A SYNOPTIC CLIMATOLOGICAL ASSESSMENT OF THE RELATIONSHIP BETWEEN
ARCTIC SEA ICE VARIABILITY AND CLIMATE ANOMALIES OVER NORTH AMERICA

A dissertation submitted to Kent State University in partial fulfillment of the requirements
for the degree of Doctor of Philosophy

by

Thomas J. Ballinger

May 2015

© Copyright
All rights reserved

Except for previously published materials
Dissertation written by

Thomas J. Ballinger

B.A., Kent State University, 2008
M.A., The Ohio State University, 2011
Ph.D., Kent State University, 2015

Approved by

Scott C. Sheridan, Professor, Ph.D., Department of Geography, Doctoral Advisor

Thomas W. Schmidlin, Professor, Ph.D., Department of Geography

Mandy J. Munro-Stasiuk, Professor, Ph.D., Department of Geography

Joseph D. Ortiz, Professor, Ph.D., Department of Geology

Daniel K. Holm, Professor, Ph.D., Department of Geology

Accepted by

Mandy J. Munro-Stasiuk, Professor, Ph.D., Chair, Department of Geography

James L. Blank, Professor, Ph.D., Dean, College of Arts and Sciences
TABLE OF CONTENTS

TABLE OF CONTENTS.............................................................................................................iii

LIST OF FIGURES..................................................................................................................vii

LIST OF TABLES....................................................................................................................xi

PREFACE...............................................................................................................................xii

ACKNOWLEDGMENTS..........................................................................................................xiii

CHAPTER 1: INTRODUCTION AND BACKGROUND

1.1 Introduction.....................................................................................................................1

1.2 Background.....................................................................................................................3

1.2.1 Arctic Sea Ice Behaviors............................................................................................3

1.2.2 Causes of Arctic Sea Ice Variability and Change.......................................................6

1.2.3 Impacts and Implications of Arctic Sea Ice Variability and Change.....................9

1.2.4 Datasets and Methods Often Employed in Sea Ice-Climate Studies....................12

1.3 Motivation for Research, Layout, and Summary of Dissertation Objectives........16

References............................................................................................................................19

CHAPTER 2: ASSOCIATIONS BETWEEN CIRCULATION PATTERN FREQUENCIES AND SEA ICE MINIMA IN THE WESTERN ARCTIC

Abstract...............................................................................................................................29

2.1 Introduction.....................................................................................................................30

2.2 Data and Methodology..................................................................................................34

2.2.1 Data..........................................................................................................................34

2.2.2 Methodology............................................................................................................36

2.3 Results............................................................................................................................37
CHAPTER 3: REGIONAL ATMOSPHERIC PATTERNS AND THE DELAYED SEA-ICE FREEZE-UP IN THE WESTERN ARCTIC

Abstract.................................................................................................................62

3.1 Introduction........................................................................................................63

3.2 Data......................................................................................................................66

3.2.1 Reanalysis Fields...............................................................................................66

3.2.2 Sea Ice Data......................................................................................................67

3.3 Methodology..........................................................................................................68

3.3.1 General Overview of the Two-Step Cluster Analysis........................................68

3.3.2 Construction of the MSLP AP Classification.....................................................69

3.3.3 Construction of the 1000-500 hPa Thickness Classification...............................70

3.4 Results....................................................................................................................71

3.4.1 Frequency Analysis.............................................................................................71

3.4.1.1 MSLP Classification......................................................................................71

3.4.1.2 1000-500 hPa Thickness Classification.......................................................72

3.4.2 Assessing AP Temporal Differences through Compositing...............................73

3.4.2.1 MSLP.............................................................................................................73
CHAPTER 4: SEA ICE IMPACTS ON POLAR WEATHER TYPES IN THE NORTH AMERICAN ARCTIC

Abstract .................................................................................................................. 105

4.1 Introduction ......................................................................................................... 106

4.2 Data and Methods ............................................................................................. 109

4.2.1 Freeze-up Data ............................................................................................... 109

4.2.2 The Spatial Synoptic Classification ................................................................. 110

4.2.3 Analysis of SSC and Freeze-up Datasets ......................................................... 111

4.3 Results .................................................................................................................. 112

4.3.1 MP/DP Climatology and Trends ..................................................................... 112

4.3.2 Long-term MP/DP Associations with Freeze-up ........................................... 114

4.3.3 Late versus Early Freeze Year Composites .................................................... 115
4.3.4 Highlighting Individual Extreme Freeze Years.......................................................117
4.4 Discussion..................................................................................................................118
4.5 Conclusions................................................................................................................121
Acknowledgments..........................................................................................................123
References.......................................................................................................................124

CHAPTER 5: DISCUSSION AND CONCLUSIONS

5.1 Synthesis of Research Findings..................................................................................152
5.2 Discussion of Results..................................................................................................154
5.3 Potential Improvements and Research Directions....................................................158
References.......................................................................................................................160
LIST OF FIGURES

Figure 1.1. Map of the western Arctic (WA) domain. The Beaufort Sea (BS) and Chukchi Sea (CS) comprising the western Arctic Ocean are outlined.................................28

Figure 2.1. Map of the western Arctic domain. The 11 longitudes from which the latitude of the mid-September sea ice retreat is measured are indicated..............................55

Figure 2.2. The 15 SLP circulation patterns generated from NCEP/NCAR reanalysis........56

Figure 2.3. Observed sea ice retreat latitudes (solid) and predicted values (dashed) based on the SMLR equation for (a) L176, (b) L151, and (c) L126. Total explained sea ice extent variance (\(r^2\)) by the melt season synoptic patterns is indicated.................................57

Figure 2.4. Pie charts indicating the frequencies of circulation patterns 11-13 over June, July, and August for 1979-2006, 2007, and 2008-2011. The number indicates the average occurrence for 1979-2006, and 2008-2011, while the 2007 number indicates the occurrence for that year.................................................................58

Figure 3.1. Freeze-up dates for the Beaufort Sea, Chukchi Sea, and Western Arctic Ocean for the 1979-2013 period..................................................................................................................91

Figure 3.2. Same as Figure 3.1, but for melt onset dates.....................................................92

Figure 3.3. Full catalogue of the 15 MSLP patterns.................................................................93

Figure 3.4. Warm season-dominant MSLP patterns P11-14. See Figure 3.3 for the full catalogue.................................................................................................................................94

Figure 3.5. Full catalogue of the 15 1000-500 hPa thickness patterns......................................95

Figure 3.6. Warm season-dominant 1000-500 hPa thickness patterns T6, T9, and T13-15. See Figure 3.5 for the full catalogue.....................................................................................................................95

Figure 3.7. Quintiles of P11-14 frequencies by warm season months, May-October. Each quintile represents a consecutive 7-year period across the entire time series (1979-2013).....................................................................................................................96

Figure 3.8. Time series of P11 across the warm season months for the Early, Middle, and Late Periods described in Figure 3.7. The frequency represents the total number of days classified for each day (May-October) within each period. A 14-day centered moving average is then applied to smooth the time series across all months.................97

Figure 3.9. Quintiles of T6, T9, and T13-15 frequencies by warm season months, May-October. Each quintile represents a consecutive 7-year period across the entire time series (1979-2013).................................................................................................................................99
Figure 3.10. Time series of T9 across the warm season months for the Early, Middle, and Late Periods described in Figure 3.9. The frequency represents the total number of days classified for each day (May-October) within each period. A 14-day centered moving average is then applied to smooth the time series across all months.

