

In 1959 some grids with reporting stations are gained in Canada, but lost in the eastern United States. The maximum values are now 50 to 60 reporting stations, and occur in California and the Northeast United States.

The rest of the period experiences relatively small changes in station density compared to the first half of the record. The station distribution continues to increase in Alaska and Saskatchewan in the 1969 snow season. The eastern half of the United States and California continue to have the largest density of reporting stations. However, in the 1979 snow season the number of reporting stations starts to decrease in many areas, but the number of grids with any reporting stations is nearly the same as the 1969 map. Maximum values are reduced to the range of 30 to 40 stations. There is little change in the 1989 map, which looks very similar to 1979. The maximum number of reporting stations continues to decrease to the 20 to 30 station
range in the 1999 map. Again, the number of grids with any reporting stations has not changed significantly from the last map.
Figure 2.4  The mean number of stations in each grid cell for the 1900 snow season.
Figure 2.5  The mean number of stations in each grid cell for the 1909 snow season.
Figure 2.6  The mean number of stations in each grid cell for the 1919 snow season.
Figure 2.7  The mean number of stations in each grid cell for the 1929 snow season.
Figure 2.8  The mean number of stations in each grid cell for the 1939 snow season.
Figure 2.9  The mean number of stations in each grid cell for the 1949 snow season.
Figure 2.10  The mean number of stations in each grid cell for the 1959 snow season.
Figure 2.11  The mean number of stations in each grid cell for the 1969 snow season.
Figure 2.12  The mean number of stations in each grid cell for the 1979 snow season.
Figure 2.13  The mean number of stations in each grid cell for the 1989 snow season.
Figure 2.14 The mean number of stations in each grid cell for the 1999 snow season.
2.3 Period of Record

The period of time that a given cell had at least one station within its borders is used to calculate the period of record for each cell. Two different measures of period of record are used in this research. One involves calculating the difference between the first and last season when stations are present in a given cell. The second is by counting the number of seasons when reporting stations were present in a cell. This is an important aspect of determining the reliability of the grid cells.

Figure 2.15 shows the date of the first season with reporting stations in a grid. The earliest starting seasons are in the Great Lakes region of the United States and Canada, into the Ohio River Valley, and a swath along the United States/Canada western boarder. These areas have starting seasons in the early 1900s. The regions with the next earliest starting seasons are along the eastern coast of the United States, and the West Coast. Reporting stations come “online” in these regions in the 1920s. Green and blue colors in the figure indicate a later starting season, from the 1940s and later. Grid cells scattered over North America and an area along central Canada and almost all of Alaska date to this period. Grids without color have never had any reporting stations within the period of record.

The last season a grid cell had station data is shown in figure 16. The majority of grids with reporting stations have those stations throughout the 1999 season. There are a few scattered grids that lose their reporting stations earlier, and are predominately located in Northern Canada.

The difference between the last season with reporting stations and the first season gives a measure of the period of record for grids with reporting stations.
Figure 2.17 shows the period of record for all grid cells. The area with the longest period of record is in the Great Lakes region, into the Ohio River Valley, parts of the Great Plains and a swath along the United States/Canada western boarder. These regions have periods of record close to 100 seasons. The eastern coast of the United States and the West Coast have periods of record of approximately 70 to 80 seasons. Scattered grids all over North America, along central Canada, and over almost all of Alaska have periods of record of 50 years or less.
Figure 2.15  The first season with reporting stations.
Figure 2.16   Last season with reporting stations.
Figure 2.17  Period of Record with Station Data. Calculated as last season minus first season from 1900 to 1999.
In order to determine the snow seasons and grids most reliable for further analysis it is useful to look at maps of grid cells based on their period of records. Figure 2.18 through 2.22 show all the grid cells with period of records of 100, \( \geq 90 \), \( \geq 75 \), \( \geq 50 \), and \( \geq 25 \) seasons. A compromise must be made between choosing the longest period of record possible, and as many grid cells with reporting stations as possible.

Figure 2.18 includes the grid cells with 100 seasons as period of record. The grid cells that qualify are located in the Great Lakes, Ohio River Valley, and the United States/Canada boarder area as mentioned previously. Figure 2.19 includes grid cells with 90 or more seasons as period of record. More grid cells are located in the central plains and Southwest United States. The northeastern United States, central United States and Southern Canada have more grid cells with stations when the period of record is lowered to 70 years or greater. However, the western United States, Alaska, and most of Canada are still without reporting stations. Figure 2.21 includes grid cells with 50 seasons or greater period of record. This solidly covers the U.S. and southern Canada. For this period of record there are a number of grids with reporting stations in Alaska as well. Figure 2.22 shows grids with 25 or greater seasons with reporting stations. While figure 2.22 captures nearly all the grids that have reporting stations, 25 seasons is not a suitable period of record for the application of this study. A period of record of 50 seasons (Figure 2.21) is deemed an acceptable compromise between spatial and temporal coverage. For this reason the 1949 season will be the starting season for the majority of the analysis done in this study.
Figure 2.18  Grids with 100 seasons period of record. Calculated as last season minus first season from 1900 to 1999.
Figure 2.19  Grids with greater than or equal to 90 seasons period of record. Calculated as last season minus first season from 1900 to 1999.
Figure 2.20  Grids with greater than or equal to 75 seasons period of record. Calculated as last season minus first season from 1900 to 1999.
Figure 2.21  Grids with greater than or equal to 50 seasons period of record. Calculated as last season minus first season from 1900 to 1999.
Figure 2.22  Grids with greater than or equal to 25 seasons period of record. Calculated as last season minus first season from 1900 to 1999.
A histogram showing the period of record for all grid cells is displayed in figure 2.23. There are 1166 grids that never have any reporting stations, and 338 grids that have reporting stations for all 100 seasons. This figure shows that a large number of grids have periods of record greater than 50 years.

![Histogram of grid cell period of record.](image)

**Figure 2.23** Histogram of grid cell period of record.

2.4 Seasonal Count

Because the period of record is calculated by taking the difference between the last year with reporting stations and the first, it does not take into account periods lacking reporting stations in the middle of a grid cell’s record. In order to
account for missing seasons, a count of the number of seasons with reporting stations was completed for each grid cell.

Figure 2.24 is a histogram showing the number of grid cells with a reporting station for each period of record. There are 1165 grids that never have a season with a reporting station. On the other extreme, 257 grids have 100 seasons with reporting stations. This indicates that several of the grid cells with data that span the entire period of record have at least some missing data. However, the figure indicates that the majority of stations that have at least one year with a reporting station, have at least 50 years.

![Histogram of grid cell seasonal count.](image)

**Figure 2.24** Histogram of grid cell seasonal count.
2.5 Differences in the number of reporting stations between significant seasons

Due to data reporting and digitization practices, there are several seasons during the period of record that experience significant changes in the number or distributions of reporting stations. These seasons are now examined more closely.