Figure 4.1. Time series of western Arctic freeze-up date anomalies, 1979-2013. The 7 most extreme years, with respect to ±1 standard deviation (sigma) from the 1981-2010 freeze-up mean (Day of Year = 276) are utilized in composite and individual freeze year analyses described in Sections 4.3.3 and 4.3.4.

Figure 4.2. Map of 27 first-order weather stations poleward of 60°N with Spatial Synoptic Classification (SSC) weather types in North America (including Greenland). Table 4.1 provides details on each location. The green polygon outlines the approximate western Arctic sea ice freeze-up domain used in the analysis.

Figure 4.3. Monthly mean frequencies of MP classified days for October-March following data records listed in Table 4.1.

Figure 4.4. Monthly mean frequencies of DP classified days for October-March following data records listed in Table 4.1.

Figure 4.5. Linear trend (slope) of MP classified days year$^{-1}$ for October-March following data records listed in Table 4.1.

Figure 4.6. Linear trend (slope) of DP classified days year$^{-1}$ for October-March following data records listed in Table 4.1.

Figure 4.7. Pearson bivariate correlations between November MP weather type occurrences and the Beaufort Sea, Chukchi Sea, and western Arctic freeze-up dates.

Figure 4.8. Pearson bivariate correlations between October-March MP weather type occurrences and the freeze date of western Arctic sea ice. Correlation coefficients of -0.34 > r > +0.34 are statistically significant.

Figure 4.9. Pearson bivariate correlations between October-March DP weather type occurrences and the freeze date of western Arctic sea ice. Correlation coefficients of -0.34 > r > +0.34 are statistically significant.

Figure 4.10. Monthly differences in MP classified days observed for the 7 latest minus 7 earliest freeze-up years. The difference between the average of these two 7-year periods is portrayed in each plot.

Figure 4.11. Monthly differences in DP classified days observed for the 7 latest minus 7 earliest freeze-up years. The difference between the average of these two 7-year periods is portrayed in each plot.
Figure 4.12. MP October frequency anomalies versus the 1981-2010 MP October mean during the extreme early (E) freeze years (1983, 1985, and 1994) and late (L) freeze years (2007, 2011, and 2012)........................................................................................................139

Figure 4.13. DP October frequency anomalies versus the 1981-2010 DP October mean during the extreme early (E) freeze years (1983, 1985, and 1994) and late (L) freeze years (2007, 2011, and 2012)........................................................................................................140

Figure 4.14. MP November frequency anomalies versus the 1981-2010 MP November mean during the extreme early (E) freeze years (1983, 1985, and 1994) and late (L) freeze years (2007, 2011, and 2012)........................................................................................................141

Figure 4.15. DP November frequency anomalies versus the 1981-2010 DP November mean during the extreme early (E) freeze years (1983, 1985, and 1994) and late (L) freeze years (2007, 2011, and 2012)........................................................................................................142

Figure 4.16. MP December frequency anomalies versus the 1981-2010 MP December mean during the extreme early (E) freeze years (1983, 1985, and 1994) and late (L) freeze years (2007, 2011, and 2012)........................................................................................................143

Figure 4.17. DP December frequency anomalies versus the 1981-2010 DP December mean during the extreme early (E) freeze years (1983, 1985, and 1994) and late (L) freeze years (2007, 2011, and 2012)........................................................................................................144

Figure 4.18. MP January frequency anomalies versus the 1981-2010 MP January mean during Januaries immediately following the extreme early (E) freeze years (i.e. 1984, 1986, and 1995) and late (L) freeze years (i.e. 2008, 2012, and 2013).................145

Figure 4.19. DP January frequency anomalies versus the 1981-2010 DP January mean during Januaries immediately following the extreme early (E) freeze years (i.e. 1984, 1986, and 1995) and late (L) freeze years (i.e. 2008, 2012, and 2013).................146

Figure 4.20. MP February frequency anomalies versus the 1981-2010 MP February mean during Februarys immediately following the extreme early (E) freeze years (i.e. 1984, 1986, and 1995) and late (L) freeze years (i.e. 2008, 2012, and 2013).................147

Figure 4.21. DP February frequency anomalies versus the 1981-2010 DP February mean during Februarys immediately following the extreme early (E) freeze years (i.e. 1984, 1986, and 1995) and late (L) freeze years (i.e. 2008, 2012, and 2013).................148

Figure 4.22. MP March frequency anomalies versus the 1981-2010 MP March mean during March months immediately following the extreme early (E) freeze years (i.e. 1984, 1986, and 1995) and late (L) freeze years (i.e. 2008, 2012, and 2013).................149
Figure 4.23. DP March frequency anomalies versus the 1981-2010 DP March mean during March months immediately following the extreme early (E) freeze years (i.e. 1984, 1986, and 1995) and late (L) freeze years (i.e. 2008, 2012, and 2013).
LIST OF TABLES

Table 2.1. Monthly frequencies (%) of the 15 circulation patterns from 1979-2011. The season in which they occur most frequently, winter (Wi), spring (Sp), summer (Su), and autumn (Au) are indicated in the final row..........................................................59

Table 2.2. Statistically significant correlations between September ice extent (along 11 longitudes from 176°-126°W indicated at the top) and CP frequency from 1979-2011. Italicized values are significant at $\alpha = 0.05$ and bold values are significant at $\alpha = 0.01$. Positive and negative correlation coefficients are indicated by the appropriate sign after the corresponding correlation month. CPs 2, 4, 5, 6, and 9 had no significant correlations during the melt season and were therefore omitted from the table..............................................................60

Table 2.3. Same as Table 2.2, but for monthly teleconnections and CP frequency. CPs 2, 4, and 15 have no significant correlations over the melt season..............................................................61

Table 3.1. Mean sea-level pressure pattern (P) classification monthly mean frequency of occurrence (%) from 1979-2013..........................................................101

Table 3.2. Mean absolute correlation coefficients $|r_p|$ between the western Arctic freeze-up dates and the MSLP and 1000-500 hPa thickness AP frequencies by month. The total number of patterns which occur at least one day during half of the respective months of the time series (e.g. ≥18 Januaries) are used to calculate the $|r_p|$ and listed accordingly ($#P/T_{patterns}$).................................................................102

Table 3.3. Same as Table 3.1, but for the 1000-500 hPa thickness pattern (T) classification.................................................................103

Table 3.4. Statistically significant Pearson bivariate correlations between the respective warm season patterns and sea ice freeze-up dates for the Beaufort, Chukchi, and Western Arctic ($p \leq 0.05$). Numbers in columns 2-4 represent calendar months and the superscripts indicate the sign of the correlation coefficient........................................104

Table 4.1. North American Arctic weather stations with SSC weather types corresponding to Figure 4.2 for Alaska (AK), USA, Yukon Territory (YT), Northwest Territories (NWT), and Nunavut (NU), Canada, and Greenland (GL). Stations 3 (YRB), 17 (YFB), 20 (YZF), 24 (BET), and 25 (YXY) are not included in the composite analyses because their records terminate prior to 2013........................................151
PREFACE

Dr. Scott Sheridan helped develop many of the ideas that form the basis of this dissertation, including its two resulting manuscripts (i.e. Chapters 2 and 3) thus far. He was also involved in editing these manuscripts prior to submission and during their revision stages. The author of this dissertation was responsible drafting its content, including writing the aforementioned manuscripts, creating their respective figures and tables, and editing the documents during all phases of the review process.
ACKNOWLEDGMENTS

This dissertation is largely the result of many individuals who graciously offered their time and support to me in one way or another through my four years as a doctoral student. First off, the biggest “thank you” of all is owed to my family for encouraging me throughout this process. They are, and will always be, my greatest supporters and I cannot possibly thank them enough.

I would like to thank the Kent State University Department of Geography for affording me many professional development opportunities (i.e. teaching, research, and travel-related) as I completed my doctorate. My dissertation committee, comprised of Drs. Scott Sheridan, Tom Schmidlin, Mandy Munro-Stasiuk, and Joe Ortiz, provided lots of constructive feedback on my project, which motivated me to strive to improve its content during analysis, writing, and editing phases.

Two individuals have been especially influential through not only during my doctoral study, but throughout much of my academic career. I am grateful to my doctoral advisor, Scott Sheridan, for his consistent guidance and support. I appreciate all the time he spent introducing the intricacies of synoptic climatological classification to me as well as his willingness to engage in several different Arctic projects. I would also like to thank Tom Schmidlin for being a great mentor and collaborator over the years. Tom sparked my interest in polar research as an undergraduate student when I took his Polar and Alpine Environments class during Spring 2007. Out of all of the classes I have taken, that was undoubtedly my favorite course and marked the starting point of my present academic journey.
I have been fortunate to receive external research funding from a couple different sources. In particular, the Association of American Geographers (AAG) Dissertation Research Grant assisted with some conference and workshop-related travel expenses during 2013. I am also grateful for the opportunity to have been a National Science Foundation (NSF) Integrative Graduate Education and Research Training (IGERT) Fellow (Grant DGE 0904560) during the 2012-2013 academic year. That fellowship allowed me the opportunity to participate in interdisciplinary research projects and attend multiple conferences and workshops that undoubtedly had a positive professional impact.

A number of individuals provided technical support and data assistance, which undoubtedly made this dissertation project a reality. They are separately thanked at the conclusion of Chapters 2-4. However, I want to especially thank Mary Haley (NCAR/CISL) for allowing me participate in the NCL workshop during June of 2013 and for her time and willingness to answer follow-up questions thereafter that led to the creation of graphics in Chapters 3 and 4 and subsequent manuscripts.

Fellowship was essential to successfully completing this dissertation. In particular, I would like to thank past KSU Geography graduates Mike Allen, Cameron Lee, and David Widner for their friendship and mentorship during the past couple of years. I will always appreciate the time we spent exploring a wide range of topics both within and beyond the walls of McGilvrey Hall.
CHAPTER 1

INTRODUCTION AND BACKGROUND

1.1 Introduction

The boreal, high latitude cryosphere, where a portion of the Earth’s water resources are located in the frozen form, has witnessed rapid change since the inception of the modern satellite era in the late 1970s (e.g. Hodgkins, 2015). These changes include, but are not limited to, accelerated mass loss of the Greenland Ice Sheet (e.g. Box et al., 2012), thawing permafrost (e.g. O’Donnell et al., 2012), increases in autumn snow cover (e.g. Cohen et al., 2012), and perhaps most notably, substantial loss of Arctic sea ice cover (e.g. Stroeve et al., 2012). Each element of cryospheric change has its own associated causes and impacts, however those related to recent sea ice variability and change have piqued the interest of various climate researchers.

The ice pack has fundamentally changed throughout its annual cycle of variability, highlighted by the springtime maximum and late summer/autumn minimum ice cover. In particular, Arctic sea ice extent (i.e. the ice edge) has exhibited a negative trend over the last three decades, most notably during the September climatological minimum cover (e.g. Cavalieri and Parkinson, 2012). Spring melt onset and autumn freeze-up periods have shifted, indicating some change in melt seasonality paralleling the notable minimum extent
trends observed (Stroeve et al., 2014). There are a plethora of mechanisms responsible for
the sea ice changes and these can vary, along with the ice cover, throughout different time
periods. Some of these factors include changes to Arctic Ocean near-surface wind fields
(e.g. Ogi and Wallace, 2012; Ogi and Rigor, 2013), the influx of warm, sub-Arctic ocean
water (e.g. Shimada et al., 2006), increases in the solar heating of the boreal polar ocean
(e.g. Arndt and Nicolaus, 2014), and coupled interactions involving different phases of
atmospheric and oceanic teleconnections combined with large-scale hemispheric warming
(e.g. Ballinger and Rogers, 2014).

The broad demise of the ice cover has been associated with both physical and
societal impacts in recent years. Sea ice cover variations affect numerous aspects of the
climate system (e.g. Budikova, 2009), such as the energy budget (Lindsay and Schweiger,
2015), and may play a critical role in synoptic-scale weather pattern persistence (e.g.
Francis and Vavrus, 2012). Diminishing ice cover has broad ecological, cultural, and
economic impacts. For instance, observed ice losses have been linked to summertime
warming and plant growth/expansion across high latitude lands, as well as the abundance
and distribution of terrestrial and marine animal species (e.g. Post et al., 2013), which
directly affects the subsistence lifestyle of local inhabitants (e.g. George et al., 2004). The
recent summer decline of sea ice has also coincided with increases in shipping traffic,
exploration, and tourism across the Arctic Ocean (Lasserre and Têtu, 2015).

While the causes of pan-Arctic ice loss are well-documented and the impacts of
those losses are beginning to be better understood across space and time, it is less clear
what atmospheric factors are most responsible for marginal Arctic sea ice losses and what
the impacts of those regional losses are on the climate of the high and middle latitudes.
This realm of causes and consequences research is especially understudied in the western Arctic (Figure 1.1), where the most dramatic changes to the ice cover have been noted (e.g. Ballinger and Rogers, 2013). As such, this dissertation research provides new insights into regional ice-climate interactions by exploring the relationships between western Arctic warm season-dominant atmospheric patterns, the region’s ice cover and freeze onset, and the impacts of the aforementioned ice variability on North American Arctic weather conditions. The forthcoming subsections of this introductory chapter describe previous Arctic sea ice-climate research and will provide a framework to describe the objectives of the dissertation research. This chapter is structured as follows: Section 1.2 discusses the broad research previously conducted on the causes and impacts of Arctic sea ice variability. Section 1.2.1 highlights previous research involving Arctic sea ice extent and freeze-up variability during the satellite era, emphasizing western Arctic sea ice behaviors. Section 1.2.2 outlines some of the atmospheric causes of areal sea ice decline, while Section 1.2.3 summarizes the recent sea ice-related climate impacts research. Section 1.2.4 discusses some of the methods that have been previously used for studying sea ice-climate relationships. Section 1.3 concludes the chapter by indicating the motivation for the research, providing an outline for the remainder of the dissertation, and specifying the proposed outcomes of the project, including the broad contributions.