Figure 2.25 shows the differences between the 1949 and 1948 seasons. A large area with increased number of reporting stations is found along the East Coast of the United States. Parts of Southern California and Alaska also show increases in the number of reporting stations. In the Ohio River valley there is a decrease in reporting stations in some areas. Much of the increase is associated with the availability of data in a digitized format.

Because of changes in the number of reporting stations and the subsequent change in station distribution, the time period for analysis of trends and association with teleconnection patterns must be carefully considered.
2.6 Within grid cell station densities

Inhomogenities of within grid cell station density will have an important effect on the subsequent analysis of trends. To illustrate the potential problems, graphs are presented showing changes in station density within grid cells.

Spurious trends are a likely problem where station densities change rapidly within a grid cell. In the following two figures the grid has no reporting stations at the beginning of the record and has purely interpolated values. As the number of reporting stations in the grid cell increases the snowfall values for the grid cell change dramatically.
In figure 2.26 the timeseries of the number of snowfall events per snow season (July 1 to June 30) at grid 121°E longitude/ 46°N latitude is plotted. The number of reporting stations within the grid is also plotted, along with the trend line. In the first 25 seasons of this record there are no reporting stations within the grid cell. The number of events are around 50 per season and there is somewhat low variability. In the late 1930s and 1940s a few reporting stations come into service within the grid cell, and the number of events increases in magnitude and variability. In the late 1940s the number of reporting stations increases dramatically and the number of events increases while the variability decreases. This suggests that the low values at the beginning of the record are not representative of the grid 121°E longitude/ 46°N latitude. Instead, these low values were associated with the values from surrounding grid cells and the interpolation algorithm. The approximate mean elevation of this grid cell is 4,500 feet, and the elevation of the reporting stations used outside of this grid cell influence the grid’s mean value early in the record. This figure suggests that the trend line is spurious for this grid cell, and that a trend should be calculated only after the 1948 jump in station distribution.

There are similar issues present in figure 2.27. At this grid point (114°E longitude/ 36°N latitude) there are zero reporting stations before the late 1940s. During this time the total snowfall per season is high, but as the number of reporting stations in the grid increases, the total snowfall per season decreases. This grid cell has an approximate elevation of 2,500 feet and is located near the Rocky Mountains. This is likely a sign that the grid from which the interpolation is drawing station data is either not near by, or is near by and has higher elevation stations. As the number of reporting stations in the grid increases, the average for the grid decreases. This also
gives a spurious trend, which is negative. It is apparent that the station density must be closely examined before any trend analysis is complete.

Figure 2.26  Timeseries of the number of snowfall events per snow season for grid cell 121°E longitude/ 46°N latitude.
2.7 Unreliable grid cells

It is essential that the grids used in this analysis are deemed reliable. A grid cell with several large changes in the number of reporting stations would be considered unreliable for use in a trend analysis. Several different methods were tested to find a suitable way to screen out unreliable grid cells.

Figure 2.28 shows the criteria that are chosen to identify unreliable cells. All of the grid cells represented in black have 10 or more seasonal changes in the number of reporting stations that are greater than or equal to 10% of the number of stations in the previous season. Differences are calculated by subtracting the previous
season’s number of reporting stations from the current season’s number of reporting stations. The absolute value of the differences are compared to the average number of stations in the previous season. This method provided the best qualitative balance between the reliability of the grid cells accepted and the period of record.

The grid cells that have no reporting stations during the entire period of record are also deemed unreliable. “Data” at these locations are solely a result of interpolation. In some cases, the distance to the nearest grid cell with a reporting station can be several grid cells away. A distance of several hundred kilometers is significant when dealing with daily precipitation data (Bussieres and Hogg, 1989). Grid cells with no data are shown in gray in figure 2.28. In this figure all of the unreliable grids are marked.
Figure 2.28  Grids with 10 or more seasonal differences that are greater than or equal to 10% of the number of stations over the period 1949 through 1999. Gray cells indicate no station data in the cell for the period of record.
2.8 Summary and Analysis

Several potential data problems are addressed in this chapter. Figures 2.2 to 2.13 show that the number of stations vary greatly with time. Also, it was shown that only 50% of grid cells have a reporting station within them at any given time. Spatially, before the 1940s there are few regions that have a continuous coverage of grid cells containing stations within them outside of the eastern half of the United States. After the late 1940s the United States and southern provinces of Canada are solidly covered with grid cells that have stations in them during at least part of the snow season. This distribution changes little in the remaining time. Examination of the grid cells’ period of record also confirms these results. The display of station timeseries graphs in section 2.6 illustrates how changes in within grid cell station density can effect trend analysis and produce spurious trends.

These data distribution problems are likely a function of several things. It is likely that station data has historically been available in locations where there are people to record it. This is seen early in the period of record, when there is little data in the northern parts of North America. The number of stations at which observations take place may also be dependent upon the amount of government funding available. This would surely affect the ability to maintain a network and add additional stations. Finally, earlier data that has been recorded on paper is still, in many cases, in the process of being digitized, and was not used in this interpolation (Midwestern Regional Climate Center, 2007).

The reliability of this data set is maximized by selecting the time period with the most consistent data distribution possible. This reduces the amount of
variability and bias in the climate record due to uneven spatial coverage, which has previously been explored in Willmott et al. (1991) and Willmott et al. (1994). In the current study the time period of 1949 to 1999 appears to have the most consistent data distribution over the longest time period, which is the time period chosen for subsequent trend analysis.

Further efforts to extend the usefulness of these data entail deeming some grid cells unreliable and leaving them out of the trend analysis. This includes grid cells without any stations used in the interpolation as well as grid cells with a large number of stations coming on- or off-line in a small amount of time. After these data verification analyses are complete, a reliable interpolated data set was available for subsequent analysis.

This chapter illustrates the importance of data verification before analysis can begin. There were several things that needed to be considered and constraints that needed to be applied before this data set could be considered reliable and ready for analysis. It is all too common that climatologists must use whatever data is available, and serious efforts must be made to reduce interpretation errors due to a lack of knowledge of the data used for analysis.
Chapter 3  
NORTH AMERICAN SNOWFALL CLIMATOLOGY

3.1 Monthly Climatology

This chapter examines the climatology of North American snowfall, by month, for the time period 1949 to 1999, in order to assess the accuracy and robustness of the gridded product. Monthly distributions of mean snowfall, coefficient of variation, maximum snowfall, and minimum snowfall were produced from the gridded data. The mean snowfall value was calculated by totaling the daily values for a given month, then calculating the mean value for all of those months in the period of record. The maximum and minimum values mapped are the highest and lowest monthly totals at each grid cell during the period of record. The coefficient of variation is the standard deviation of the snowfall values divided by the mean for each grid cell, and it shows the dispersion of the data around the mean. Because coefficient of variation is greater with a smaller mean, this variable highlights ephemeral snow regions. When seasonal snowfall statistics are calculated, a snow season from July 1st to June 30th is used. Blackout grids, mentioned in Chapter 2, are not applied to these maps because trends are not being calculated.

In figure 3.1 the mean, coefficient of variation, maximum, and minimum January snowfall are shown. For mean snowfall the highest gridded values are around 950 to 1000 mm, occurring in the Canadian Rockies, with other high values in the United States’ Rocky and Cascade Mountains. Maximum values in this area surpass
2,000 mm of snowfall. There is a second area of high snowfall values in the Maritime Provinces/Great Lakes region. The mean snowfall in this area is as high as 1000 mm in January. Maximum values near 2,000 mm occur downwind of the Great Lakes, as clear indicators of lake-effect snow regions. The coefficient of variation shows that the data varies little around the mean for most of North America. Regions along the ephemeral snow line have the largest values representing a large standard deviation and a small mean value. Minimum values show snow-free areas throughout the southern half of the United States. The areas with the largest minimum values are in the Northeast/Great Lakes region, which receives as much as 360 mm during the lowest January monthly totals.

February’s snowfall is examined in figure 3.2. Mean values are only slightly smaller than January’s; with the highest mean snowfall around 950 mm along the gulf of Alaska. Other large areas of snowfall in the 600 to 700 mm range occur in the Rocky Mountains, Sierra/Cascade Mountains, and in the Northeast/Great Lakes regions. February’s coefficient of variation map shows the largest values in the ephemeral snowfall regions (southern tier of the United States), where they would be expected. Maximum monthly values occur in Southeastern Alaska, with values as high as 3,000 mm. Minimum February values leave large areas in the contiguous United States without snow, while the largest minimum values are centered in the Northeast/Great Lakes region.

Figure 3.3 details March snowfall. During March large homogeneous regions of high mean snowfall values are generally in the same locations and mean snowfall range as in February (600mm to 700 mm in the inter-mountain west, and Eastern/Great Lakes). However, in March higher mean values have spread into the
central and northern great plains. As in previous months, the coefficient of variation is largest along the ephemeral snow line which has extended slightly north in this month. Minimum values show much of the United States without snowfall, except in the Rockies and the Northeast/Great Lakes region. Minimum values in the Maritime Provinces are has high as 160 mm, which indicates that there is nearly always snow in the Maritime Provinces in March. This is also an area of high values in the maximum monthly map. Maximum March values are as high as 2,000 mm in the Maritime Provinces, southern Canada, and the Sierra/cascade Mountains.

April snowfall maps are shown in figure 3.4. Mean values as high as 500 mm of snowfall are located in the southern and central Rockies, and in maritime Quebec and Labrador. The coefficient of variation shows that these values have a relatively low standard deviation. It also helps to detail the northward progression of the ephemeral snow line across the United States. Maximum snowfall values are the highest in Quebec and Labrador Province, as well as at scattered locations in the High Plains and Alaska. These values are as high as 2,000 to 2,200 mm. Minimum April snowfall shows most grid cells without snow in the United States and southern Canada with the exception of the Rocky Mountains and around the Great Lakes.

Figure 3.5 displays May snowfall statistics. On average 100 mm of snow or less falls across most of the United States, with peaks near 150 mm in some of the higher elevation regions in the Rockies. In Northern Canada there is as much as 150 mm of snowfall for May, with the highest mean snowfall occurring around Baffin Island and Foxe Basin (150 to 200 mm). The highest coefficient of variation is in the eastern United States along the ephemeral snow line where snowfall events are sporadic occurrences during this time of year. Maximum May snowfall indicates
some amount of snowfall in most grid cells except the southeastern United States. The highest maximum values occur around the Hudson Bay in Canada, nearly 450 mm. Minimum May snowfall only shows northern Canada with regions that receive any snowfall at all, and in most grid cells the range is from 0.1 to 20 mm.

June, July, August, and September snowfall maps are shown in figures 3.6 through 3.9. These summer months have very few grid cells with appreciable mean snowfall. The coefficient of variation is large across the continent, indicating that snowfall values have a large standard deviation because of the few snowfall events. Maximum values are in the 0.1 mm to 300 mm range, except in September, which has as much as 600 mm of snow North of Hudson Bay, along the Rocky Mountains, and in Alaska. For all of these months the minimum value for all grids is at or near zero snowfall.

Figure 3.10 shows October snowfall statistics. Mean values above zero cover a large portion of the United States, with the highest values observed in Northern Canada, from 300 to 500 mm. The coefficient of variation is still large, but is decreasing as more regions begin to receive consistent snowfall amounts. Snow is found in most grid cells in maximum months with values as high as 1,200 mm to 1,500 mm in Alaska, the Yukon, and the Rocky Mountains. Minimum October snowfall still shows most of the continent with zero values except Northern Canada, parts of Alaska, and a few grid cells in the Rocky Mountains.

In figure 3.11 November snowfall is examined. The Southeastern United States is the only region with mean snowfall values equal to zero for this month. The highest mean values occur in the Quebec and Labrador Provinces, and the inter-mountain west with slightly smaller values in the Great Lakes region, Southeastern
Alaska, and in the Canadian Rockies. These mean values are as high as 500 mm to 600 mm. The coefficient of variation is low for all of Canada, and the northern half of the United States, but the values increase towards the south and the ephemeral snow line. Maximum snowfall ranges from 0.1 mm to 1600 mm over most of the continent, with maximum values of 3000 in a few grid cells in southeastern Alaska. The Minimum November snowfall map shows a lack of snowfall over most of the United States, but still as much as 160 mm in parts of Canada.

December snowfall is examined in figure 3.12. Mean monthly values are the highest in the Rockies, Cascades, Southeastern Alaska, Great Lakes region, and Maritime Provinces. Values are as high as 950 mm to 1000 mm. The coefficient of variation is similar to the other winter maps where most of the continent has values between 0 and 1 for most of the continent, and higher values appearing along the ephemeral snow line in the southern United States. Maximum values for December are as high as 2400 mm in Southeastern Alaska. Other areas of high values occur in the inter-mountain west, Great Lakes regions, and the eastern United States/Maritime Provinces of Canada. Minimum December values range from 0 to 300 mm, with the highest minimum values occurring in the Quebec/Ontario Great Lakes region and the Canadian Rockies.