1.2 Background

1.2.1 Arctic Sea Ice Behaviors

Cryospheric research has been dedicated to the consistent monitoring of the Arctic sea ice pack over time. Pan-Arctic sea ice extent has declined in a statistically significant
manner during all months and seasons across the modern satellite era (November 1978-present), and these losses have been most significant during September, equating to roughly $3 \times 10^6$ km$^2$ from 1979-2013 (Simmonds, 2015). Recent ice behaviors have largely influenced the overall temporal variability in the end-of-summer cover measurements. In fact, the last ten Septembers (i.e. 2005-2014), which is the month of the climatological (i.e. mean) sea ice extent minimum, or period of maximum open ocean surface, in the Northern Hemisphere, represent the ten lowest sea ice extent measurements obtained during the modern period (National Snow and Ice Data Center (NSIDC), 2014). This suggests that over the past decade, Arctic sea ice has entered a new era where high summer losses are common. Aside from dramatic extent changes, other low frequency changes to the ice cover have transpired including dramatic thinning of the annual ice pack over much of the period (e.g. Lindsay and Schweiger, 2015), and melt season expansion through both earlier spring melt onset and delayed freeze-up in autumn (e.g. Stroeve et al., 2014).

While the overall degradation of the areal cover has been noted in previous research articles, regional studies of the Arctic marginal seas’ ice conditions provide insight into which portions of the ice edge are changing most dramatically since 1979. Recent manuscripts have indicated that the most significant spatiotemporal ice cover changes, especially during the melt season months (i.e. spring – autumn), have been observed in the Beaufort and Chukchi Seas that together comprise the western Arctic Ocean (e.g. Ballinger and Rogers, 2013). Considerable summer ice melt began in the region in the late 1990s (Maslanik et al., 1999), preempting more recent, massive end-of-summer ice cover losses, including 2007 (e.g. Comiso et al., 2008) and 2012 (e.g. Ballinger and Rogers, 2013). A fundamental change in the ice pack has paralleled the observed ice losses. The western
Arctic represents the sector of the Arctic Ocean where the greatest losses of the oldest, thickest, multiyear ice have taken place (Maslanik et al., 2011) as have the largest perennial, or first-year, sea ice losses (Comiso, 2012). Along with a transition to a younger, thinner, and more fragile ice pack, measurements of the region’s ice edge have revealed anomalous rates of change here, with respect to the other marginal seas, during the months of August through October from 1979-2012 (Xia et al., 2014). The authors also note that the September ice retreat rate (northward by 0.2° year^{-1} or 7° over the time series) was most robust in this portion of the Arctic Ocean (Xia et al., 2014).

The autumn freeze-up time period has also been substantially altered during the modern period. The fraction of open water observed during October (1979-2012) has increased by >44% in both the Beaufort and Chukchi Seas (Wendler et al., 2014), thereby delaying the formation of ice on the western Arctic Ocean surface over time. From 1979-2013, the continuous freeze date, marking persistence in freeze conditions through the winter months, has been delayed nearly 10 days decade^{-1} in the Chukchi Sea, the highest positive trend of all marginal seas, while Beaufort Sea has formed approximately 6 days decade^{-1} later over the period (Stroeve et al., 2014). These changes in freeze timing have dramatically prolonged the end of the western Arctic melt season in particular, simultaneously altering the seasonality of ocean-atmosphere heat exchanges and potentially the weather and climate in the high northern and middle latitudes (e.g. Serreze et al., 2009; Francis et al., 2009; Francis and Vavrus, 2012) as will be discussed at length in Sections 1.2.3.
1.2.2 Causes of Arctic Sea Ice Variability and Change

A significant amount of research has attributed interannual and/or long-term sea ice cover observations to variations in atmospheric conditions. Spatial scales of analysis vary for these projects focused on the causal mechanisms fueling sea ice responses. Associations between large-scale atmospheric variability, Arctic climate, and pan-Arctic sea ice changes (e.g. Rigor et al. 2002; Maslanik et al., 2007; Overland et al., 2008; Ikeda, 2012; Ogi and Wallace, 2012; Ogi and Rigor, 2013) are undoubtedly more common compared to regionalized ice-atmosphere interaction studies constrained to Arctic marginal sea domains (e.g. Rogers, 1978; Drobot and Maslanik, 2003; Wendler et al., 2010). For instance, Maslanik et al. (2007) noted that the Arctic Oscillation (AO), a principal mode of lower tropospheric atmospheric circulation variability in the Northern Hemisphere (Thompson and Wallace, 1998), was primarily in a positive phase (i.e. low Arctic surface pressure and largely zonal winds throughout the Arctic basin) in the 1980s and 1990s as the sea ice began to weakly decline. The AO shifted to more neutral conditions in the mid-2000s (Maslanik et al., 2007) followed by increasingly negative AO values (i.e. high Arctic surface pressure and predominantly meridional winds across the Arctic Ocean) in recent summers (e.g. Overland et al., 2012) as ice cover losses have been more pronounced relative to the weaker-loss years witnessed in the 1980s and 1990s.

The Pacific-North American Pattern (PNA; Wallace and Gutzler, 1981), which is a leading mode of Northern Hemisphere monthly middle tropospheric (i.e. 500 hPa) circulation anomalies, has also been tied to Arctic climate on decadal and subdecadal time scales (Overland et al., 2008). Much like the positive phase of the AO, a positive PNA (i.e. strong upper-level zonal winds across the Arctic Ocean) during the 1980s and 1990s was
associated with Arctic warming and patterns of sea ice variability in those decades but has since explained less Arctic climate variability during more recent years with few exceptions (Overland et al., 2008). The record-setting positive PNA during summer (i.e. July-September) 2007, which was roughly three standard deviations above the 1950-2007 mean, was associated with a large, persistent upper-air anticyclone over the western Arctic that likely contributed to the then record-setting ice losses of that summer (L’Heureux et al., 2008). It appears as though less sea ice variability is generally explained by these teleconnections, or drivers of large-scale natural climate variability, over the last decade (e.g. Ballinger and Rogers, 2014), except during years in which anomalous phases of these atmospheric mechanisms persist during melt season months (e.g. L’Heureux et al., 2008).

Atmospheric circulation within the Arctic basin is also driving some changes in sea ice behaviors over time. Ogi and Rigor (2013) noted a substantial drop in summer sea ice across most of the Arctic marginal seas coincident with a near-surface summer (June-September) wind shift at the 925 hPa atmospheric level from cyclonic (1979-1996) to anticyclonic (1996-2010) conditions. During the high ice loss years of 2007-2011, 925 hPa winds were much more anticyclonic versus those observed from 1979-2006 (Ogi and Wallace, 2012), perhaps driving the trend toward anticyclonic conditions within the basin since the mid-1990s (Ogi and Rigor, 2013). Overland et al. (2012) also connect changes in the Arctic Dipole (AD) pattern, the second mode of Northern Hemisphere mean sea level pressure (MSLP) variability poleward of 70°N (e.g. Wang et al., 2009), to recently observed changes in high-latitude synoptic fields. The authors link the common occurrence of a negative, summer AD pattern from (2007-2012) with higher-than-average MSLP and 700 hPa positive meridional winds over the western Arctic. The persistence of southerly winds
would not only dynamically force ice away from the coast, but also advect warm air and moisture into the region and aid in the extreme ice edge melting during the last several years.