In these monthly maps the annual transition is clear. In the summer months there are few grid cells with mean snowfall values above zero. In October, snowfall starts to increase and more grid cells have an average above zero. Snowfall continues to increase into the winter months where December (closely followed by January) has the largest mean snowfall values. Values continue to stay large
throughout winter and into March and April. While some areas experience peaks in snowfall in the spring, generally snowfall decreases into the summer months.
Figure 3.1  January mean snowfall (A), coefficient of variation (B), maximum snowfall amount (C), and minimum snowfall amount (D) over the time period 1949 to 1999.
Figure 3.2  February mean snowfall (A), coefficient of variation (B), maximum snowfall amount (C), and minimum snowfall amount (D) over the time period 1949 to 1999.
Figure 3.3  March mean snowfall (A), coefficient of variation (B), maximum snowfall amount (C), and minimum snowfall amount (D) over the time period 1949 to 1999.
Figure 3.4  April mean snowfall (A), coefficient of variation (B), maximum snowfall amount (C), and minimum snowfall amount (D) over the time period 1949 to 1999.
Figure 3.5  May mean snowfall (A), coefficient of variation (B), maximum snowfall amount (C), and minimum snowfall amount (D) over the time period 1949 to 1999.
Figure 3.6  June mean snowfall (A), coefficient of variation (B), maximum snowfall amount (C), and minimum snowfall amount (D) over the time period 1949 to 1999.
Figure 3.7  July mean snowfall (A), coefficient of variation (B), maximum snowfall amount (C), and minimum snowfall amount (D) over the time period 1949 to 1999.
Figure 3.8  August mean snowfall (A), coefficient of variation (B), maximum snowfall amount (C), and minimum snowfall amount (D) over the time period 1949 to 1999.
Figure 3.9  September mean snowfall (A), coefficient of variation (B), maximum snowfall amount (C), and minimum snowfall amount (D) over the time period 1949 to 1999.
Figure 3.10  October mean snowfall (A), coefficient of variation (B), maximum snowfall amount (C), and minimum snowfall amount (D) over the time period 1949 to 1999.
Figure 3.11 November mean snowfall (A), coefficient of variation (B), maximum snowfall amount (C), and minimum snowfall amount (D) over the time period 1949 to 1999.
Figure 3.12  December mean snowfall (A), coefficient of variation (B), maximum snowfall amount (C), and minimum snowfall amount (D) over the time period 1949 to 1999.
3.2 **Seasonal Climatology and Regional Maps**

Mean seasonal snowfall (July 1 to June 30) for 1949 through 1999 is shown in Figure 3.13. This is the average of all the total seasonal snowfall values for each grid cell. The highest values occur in high elevation mountainous regions of the Rockies and Cascade/Sierra Mountains, Southeastern Alaska, the Great Lakes region and the Maritime Provinces. Values are has high as 4690 mm in these regions.

The mean seasonal map is consistent with the monthly maps already discussed in this chapter. The high elevation regions (including southeastern Alaska) have the highest snowfall values in almost all months. The Great Lakes and Maritime Provinces have the highest grid cell totals in the fall and spring months.

The seasonal coefficient of variation map in Figure 3.14 is also consistent with the monthly maps previously mentioned. Small values over most of the continent show small variations in snowfall away from the grid cell’s mean in all areas except the southern United States. This is a region where snowfall is not consistent and varies greatly from year to year.

Figure 3.15 details the maximum seasonal snowfall values for each grid cell over the 51 year period from 1949 to 1999. The grid cell with the largest seasonal snowfall is located in southeastern Alaska with a value greater than 14,400 mm. Other regions that consistently have the largest maximum values in the monthly maps also have the highest maximum seasonal values. These places include southeastern Alaska, the inter-mountain west, the Great Lakes region, and the Maritime Provinces.

The geographical regions that are predominant in the maximum seasonal snowfall map are as notable in Figure 3.16 detailing the minimum seasonal snowfall.
The highest values occur in the Rocky Mountains and in the Quebec and Labrador Provinces. Along the southern tier and west coast of the United States there are many grid cells with as little as zero snowfall during the minimum season.

The high resolution of this data set also makes it possible to look at specific regions in more detail. For purposes of data verification, a few select regions are examined further to verify that our climatology matches known patterns of snowfall based on elevation, latitude, and continentality.

The Pacific Northwest is examined in figure 3.17. Lower snowfall values can be seen in the coastal areas and increases in the high elevation areas in the Cascade/Sierra Mountains, and the Rocky Mountains. Higher snowfall values also highlight the various smaller mountain ranges in the Yukon, to the North of the Rocky Mountains, like the Mackenzie and Ogilvie Mountain ranges. The highest seasonal snowfall values that occur in southeastern Alaska can also be seen to exist in the high elevation mountains around Yukutat Bay.

Figure 3.18 shows the Northeast mean seasonal snowfall for 1949 to 1999. In this figure, regions that can be picked out based on higher seasonal snowfall values are the lake-effect areas around Lakes Erie, Ontario, Huron, Superior, and near the Ungava Bay. In North Carolina, Virginia, West Virginia, western Maryland, Pennsylvania, and into New York the Appalachian Mountains appear with higher snowfall values. This clearly shows the ability of the 1° x 1° resolution of this data set to make smaller scale features identifiable.
Figure 3.13  Mean seasonal snowfall over the time period 1949 to 1999.
Figure 3.14  Seasonal coefficient of variation over the time period 1949 to 1999.
Figure 3.15  Seasonal maximum snowfall over the time period 1949 to 1999.
Figure 3.16  Seasonal minimum snowfall over the time period 1949 to 1999.
Figure 3.17  Pacific Northwest mean snowfall over the time period 1949 to 1999.
Figure 3.18  Northeast mean snowfall over the time period 1949 to 1999.
3.2 Summary and Analysis

The annual cycle of North American snowfall is well documented by this data set. With the exception of northern Canada, throughout the summer months there are few grid cells with any snowfall. In autumn (September, October, and November) consistent snowfall moves south from Canada, first in the area of the Rocky Mountains, then throughout the western United States, into the northeastern United States and the Great Lakes region. During this time there are also several grid cells in Canada and Alaska that experience minimum snowfall values above zero, indicating that even in the most unfavorable conditions these regions receive some snowfall greater than zero during the time period. In the winter months, high mean snowfall values are seen in southern Alaska, the inter-mountain west, eastern Canada and the Great Lakes region. As the months move into spring the mean snowfall values start to decrease. The ephemeral snow line moves north as snowfall becomes more sporadic with the last areas to stop receiving a mean snowfall above zero being the Rocky Mountains and eastern Canada.

Three areas in particular stand out as large snowfall accumulation regions across North America. First, mean snowfall values are among the largest in high elevation regions, such as the Rocky, Cascades, and Sierra Nevada ranges. Considering maximum monthly values, October through April have the highest snowfall values at high elevation locations. The second region with large snowfall values is the Maritime Provinces of Canada. Consistently widespread and large mean monthly values are found in November through March. Thirdly, the Great Lakes region shows evidence of lake effect snow enhancement in the winter and spring. The
highest mean monthly values occur in the Great lakes region in the winter and spring months of November through March.

This chapter identifies the ability of the data set to reasonably capture the climatology of North American snowfall when properly constrained. The monthly and seasonal statistics are consistent with what is expected as far as spatial and temporal behavior of snowfall. The valuable property of this data set’s fine scale resolution is also highlighted, by exploring the ease of identifying smaller scale features in the regional maps.
4.1 Calculation of trends

In order to identify potential trends in snowfall variables, least squares linear regressions are calculated between each variable and time. The slope of the linear regression line identifies temporal changes in the dependent variable. For the majority of analyses in this chapter, the slope, or trend, is expressed as a value over the 51-year period, 1949 to 1999. Positive values indicate an increase in the dependent variable over the time period, and negative values indicate a decrease.

Correlation coefficients are calculated for the linear regressions but are not mapped. This study seeks to identify homogeneous areas of physically significant changes in snowfall over time. Because correlation coefficients may highlight statistically significant grid cells with physically insignificant trends they are not shown.

4.2 Seasonal Trends over 1949-1999

Changes in several snowfall variables with time are examined in this chapter. As seen in the data description (see Chapter 2) several grids have data over the entire period of record allowing trends to be calculated back to 1900. However, the aim of this study is to examine the entire continent for synoptic or larger scale patterns in snowfall change. As shown in Chapter 2, the best coverage of reliable grid
cells over North America does not occur until 1949. For this reason, only the last 51 years are used in the subsequent trend analysis.

The changes in seasonally calculated variables are shown first. These variables are total seasonal snowfall, number of seasonal snowfall events, date of the last seasonal snowfall, date of the first seasonal snowfall, and length of the snowfall season. In all figures, grey grid cells have no reporting stations during the entire record. Also, blacked out grids have 10 or more years with differences in reporting stations greater than 10% during the period of record 1949 to 1999.

The slope of the linear regression line for total seasonal snowfall is shown in figure 4.1. In the Pacific Northwest general decreases in total seasonal snowfall are seen, ranging from -0.5 mm to -30 mm over the half-century. Many of the largest decreases occur in spatially small regions, at high elevations. Several areas experience increases in snowfall. Increases are seen in Alaska, the Great Plains, the Great Lakes, Northeast United States, and most grid cells in the northern half of Canada. The increases in total seasonal snowfall range from +0.5 mm to +30 mm over the period.

The trend of the number of seasonal snowfall events is examined in figure 4.2. Each day that measurable snowfall is recorded is considered a snowfall event. Few grid cells with any appreciable trends are present for the period of record.

Figure 4.3 represents the trends of the date of the first seasonal snowfall. The south central United States, the Rocky Mountains, and a small area of grid cells in central Manitoba have experienced a later date of first snowfall, by as much as 4 days over the half-century. The South central United States has an area of earlier first snow dates directly north of a swath of later first snow dates. These are also trends of as
much as 3 to 4 days. However, this is a region where snowfall is infrequent, and these trends may only indicate the change in a few rare storm events.

Changes in the date of last seasonal snowfall are depicted in figure 4.4. The southeastern United States (along the ephemeral snow line) has experienced a later date of last seasonal snowfall by .5 to 3 days over the half-century. There are a few grid cells in the mountains of southern California with an earlier date of last seasonal snowfall, also along the ephemeral snow line.

Subtracting the date of first snowfall from the date of last snowfall gives a length of snowfall season, shown in figure 4.5. There is a swath of longer snowfall season across the southeastern ephemeral snow line in the United States. This reflects the earlier arrival of the first day of snowfall in this region. There is a shorter snow season in the mountains of Southern California and in parts of the Rocky Mountains of as much as 5 days. This is caused by a later first snowfall date in the Rockies and an earlier end to the snow season in southern California.
Figure 4.1  Total seasonal snowfall, slope of the linear regression for 1949 to 1999.
Figure 4.2  Number of seasonal snowfall events, slope of the linear regression for 1949 to 1999.
Figure 4.3  Date of first seasonal snowfall, slope of the linear regression for 1949 to 1999.
Figure 4.4  Date of last seasonal snowfall, slope of the linear regression for 1949 to 1999.
Figure 4.5  Length of snowfall season, slope of the linear regression for 1949 to 1999.
4.3 Monthly Snowfall Trends over 1949-2000

January snowfall slope of the linear regression is displayed in figure 4.6. In January, the majority of reliable grids in the eastern half of the continent show an increase in total January snowfall of as much as 15 mm over the approximate half-century. The highest values occur in the Great Lakes region and along Newfoundland and Labrador of the Maritime Provinces. Decreases in snowfall occur over much of the western United States into the southern portions of British Columbia and Alberta provinces. These decreases are as much as -20 mm over the approximate half-century.

February’s snowfall trends are examined in Figure 4.7. In this month there are few regions with trends greater than +/-5 mm/half-century.

Figure 4.8 details March snowfall trends. In the Pacific Northwest the high elevation regions show a decrease in total March snowfall, from -4 to -14 mm/half-century. Eastern Canada shows slight increases in snowfall, with the maximum reaching 6 mm/half-century.

April snowfall trends are shown in figure 4.9. The highest increases in snowfall occur in the New England/Eastern Great Lakes region north, into Quebec. There are also increases in the central United States. None of these increases exceed 4 mm/half-century. Decreases occur in the western great lakes region, and in the western United States. These decreases are as much as -8 mm/half-century around Montana and parts of the Sierra Nevada Mountains.

Figure 4.10 displays the slope of the linear regression of May snowfall. There are small decreases in snowfall in the intermountain west of the United States, Upper Great Lakes region, and Southeastern Canada. The largest decreases are 4 mm
over the half-century. The majority of grid cells in Northern Canada have experienced increases in snowfall of as much as 4 mm/half-century.

June, July, and August have few or no grid cells with snowfall data so these months are not analyzed in this section. Figure 4.11 details September snowfall slope of the linear regression. Northern grid cells have frequent snow during this month, however only small trends are apparent. A swath of snowfall decreases exists in the Rocky Mountains near the United States-Canada border. The largest values are around –2 mm/half-century. Several of the northern grid cells have a positive trend, centralized in southeastern Alaska. These values are has high as 3 mm/half-century.

Trends become larger as the snow season moves into October. Figure 4.12 shows these trends. Several of the northern grid cells in Canada and Alaska have positive trends of as much as 6 mm/half-century. In addition, snowfall increases of up to 2 mm/half-century are found across the High Plains of the central United States.

November Snowfall trends are examined in figure 4.13. Negative trends are widespread in the Upper Great Lakes region of –5 to –10 mm/half-century. The largest positive trends occur in southeastern Alaska with values as high as 15 mm/half-century. Smaller positive trends are apparent over most of the northern Great Plains of the United States into southern Canada.