Some studies have focused on even smaller spatial domains to assess atmospheric mechanisms and their interactions with sea ice cover. For instance, the Beaufort Sea High (BSH), an anticyclonic pattern common over the western Arctic Ocean during the melt season, plays an important role in interannual, regional-to-areal sea ice variability depending on its geographic positioning and persistence (e.g. Serreze and Barrett, 2011). Rogers (1978) found the Beaufort summer sea ice cover to be lighter (heavier) with higher (lower) MSLP across the Beaufort Sea. Further, high ice loss summers often have a more intense pressure gradient and strong easterly/southeasterly surface winds that force ice out of the region, whereas low loss summers often experience a weaker gradient and weakened regional winds (e.g. Drobot and Maslanik, 2003). Observations have shown a strengthening of the summertime BSH pressure field since the mid-1990s (Moore, 2012) concurrent with an increase in its frequency during early summer (Ballinger et al., 2014). The change in frequency and strength of the BSH pattern may be driven by high latitude warming and/or increasingly amplified, wavy atmospheric circulation fields across the high and middle latitudes during the last two decades (e.g. Moore et al., 2012; Overland et al., 2012; Ballinger et al., 2014; Francis and Vavrus, 2015).

Observed changes in atmospheric conditions, specifically toward a more common large-scale anticyclonic circulation regime during summer, have yielded clearer skies and led, in part, to significant amounts of downwelling radiation being absorbed by the Arctic Ocean (e.g. Perovich et al., 2007; Kay et al., 2008; Perovich et al., 2011; Stroeve et al., 2014).
Perovich et al. (2011) noted increasing trends in solar heat flux across large swaths of the Arctic Ocean, maximized in the Chukchi Sea at roughly +4% decade\(^{-1}\). Kay et al. (2008) found that summertime solar irradiance generally rose at Barrow, AK from 1998-2007, especially from 2006 to 2007 when skies in the western Arctic region were clearer. Earlier spring sea ice melt onset during the last three decades also explains some of the increases of solar radiation absorption into the open ocean (Stroeve et al., 2014). These interconnected processes have prompted Arctic sea surface temperatures (SSTs) to increase and sea ice melt rates to accelerate, especially during 2007-2011 (e.g. Stroeve et al., 2014). In sum, higher warm season SSTs and lower tropospheric air temperatures, partially driven by atmospheric circulation processes, enhance the ice-albedo feedback process, leading to more melt and later autumn freeze-up over time.

1.2.3 Impacts and Implications of Arctic Sea Variability and Change

Some of the relationships between atmospheric circulation and Arctic sea ice variability outlined in Section 1.2.2 have been studied for decades. Far less studied and surveyed by climate researchers, until the last decade, are the potential impacts of sea ice changes on aspects of Northern Hemisphere weather and climate, including spatiotemporal variability in storm tracks, temperature, and precipitation (e.g. Dethloff et al., 2006; Budikova, 2009; Higgins and Cassano, 2009; Serreze and Barry, 2011; Screen, 2014; Vihma, 2014). The sequence of consecutive large sea ice loss summers have preceded some extreme weather events across the boreal high and middle latitudes, such as cold outbreaks during several recent winters (e.g. Overland et al., 2011; Cohen et al., 2012; Francis and Vavrus, 2012; Ballinger et al., 2014). This has motivated some researchers to investigate
links between the sea ice changes and persistence in large-scale atmospheric patterns which may in turn be driving some of the recent, regional weather and climate anomalies (e.g. Francis et al., 2009; Overland et al., 2011; Cassano et al., 2014).

Arctic amplification (AA), the strong surface air temperature (SAT) warming of the Arctic relative to the overall warming observed throughout the Northern Hemisphere (e.g. Serreze and Barry, 2011), may be driving the proposed changes in circulation. Arctic surface temperature warming has been observed both in the terrestrial weather station datasets (e.g. Bekryaev et al., 2010) as well as the hindcast weather/climate model simulations, often referred to as global reanalyses (e.g. Screen et al., 2012), particularly since 1995 (Francis and Vavrus, 2015). AA, while strongest at the surface (e.g. Serreze et al., 2009), has also been detected alongside increases in humidity in the lower-to-middle troposphere observational (i.e. radiosonde) and reanalysis data (e.g. Serreze et al., 2012). While other factors such as earlier spring snow melt onset (e.g. Brown et al., 2010) and poleward heat flux (e.g. Skific and Francis, 2013; Lee and Yoo, 2014) explain some of the variability in the lower tropospheric AA signal, a large amount of the near-surface Arctic warming has been directly attributed to the catastrophic, recent decline of the summer sea ice cover (e.g. Kumar et al., 2010; Screen and Simmonds, 2010).

Massive summer sea ice losses have altered both Arctic ocean-atmosphere heat exchanges at high latitudes, but also influenced the equator-to-pole temperature gradient that drives atmospheric circulation across the hemisphere. During summer, sea ice represents a barrier between the warm ocean surface and the cold atmosphere. Significant warm season Arctic Ocean ice melt, especially manifested with large losses of the September minimum extent (e.g. Simmonds, 2015), reveals more ice-free fetches that
prolong freeze-up during the autumn season (i.e. October-December). Extended periods of open water in autumn allow more ocean-to-atmosphere sensible and latent heat exchanges to take place (e.g. Serreze et al., 2009). These exchanges effectively warm the lower troposphere in the Arctic, but also act to expand the atmospheric column thereby reducing the temperature gradient that typically exists between the high and middle latitudes (e.g. Francis and Vavrus, 2012). This large-scale reduction in the hemispheric temperature gradient may be driving the increasingly meridional, or wavy, upper-level atmospheric circulation that has been observed in recent years (e.g. Francis et al., 2009; Francis and Vavrus, 2012), especially over North America during winter (i.e. January – March) and summer (i.e. July – September) since 1995 (Francis and Vavrus, 2015). Studies have also noted that sub-Arctic SST changes (e.g. Screen and Simmonds, 2013) and natural climate variability (e.g. Overland et al., 2011), along with the sea ice losses, may also be driving some of the increasing meridionality in atmospheric circulation.