December trends are depicted in figure 4.14. Increases in snowfall occur in southern Alaska, the Great Lakes region, the High Plains, Newfoundland and Labrador. These increases are as much as 20 mm over the half-century. Decreases are mainly found in the Pacific Northwest. The largest decreases are 15 mm/half-century.
Figure 4.6  Total January Snowfall, slope of the linear regression for 1949 to 2000.
Figure 4.7  Total February Snowfall, slope of the linear regression for 1949 to 2000.
Figure 4.8  Total March Snowfall, slope of the linear regression for 1949 to 2000.
Figure 4.9  Total April snowfall, slope of the linear regression for 1949 to 2000.
Figure 4.10  Total May snowfall, slope of the linear regression for 1949 to 2000.
Figure 4.11  Total September snowfall, slope of the linear regression for 1949 to 2000.
Figure 4.12  Total October snowfall, slope of the linear regression for 1949 to 2000.
Figure 4.13  Total November snowfall, slope of the linear regression for 1949 to 2000.
Figure 4.14 Total December snowfall, slope of the linear regression for 1949 to 2000.
4.4 Summary and Analysis

This chapter examines trends in snowfall data over the last half-century as variables calculated per snow season and by month. Large seasonal trends include: decreases of up to -30 mm of snowfall per half-century in the Pacific Northwest; an increase in the total seasonal snowfall in Alaska, the Great Plains, the Great Lakes, Northeast United States, and reliable grid cells in northern Canada of as much as 30 mm over the period; and a shorter snow season in Southern California and parts of the Rocky mountains, reduced at both ends of the annual cycle. Monthly trend maps identify the largest changes in monthly snowfall as occurring in the winter and spring months. These decreases in the western United States snowfall are primarily located in high elevation regions.

The Pacific Northwest was identified by Scott and Kaiser (2004) as an area of snowfall change. They found decreases in total and extreme snowfall events, leading to declines in overall snowfall of the Pacific Northwest. Snow Water Equivalent studies similarly found decreases in April 1st SWE (Mote et al., 2004). Similarly, this study identifies decreased snowfall in the western United States primarily due to less snowfall in December through March. These results further justify the reliability of this data set over the chosen period of study.

Another region given significant attention by the previous literature is the Great Lakes region. The current study found increases in the Great Lakes region snowfall, as a result of significant increases in December, January and April snowfall totals. This finding is corroborated by Leathers and Ellis (1996), Smith et al. (2001), Burnett et al. (2003), and Scott and Kaiser (2004).
While the status of changes in snowfall over the last half-century in North America is addressed in this chapter, the reason for these trends is still unknown. A few suggestions for future research are presented, with the hope of establishing causal relationships. In the first step, the snowfall time series should be examined to identify the percentage of snowfall variation that can be attributed to teleconnection patterns, which is the focus of Chapter 5. Secondly, work is needed to separate the effects of elevation from other forcing mechanisms of snowfall trends. Finally, it is very likely that half of a century is not enough time to identify long-term trends in snowfall. Therefore, work is needed to establish a means of accurately monitoring long-term trends for future study.
5.1 Teleconnection Data

Arctic Oscillation (AO) data were obtained from the Climate Prediction Center’s Monitoring and Data Index page (Climate Prediction Center Internet Team, 2004). These data are composed of monthly index values for the years 1950 to 2004. AO values were calculated by projecting daily and monthly mean 1000 mb height anomalies onto the leading EOF of the monthly mean height anomalies at 1000 mb. The time series was then normalized by the standard deviation of the monthly index.

The North Atlantic Oscillation (NAO) index values are also from the Climate Prediction Center’s Monitoring and Data Index page. The monthly and annual values extend from 1821 to 2000. The index values used in this study were calculated based on a Rotated Principal Component Analysis (RPCA) (Barnston and Livezey, 1987) applied to monthly standardized 500 mb height anomalies in the region from 20° N to 90° N between January 1950 and December 2000. The leading principal components refer to teleconnection patterns and the solution to a least squares regression of the system of equations yields the teleconnection indices. Similarly, Pacific North American (PNA) index values were calculated along with the NAO index in the RPCA. These monthly values extend from 1950 to 2004.

Pacific Decadal Oscillation (PDO) values were obtained from the University of Washington website (Mantua, 2000) PDO index. The PDO is defined as
a leading mode of multi-decadal variability in sea surface temperatures (SSTs) in the extratropical North Pacific. It was computed as the time series scores associated with the leading principal component of SSTs in the Pacific, poleward of 20°N. This monthly index is available for the years 1900 to 2000.

Southern Oscillation Index (SOI) data were obtained from the University of East Anglia Climate Research Unit’s Data site (Salmon, 2006). To derive SOI indices, pressure differences between Tahiti and Darwin were normalized as described in Ropelewski and Jones (1987). This monthly, annual and seasonal data extends from 1866 to 2003.

For individual analyses with these data sets, 1949/1950 to 2000 values are used in order to overlap with the years of reliable snowfall data. When all teleconnection data are included in the analyses, the time period 1950 to 2000 is used, limited by the start date of both the AO and PNA.

5.2 Monthly linear correlations

In order to identify basic relationships between snowfall and teleconnection patterns, simple linear regressions as well as Pearson correlation coefficients are calculated. These calculations are done monthly for each of the 5 teleconnection patterns and snowfall. Only maps with strong, visible signals are shown.

The correlation coefficients between the Arctic Oscillation and monthly snowfall are displayed in figure 5.1. In the fall, negative correlations are seen in the southeastern United States starting in November, moving north and expanding into much of the Ohio River Valley and Central Plains in December. These correlation coefficients become stronger in January with values as low as -0.6 to -0.7 in
Kentucky, Tennessee, western Virginia and West Virginia. These negative correlations persist into February and March where they remain strong, but move west, toward the Central Plains (especially in March). In general, from late fall into early spring there are negative correlations between snowfall and the AO over much of eastern North America, which moves towards the west in March. The $R^2$ values for these strongly correlated regions are as high as .49, implying that the AO explains close to 50 percent of the variance in snowfall at these locations.

The correlation coefficients between the North Atlantic Oscillation and monthly snowfall are shown in figure 5.2. Only two months exhibit homogeneous regions of appreciable correlations. September has negative correlations centered in northwestern Minnesota and North Dakota, with weaker negative correlations in much of the reliable grid cells in the northern part of the continent. In October the negative correlations are located in the Great Lakes region along with a swath extending into the south and south central United States. There are some regions of small positive correlations in the inter-mountain west in September and a broader area in the western half of the continent in October.

In figure 5.3 the relationship between the Pacific Decadal Oscillation and monthly snowfall is detailed. Several months are included in this figure because most of the fall, winter, and spring months have persistent and clear areas of large correlations. Beginning in the fall, October displays some smaller negative correlations over much of the western half of the continent and small positive correlations in eastern North America (0 to 0.4). During the months of November, December and January the positive values decreases spatially and move to the south central United States, while the negative values intensify and expand. There are large
areas of the Pacific Northwest, Alaska, and southern Canada with correlation coefficients as low as -0.7. In February and March the areas of negative correlations decrease to mainly the Pacific Northwest and a small area in the Great Lakes region. In March and April the areas of positive correlations strengthen to as much as 0.4 to 0.5 and move into much of the central plains as well as the southeastern United States.