Previous studies have recognized a lower troposphere thermal response across portions of North America following high pan-Arctic sea ice loss years (e.g. Royer et al., 1990; Murray and Simmonds, 1995; Budikova, 2009). Effects of sea ice-related AA have been examined at multiple atmospheric levels from the surface to the middle troposphere, including SAT (e.g. Wendler et al., 2014), MSLP (e.g. Overland et al., 2008), 850 hPa temperatures and geopotential heights (e.g. Overland et al., 2011), and 1000-500 hPa thicknesses (e.g. Francis and Vavrus, 2012), and 500 hPa zonal and meridional wind fields (e.g. Francis and Vavrus, 2015). While these studies have suggested that AA affects different synoptic fields, the geographic extent of the AA impacts largely remains in question.
Localized impacts of pan-Arctic sea ice behaviors are apparent in the SAT warming trends witnessed at high latitude coastal weather stations over the satellite era (e.g. Bekryaev et al., 2010). However, large-scale analyses of atmospheric fields, using composite and multiyear aggregation techniques, show some spatial inconsistencies in climatic responses. For instance, Francis et al. (2009) showed that autumn (i.e. October – November) MSLP (SAT) is lower across portions of eastern (central) North America during high versus low ice loss summers, and Francis and Vavrus (2012) revealed statistically significant 1000-500 hPa thickness anomalies (2000-2012 versus 1970-1999) during autumn (i.e. October – December) and winter (i.e. January – March) over easternmost Canada. Interannual studies comparing the atmospheric conditions following disparate high loss years do not necessarily show a consistent spatial response. Overland et al. (2011) displays 850 hPa geopotential height fields during December 2009 and 2010. While heights are generally lower (indicative of colder lower tropospheric air temperatures) over the North Pacific and eastern North America, height minima are shifted slightly to the east across both regions in 2010 versus 2009.

1.2.4 Datasets and Methods Often Employed in Sea Ice-Climate Studies

Sea ice-climate interaction studies typically associate atmospheric conditions (e.g. MSLP and/or 1000-500 hPa thickness) to sea ice variability (or vice-versa) across either multidecadal time periods or during extreme ice years, when anomalously high or low ice conditions are found across the Arctic Ocean. Two ice metrics are used throughout this dissertation to assess sea ice-climate interactions, sea ice extent and freeze-up dates. Most sea ice measurements, including the sea ice extent and freeze-up measurements used in
this dissertation, are derived from passive microwave satellite sources, dating from 1979 onward, and the output is often gridded across the Arctic Ocean. Sea ice extent is measured in terms of the fraction of ice cover (i.e. ice concentration) observed within a particular grid box. This amount typically ranges from 15-50% coverage, depending on the study, and is used to trace the ice edge for the domain and time period specified (e.g. Ballinger et al., 2013; Ballinger and Rogers, 2014). Freeze-up dates are observed when the sea ice cover completely forms over an area of the Arctic Ocean. This date represents the time, typically during the autumn season, when open ocean is no longer observed and ice completely and continuously fills the grid box from that date until melt occurs during the following spring (Markus et al., 2009). All grid boxes in a particular domain are then averaged to get a mean freeze-up date across the marginal sea of interest. More details regarding the specifics of the sea ice datasets and their origins are outlined in Chapters 2-4 of this dissertation.

Most of the aforementioned ice-climate studies discussed throughout this introduction also share similar datasets and aggregation techniques. The common type of data used in these analyses is derived from one of the global reanalyses. The premise of a “reanalysis” is that climatic variables are not observed at consistent spatial resolutions across Earth’s surface. As such, reanalyses take historical data from various sources, including but not limited to terrestrial weather stations, radiosondes, buoys, ships, aircrafts and satellites, and integrate that data into fixed numerical weather prediction models and data assimilation schemes. These climate models then hindcast weather/climate observations with reasonable accuracy at high spatial and temporal resolutions across the globe (e.g. Kalnay et al., 1996).
There are a number of different climate research centers across the world that have created reanalysis datasets covering various retrospective time periods. For this dissertation, the daily-averaged atmospheric circulation data, specifically MSLP and geopotential height fields, are taken from the National Centers for Environmental Prediction/National Center for Atmospheric Research (NCEP/NCAR) first generation reanalysis (Kalnay et al., 1996). All NCEP/NCAR reanalysis data are available at a 2.5° horizontal resolution across the globe and date from 1 January 1948, however only data since 1 January 1979 are used here to overlap the modern satellite-derived sea ice data. The MSLP and geopotential height fields are labeled as Class A variables, indicating that they closely follow the observational data assimilated into the model (Kalnay et al., 1996) and therefore yield similar results to other available reanalysis products. MSLP data is used to assess near-surface atmospheric flow, while 1000-500 hPa thickness, found by subtracting the 1000 hPa height field from the 500 hPa height field at each gridpoint in the domain for each day of the analysis period, is employed to examine the thermal character of the atmosphere from the middle troposphere to the surface as well as better understand atmospheric circulation in that layer. Further justification of the NCEP/NCAR product for atmospheric analysis across the Arctic domains studied is provided in each of the results chapters.

Most contemporary studies of Arctic sea ice-climate interactions desire a better understanding of the atmospheric characteristics in the months preceding, or following, a massive sea ice loss event or a notable change in low frequency ice behavior. These interactions are often explored via monthly mean or seasonal aggregations (i.e. summer and autumn) of reanalysis data followed by subsequent comparing/contrasting of select
atmospheric fields during high versus low sea ice loss years or a climatological period. Comparisons of short-term anomalies or anomalous loss years versus multidecadal mean atmospheric conditions are useful for diagnosing recent Arctic change, but fail to identify daily atmospheric patterns and surface weather conditions and their frequency changes over time.

This dissertation takes a different approach versus traditional sea ice-climate studies, which commonly use the aforementioned aggregation/compositing procedures, by employing synoptic climatological techniques and datasets, including circulation pattern classification (also referenced as atmospheric or map-pattern classification) and weather typing (Yarnal, 1993). These approaches holistically categorize the range of atmospheric conditions experienced over a location. More specifically, circulation pattern classification partitions an atmospheric level into typical patterns of large-scale flow, and weather typing describes surface weather conditions at a particular locale. One of the primary objectives of synoptic climatology is to relate these coherent groups of circulation patterns or weather types to the surface environment (Yarnal, 1993). In this dissertation, circulation pattern classifications will be created from the reanalysis datasets and linked to the September sea ice extent minima (Chapter 2) and freeze-up dates (Chapter 3) to better understand the influence of the daily circulation types on these ice variables across time. Data from one particular weather typing scheme, the Spatial Synoptic Classification (SSC; Sheridan, 2002), is also used to understand the surface weather responses across high-latitude North American to freeze-up date variability over time (Chapter 4). More detailed description about the creation of the circulation pattern datasets and the components of the SSC are provided in the aforementioned results chapters.
1.3 Motivation for Research, Layout, and Summary of Dissertation Objectives

As highlighted in Sections 1.2.2 and 1.2.3, the majority of sea ice change research has linked pan-Arctic ice conditions and continental-to-hemispheric large-scale atmospheric circulation. There are relatively few studies exploring regional sea ice-climate interactions, particularly in the western Arctic where the most substantial ice changes have been documented (e.g. Comiso 2012; Stroeve et al., 2014; Xia et al., 2014). More specifically, there is a need to better understand the relationships between atmospheric circulation patterns and the sea ice conditions within the western Arctic during times in annual cycle when the ice cover is changing dramatically (i.e. the September minimum extent and autumn freeze-up) and may be significantly impacting large-scale weather and climate patterns beyond the region (e.g. Francis and Vavrus, 2012). Links between the subseasonal (i.e. daily) spatiotemporal variability of the typical western Arctic synoptic circulation regimes and sea ice minimum and freeze-up conditions have yet to be explored by the cryospheric research community. Similarly, the extent to which western Arctic freeze-up variability relates to high latitude North American surface weather conditions across space and time is uncertain.

This dissertation research takes a regional approach to better understand some of the atmospheric causes and impacts of dramatic western Arctic sea ice changes since 1979. The first two results chapters specifically examine the causes of ice variability, while the final results chapter is concerned with the impacts of the ice conditions. The remaining chapters of this dissertation are organized as follows: Chapter 2 explores interactions between warm season MSLP circulation patterns and the September sea ice extent minimum. Chapter 3 analyzes relationships between warm season MSLP and 1000-500
hPa thickness patterns as they relate to the variability of the autumn western Arctic freeze-up dates. Chapter 4 evaluates the links between the aforementioned freeze-up timing and the subsequent monthly temporal variability of SSC weather types across the terrestrial North American Arctic. Chapter 5 provides a synthesis of the results, discusses the pertinent findings, and briefly proposes future research directions.

The goals of this dissertation research are multifold, and aspire to broadly fulfill clear, interconnected needs in the cryospheric research community to better understand the regional links between ice variability in a massive change region and daily atmospheric circulation and surface weather conditions. Chapter 2 aims to contribute to the literature by resolving a comprehensive, high frequency MSLP catalogue for the western Arctic and statistically assessing how frequencies of these warm season patterns relate to and explain the end-of-summer sea ice extent variability in the region. Chapter 3 seeks to expand the limited understanding of the mechanisms influencing sea ice freeze-up by evaluating the roles of western Arctic lower tropospheric circulation (i.e. MSLP) and thermal (i.e. 1000-500 hPa thickness) pattern variability on its timing. Chapter 4 strives to enhance recent sea ice-related impacts research by analyzing links between western Arctic freeze-up and the spatiotemporal variability of autumn and winter SSC-derived surface weather types across the North American Arctic.

The regional sea ice-climate results derived from this dissertation research will also benefit the short-to-medium range objectives of operational meteorologists, especially associated with North American agencies such as the National Oceanic and Atmospheric Administration (NOAA) and Environment Canada. These results will show common atmospheric patterns and surface weather responses that occur around variable sea ice
conditions. The results will likely also be of interest to climate and socio-ecological modelers seeking to better understand how regional changes in high latitude climate may continue to influence the sea ice cover, ecosystems, and society and subsequently impact large-scale human-environment interactions.

Results generated and analyzed through this dissertation research will also make a methodological contribution to the synoptic climatology subfield of geography. Synoptic climatological tools have been used infrequently in previous cryospheric studies (as mentioned in each results chapter); however this research uses these techniques, along with statistical methods, to delve into western Arctic sea ice-climate interactions in detail. Fulfillment of the research objectives summarized above will therefore have multi-dimensional impacts, benefitting future interdisciplinary research spanning societal, environmental, and climatic realms.
References


Last accessed February 18, 2015.


Figure 1.1. Map of the western Arctic (WA) domain. The Beaufort Sea (BS) and Chukchi Sea (CS) comprising the western Arctic Ocean are outlined.
CHAPTER 2

ASSOCIATIONS BETWEEN CIRCULATION PATTERN FREQUENCIES AND SEA ICE MINIMA IN THE WESTERN ARCTIC

Abstract

In this study, a synoptic climatological approach is employed to assess the relationship between the frequency of circulation patterns (CPs) and the latitude of mid-September sea ice minima in the western Arctic. Fifteen CPs are created via principal component analysis and cluster analysis from daily NCEP/NCAR reanalysis sea-level pressure (SLP) fields across a grid from 50-90°N and 150°E-100°W from 1979-2011. The frequency of these CPs are statistically compared with the latitude of the sea ice minimum from passive microwave data for each of 11 equally-spaced longitudes (176°W to 126°W) extending into the Chukchi and Beaufort Seas. Monthly frequencies for each of the 15 CPs from March to September, signifying the melt season, for each year are correlated with the ice minimum for that September. These monthly frequencies are then entered into a stepwise multiple linear regression (SMLR) and collectively, CP frequencies explain 40-79% of the total ice retreat variance across the longitudes. The frequency of one cluster, CP 11, representing a broad high pressure area over the Beaufort Sea, is highly correlated with the latitude of the

sea ice minima; June and August frequencies of this pattern are the initial predictors at 8 of the 11 longitudes and explain 22-32% of the variance. This pattern has occurred more frequently from 2007 onwards; compared with a June mean occurrence of 9 days during 1979-2006, CP 11 occurred 16 times in June 2007, and on average more than 17 days per month during June 2008-2011. The Arctic Dipole (AD), Arctic Oscillation (AO), and Pacific-North American pattern (PNA) indices are significantly correlated with CPs 11-13 frequencies throughout certain summer months, further indicating strong relationships between summer circulation and sea ice minima in the region.

2.1 Introduction

September Arctic sea ice extent has decreased nearly 12% decade⁻¹ since 1979 (Stroeve et al., 2012) with the largest extent losses occurring in the western Arctic, including the Beaufort and Chukchi Seas (Perovich and Richter-Menge, 2009). This region’s summer sea ice losses in the 1990s (Maslanik et al, 1996; 1999) and 2000s (Stroeve et al., 2005; Maslanik et al., 2007b) have largely fueled the recurring Arctic-wide record minima over the last two decades. During this time thick, multi-year ice (MYI) has dramatically declined in the Beaufort Sea and Canada Basin (Kwok and Cunningham, 2010; Maslanik et al., 2011; Derksen et al., 2012) amidst a transition to thin, fragile first-year, or seasonal, ice cover that has also rapidly dissipated (Barber et al., 2012). Several interconnected factors have potentially contributed to recently observed changes to the summer ice cover in the region including the ice-albedo feedback (e.g., Lindsay and Zhang, 2005), oceanic heat flux from the Pacific Ocean (e.g., Woodgate et al., 2010), solar heating of
the open ocean (e.g., Perovich et al., 2007), and aspects of regional atmospheric circulation (e.g., Ogi and Wallace, 2012).

Synoptic circulation patterns occurring throughout the melt season (from the March maximum to September minimum in the Northern Hemisphere) have been directly connected to western Arctic ice extent behavior during summer. Though high pressure over this region is typically stronger and broader due to enhanced baroclinity in winter and spring (Serreze and Barrett, 2011), summer anticyclone activity is arguably more important in directly dictating summer ice loss. A strong Beaufort High during summer promotes ice loss due to warm air advection (Rogers, 1978) and increases ice export out of the region (Drobot and Maslanik, 2003). In recent years, this pattern has been especially pronounced in conjunction with record ice extent losses. For instance, Ogi et al. (2008) found mean summer (JAS) SLP over the Beaufort Sea in 2007 to be higher than during any summer since 1979. The persistence of this synoptic setting permitted a reduction in western Arctic cloudiness (Kay et al., 2008) and strong solar heating of the ocean which propelled the anomalous melt (Perovich et al., 2008).