The Pacific North American Index correlations with monthly snowfall are displayed in figure 5.4. Much like the PDO, the general signal is of negative correlations in the Pacific Northwest and positive correlations in the east/southeastern United States. In the fall, negative correlations appear in the Great Plains, Four Corners regions of the United States, and the southern Canadian Prairies. Negative correlations migrate toward the western half of the continent as the season progresses into winter. By January the largest negative correlations are found along the west coast of North America. Negative correlations decrease into the spring. In most months the negative correlations are not as strong as the PDO, with the exception of January. Positive correlations also develop in the fall, primarily located in the Great Lakes region, the mid-Atlantic, and the southern tier of the United States. These positive values are highest in November and December where the maximums in the mid-Atlantic region are stronger than the PDO correlations.

Figure 5.5 shows the correlations between the Southern Oscillation Index and monthly snowfall. In all four of the months shown the most notable features are the negative correlation coefficients of as much as -0.5 in the central United States. These negative correlations are accompanied by small areas of less persistent positive correlations which are located in the Western United States in January and more
noticeably in October. The March correlation coefficients also point to a small area of positive correlations around Missouri and Arkansas.
Figure 5.1  Arctic Oscillation and monthly snowfall correlations.
Figure 5.2 North American Oscillation and monthly snowfall correlations.
Figure 5.3  Pacific Decadal Oscillation and monthly snowfall correlations.
Figure 5.4  Pacific North American and monthly snowfall correlations.
Figure 5.5   Southern Oscillation Index and monthly snowfall correlations.
5.2 Multiple linear stepwise regressions with monthly snowfall

For each month a multiple linear stepwise regression was calculated between monthly snowfall (dependent variable) and the AO, NAO, PDO, PNA, and SOI (independent variables). The time period used for this analysis is 1950 to 1999 due to availability of teleconnection data. The multiple linear stepwise regression maps are compared to the individual monthly correlation maps from the simple linear regression. By examining both analyses together, teleconnection patterns responsible for the total explained variance can be identified.

A multiple linear regression model was built with stepwise selection of independent variables. This is done using the variance-covariance matrix of the monthly snowfall and monthly teleconnection data. This method of analysis is useful in this situation because it allows the observational data (snowfall) to be characterized by several different variables (teleconnection data).

In figure 5.6 the amount of variation in snowfall explained by a linear combination of the independent variables ($R^2$) is shown for each grid cell. The highest $R^2$ values occur in the Pacific Northwest and are as high as 0.7. This explained variation is due primarily to the PDO and PNA, which are highly associated with snowfall. There are also some relatively large $R^2$ values in the mid Atlantic region in Virginia and North Carolina, which can be attributed to small negative correlations with the AO and positive correlations with the PNA index in that region.

Figure 5.7 shows the same analysis for February. There are a few grid cells in the Pacific Northwest that have $R^2$ values of 0.4 to 0.5, a few grid cells in the great lakes region with 0.2 to 0.3, and several grid cells in the western United States
that have 0.1 to 0.2. These may be due to small positive correlations with the AO in the west, larger negative correlations with the PDO and the PNA index in the Pacific Northwest and small negative correlations with the SOI in the southern Sierra Nevada Mountains. Overall, variation in snowfall attributed to teleconnections is greatly reduced compared to January.

The March analysis is shown in figure 5.7. During this month there are large $R^2$ values in the Pacific Northwest, as well as higher values moving into the southern Great Plains with a maximum of 0.4 to 0.5 in Kansas. In this month Alaska also has increases in the number of grid cells with values as high as 0.4. The AO and SOI explain some of the variance in the central Great Plains, while the PDO and PNA are responsible for most of the variance explained in the Pacific Northwest.

Figure 5.8 displays the April multilinear stepwise regression. The majority of the grid cells with $R^2$ values higher than 0.2 are along the boarder between the United States and Canada, which for this month is the ephemeral snow region. These values are attributed to the PNA which has negative correlations with snowfall in that region. The number of grid cells in the continental United States with snowfall/teleconnection associations starts to decrease in this month as fewer grid cells have appreciable snow at this time of year.

For the months of May, June, July, and August too little snow falls across most of North America for any meaningful analysis.

The September multilinear stepwise regression is shown in figure 5.9. For this month, there are only a few grid cells with $R^2$ values greater than 0.2 to 0.3. These occur mostly in scattered locations in the western United States, extending
north into Alberta and Saskatchewan. The PNA has the largest negative correlation in
this region and is primarily responsible for the larger $R^2$ values here.

The October stepwise regression in figure 5.10 is similar to September
with few grid cells with $R^2$ values greater than 0.2 to 0.3. However, most of the
higher valued grid cells occur in the Great Lakes region in southeastern Ontario and
southwestern Quebec. The $R^2$ values in this area are as high as 0.4 to 0.5.
Teleconnection patterns with correlations in this area are the NAO, PDO, PNA, and
SOI. The negative NAO correlation coefficient and positive correlation coefficient
with the PNA make up a large part of the variance explained.

In the November map, shown in figure 5.11, higher values start to appear.
In the Pacific Northwest there is a local maximum in variance explained by the
teleconnection patterns. Just north of Vancouver Island in British Columbia there are
several grid cells with $R^2$ values as high as 0.5 to 0.6, surrounded by an area of
elevated values at 0.3 to 0.5 extending west into Alberta, and south into Washington
and Oregon. This is mainly due to the large negative correlations with the PDO in this
region, and with some small correlations with the PNA.
Figure 5.12 shows the December $R^2$ values from the regression. In this month, the Pacific Northwest grid cells with high November values have intensified, and increased. There are many grid cells with $R^2$ values as high as 0.7 along the Alberta and Washington coast, extending into the higher elevations of Oregon, and to the west in Idaho, and western Montana. This is attributed to the large negative correlations with the PDO in this region. There are also a noticeable number of grid cells in the Great Lakes region with $R^2$ values of as high as 0.3 to 0.4, which may be related to negative correlations with the AO and positive correlations with the PNA index.
Figure 5.5  Multiple linear stepwise regression with Monthly snowfall, PNA, SOI, AO, NAO, and PDO for January 1950 to 1999.
Figure 5.6  Multiple linear stepwise regression with Monthly snowfall, PNA, SOI, AO, NAO, and PDO for February 1950 to 1999.
Figure 5.7  Multiple linear stepwise regression with Monthly snowfall, PNA, SOI, AO, NAO, and PDO for March 1950 to 1999.
Figure 5.8  Multiple linear stepwise regression with Monthly snowfall, PNA, SOI, AO, NAO, and PDO for April 1950 to 1999.
Figure 5.9  Multiple linear stepwise regression with Monthly snowfall, PNA, SOI, AO, NAO, and PDO for September 1950 to 1999.
Figure 5.10  Multiple linear stepwise regression with Monthly snowfall, PNA, SOI, AO, NAO, and PDO for October 1950 to 1999.
Figure 5.11  Multiple linear stepwise regression with Monthly snowfall, PNA, SOI, AO, NAO, and PDO for November 1950 to 1999.
Figure 5.12  Multiple linear stepwise regression with Monthly snowfall, PNA, SOI, AO, NAO, and PDO for December 1950 to 1999.
5.3 Summary and Analysis

This chapter examines correlations between teleconnection patterns and North American snowfall. Several associations are spatially homogeneous and temporally robust. A majority of the findings were previously established in the literature, if only for a particular section of the study area.

The general negative correlation between the eastern half of the United States and the AO are consistent with the previous literature. During a positive AO phase, the reduction of cold air outbreaks east of the Rocky Mountains leads to higher temperatures. This could cause the precipitation to fall more frequently as rain rather than snow, decreasing the snowfall amounts in these regions during a positive AO. Furthermore, Serreze et al. (1998) also found that snowfall in the southern and northeast parts of the United States is heavily influenced by the maximum temperature on precipitation days.

Correlations between the NAO and monthly snowfall are most likely related to changes of the mid-tropospheric flow. Bradbury et al. (2002) found that the NAO is associated with an eastward displacement of the eastern trough during its negative phase, which could explain the negative correlations with the northern Great Plains in September, and the Great Lakes/Eastern United States in October.

Maps of PDO and snowfall correlations have striking areas of strong correlations that are fairly persistent throughout the snow season. The negative correlations in the Pacific Northwest are consistent with previous findings of decreases in winter precipitation in the Pacific Northwest and of a weakened Aleutian low during negative (cold) phases (Gedalof and Mantua, 2002). Correspondingly,
April 1st snowpack and SWE studies find associations with the PDO (McCabe and Dettinger, 2002; Hunter et al., 2006).

Monthly correlations with the PNA look similar to those of the PDO. The general negative correlations in the Pacific Northwest correspond to the more meridional characteristics of the Pacific North American pattern during the positive phase of the PNA index. In this phase the ridge over the western part of the continent is accentuated, resulting in changes in storm tracks and higher temperatures. The positive correlations in the eastern United States are associated with the deepening of the eastern trough during a positive PNA index, which results in colder temperatures and changes in storm tracks (Leathers et al., 1991; Serreze et al., 1998; Bradbury et al., 2002; Notaro et al., 2006).

Previous literature identifies associations with ENSO and the Pacific Northwest, Great Lakes, Northeastern United States, and to some extent parts of the Rocky Mountains (Smith and O’Brien, 2001; Patten et al., 2003; Hunter et al., 2006). Interestingly, this study found the largest correlations in the central United States, along with a small area of positive correlations in the West. This difference could be due to the use of different estimations for the El Nino Southern Oscillation, and should be examined in greater detail in subsequent studies.

Beyond the individual correlation analyses, the multilinear stepwise regressions show that these teleconnection patterns can account for up to 70 % of the variance in snowfall, principally in the Pacific Northwest. This study contributes to the body of knowledge by expanding the geographic extent over which these relationships are considered, and confirming previous research. While not all of the
mechanisms behind the correlations are explained, the identification of these associations could lead to enhanced forecasting abilities.
Chapter 6

SUMMARY AND CONCLUSIONS

Several studies have investigated snowfall across North America and have identified changing snowfall characteristics with time. Regions repeatedly identified in the literature with significant change are the Pacific Northwest, and the Great Lakes (Burnett et al., 2003; Leathers et al., 1993; Ellis and Leathers, 1996; Leathers and Ellis, 1996; Mote et al, 2004; Scott and Kaiser, 2004). However, the previous body of work is limited by the study of specific regions, relatively small data sets, and short periods of record.

The goal of this study was to examine snowfall across the entire North American continent, using a larger selection of data, over a longer time period and to better understand the reasons for snowfall variability over the last century. Major findings of this study are as follows:

- The Dyer and Mote (2006) gridded snowfall product for North America is a unique and usefull data set for climatological studies. It is an improvement over other data sets specifically with respect to its spatial coverage. However, a thorough understanding of the limitations of the data is necessary before using the gridded product.

- The Rocky Mountains, coastal Alaska, Great Lakes snowbelts and southeastern Canada have the largest snowfall accumulations on the continent. Smallest accumulations are found along the southern tier of the United States.

- Temporal trends in snowfall identified in this study agree with previous work. Large decreases in snowfall are seen in the
Pacific Northwest, but to a spatial extent that is previously undocumented, and increases in snowfall are seen in several areas across the continent.

- Atmospheric teleconnections account for a substantial amount of variation in snowfall, especially the AO, PDO, and PNA.

The data verification identifies a number of potential problems that would undermine the use of the gridded snowfall product without an in-depth study of its character and the adoption of certain constraints. Period of record is a major issue. As stations “come on-line” the grid cell means can change appreciably creating spurious trends. Independent of large scale increases/decreases in the number of reporting stations, there can be small regional changes in the number of stations within grid cells, which must be addressed on a grid by grid basis. It was also determined that several grid cells never include any reporting stations leading to interpolation of data from long distances. For these reasons, the data used in this study were limited in space and time, insuring its reliability for the application of this research.

A climatology of North American snowfall was conducted in order to validate the gridded snowfall data set and to provide a detailed climatology of snowfall across the continent. Values are consistent with the expected annual cycle of snowfall. High elevation regions, the Maritime Provinces, and the Great Lakes region are all locations of high snowfall accumulation. The southeast, Gulf Coast, and southwest regions of the United States have the smallest accumulation totals.

An analysis of trends in snowfall over the last half-century points to the Pacific Northwest as a large region of homogeneous decreases in snowfall. Decreasing values are found to extend east, encompass the Rocky Mountains, and north into British Columbia and southern Alaska. Areas of increasing snowfall include the Great Lakes, Alaska, the Great Plains, Northeast United States, and
reliable grid cells in northern Canada. The monthly analysis points to winter and early spring months as the primary time of snowfall change. Previous literature is consistent with some of these results (Leathers and Ellis, 1996; Smith et al., 2001; Burnett at al., 2003; Scott and Kaiser, 2004; Mote et al., 2004). However, previous studies do not identify all of the trends found in this research and are lacking the spatial detail of the data set used here.

A preliminary study of the relationship between major teleconnection patterns and North American snowfall identified persistent associations in the Pacific Northwest and Eastern United States. The strongest correlations are seen in the Pacific Northwest, where snowfall is inversely associated with the PDO and PNA. In the Eastern United States the AO and NAO also have a somewhat persistent negative relationship with snowfall. In most cases these reflect changes in large scale air flow that either change the temperature at which precipitation is occurring, or the amount of precipitation occurring due to storm tracks or available moisture. The multiple linear stepwise regression analysis further points to the Pacific Northwest, as an area where teleconnection patterns account for a substantial amount of the snowfall variation.

This study points to teleconnection patterns as a means of understanding and possibly forecasting North American snowfall. More detailed studies of regional teleconnection/snowfall relationships could enhance our understanding of the physical processes involved in the statistical relationships shown here. Extension of the trend analysis into the past would aid in the understanding of how snowfall has changed over time, as well as help to further quantify the relationships between snowfall and atmospheric teleconnections.
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