Cyclonic activity also has an important bearing on the summer ice edge. Climatologies of broad-scale low pressure activity in the Arctic indicate that frequency typically increases through spring and peaks in summer (Serreze et al., 1993). Simmonds and Keay (2009) indicated that strength, as opposed to frequency, of recent migratory summer storms across the Arctic basin has substantially contributed to the negative September ice cover trend. Case studies by Asplin et al. (2012) and Parkinson and Comiso (2013) showed that strong winds and large waves from cyclones passing over open
western Arctic waters in early September 2009 and August 2013, respectively, aided in the breakup of an already retreating ice edge.

In connection with synoptic patterns, atmospheric and oceanic teleconnections also have multifaceted and temporally varying associations with ice in the region. The Arctic Dipole (AD) has been linked to strong meridional wind forcing and export of sea ice out of the Beaufort basin during recent high loss summers (Wang et al., 2009; Overland and Wang, 2010). The Pacific-North American pattern (PNA) was three standard deviations above its mean during summer 2007 with a pronounced 500 hPa ridge over the western Arctic (L’Heureux et al., 2008). Rigor and Wallace (2004) found the negative phase of the summer Arctic Oscillation (AO) was associated with low sea ice concentrations in the Beaufort and Chukchi Seas due to warm air advection off the adjacent land and southerly, wind-driven transport of ice out of the region. The Pacific Decadal Oscillation (PDO), which has been mostly in positive phase since the mid-1970s, has also been associated with changes in Pacific sector sea ice and western Arctic surface air temperature in recent decades (Hartmann and Wendler, 2005; Danielson et al., 2011).

Studies of regional atmospheric circulation and teleconnections over the Arctic Ocean often focus on one or two pressure features that have pronounced climatic impacts. Rather than focus solely on distinct spatial and temporal characteristics of specified semi-permanent pressure features, synoptic climatology provides an opportunity to identify a range of weather conditions related to the surface environment (Yarnal, 1993). There are two distinct synoptic classification realms: weather typing, which characterizes surface weather conditions, and circulation pattern identification, which partitions a single atmospheric level into common synoptic-scale flow regimes (Yarnal, 1993). In the western
Arctic, weather typing (e.g., Kalkstein et al., 1990) and hybrid approaches that integrate both classification schemes (e.g., Mülmenstädt et al., 2012) have been seldom used. Circulation-pattern classifications have been mainly used to evaluate surface circulation (i.e. SLP) effects on local temperature, precipitation, and wind regimes in the region (e.g., Cassano et al., 2006; Cassano and Cassano, 2010; Cassano et al., 2011). Few studies have examined relationships between atmospheric circulation patterns and sea ice (e.g., Maslanik et al., 2007a; Asplin et al., 2009; Higgins and Cassano, 2009).

The purpose of this paper is to create a synoptic circulation classification for the western Arctic in order to depict the array of SLP patterns that may play a role in the mid-September sea ice minimum in the Beaufort and Chukchi Seas. Aside from creating a holistic representation of SLP patterns over the region, this study focuses on the frequency of these events over the course of the melt season and their statistical relationships to the sea ice over time. Statistically significant relationships between monthly circulation patterns and ice across multiple longitudes are further examined in the context of recent years, during which record loss of sea ice has been observed. Relationships between circulation patterns and teleconnections known to impact western Arctic ice are also explored. This paper is divided up as follows: Section 2.2 presents the data and methodology in detail. Section 2.3 covers the results including an introduction of the synoptic patterns, statistical connections to the ice, interannual variability/evolution of select patterns, and circulation pattern relationships to atmospheric and oceanic circulation teleconnections also relevant to western Arctic climate. Section 2.4 provides a discussion of the main findings. Section 2.5 briefly summarizes the results and discusses
the potential for synoptic climatological applications in further assessing sea ice-climate interactions in the Arctic.

2.2 Data and Methodology

2.2.1 Data

Reanalysis datasets are particularly useful over the polar regions in part because weather stations in these areas are sparsely configured and typically have short temporal resolutions. In this study, we use daily mean SLP data from the NCEP/NCAR Reanalysis (Kalnay et al., 1996) on a 2.5° x 2.5° latitude/longitude grid (a total of 765 points) across the domain of 50-90°N and 150°E-100°W (Figure 2.1) from 1979-2011 to capture the SLP patterns associated with the Chukchi and Beaufort sea ice minimum in mid-September. SLP is utilized in this study because it can be used to indicate a number of different surface conditions including wind speed, wind direction, cloud cover and sunlight over an area, all of which play a vital role in modulating the ice cover. NCEP/NCAR circulation fields, SLP and otherwise, have been evaluated extensively in the Arctic and are known to produce comparable results to other reanalyses over the region (e.g., Serreze et al., 2009; Cassano et al., 2011; Serreze and Barrett, 2011). These data are obtained from NOAA/ESRL Physical Sciences Division.

Sea ice edge data for mid-September, which is the approximate climatological minimum (Comiso, 2002), are obtained from two different passive microwave datasets, HadISST and AMSR-E. Hadley Center Sea Ice and Sea Surface Temperatures (HadISST) monthly median sea ice concentration fields on a 1° x 1° grid, 1979-2007, are from the Met Office Hadley Center (Rayner et al., 2003). The sea ice edge from 2008-2011 is from the
Advanced Microwave Scanning Radiometer-EOS (AMSR-E) archive available at the University of Bremen. The AMSR-E sea ice charts date back to 2003 and these data are well correlated with the HadISST-derived ice edge for overlapping years 2003-2007 (r = 0.96 – 0.99 depending on the longitude). Ice extent for mid-September, in terms of latitudinal retreat, is found in both datasets using the 50% ice concentration threshold along 11 longitudes extending from the Alaskan and Siberian coasts into the Chukchi and Beaufort Seas (Figure 2.1). These longitudes are located at equally-spaced 5° intervals from 176°W (hereafter L176) in the eastern Chukchi Sea to 126°W (L126) in the eastern Beaufort Sea. The latitude of the ice edge is determined at sea unless the sea ice is fast to the coastline, in which case the extent is equal to the coastal latitude along the given longitude.

Four monthly atmospheric/oceanic teleconnection indices are also assessed in relation to the ice extent minimum. The Pacific North American (PNA) pattern (Wallace and Gutzler, 1981) is one of the leading modes of Northern Hemisphere atmospheric variability and is created using a rotated principal component analysis (RPCA) technique on monthly mean 500 hPa height anomalies poleward of 20°N. Similar to Thompson and Wallace (1998), the Arctic Oscillation (AO) is constructed by projecting the daily 1000 hPa height anomalies poleward of 20-90°N onto the loading pattern of the AO. The AO and PNA indices are obtained from the Climate Prediction Center (CPC). The Arctic Dipole (AD) is defined as the second Empirical Orthogonal Function (EOF) of the extended winter (NDJFM) mean SLP anomaly north of 70°N (Wang et al., 2009; Overland and Wang, 2010), and were provided by James Overland and Muyin Wang. The Pacific Decadal Oscillation (PDO) is the leading principal component of monthly sea surface temperature (SST)
poleward of 20°N (Mantua et al., 1997) and these data are available from the Joint Institute for the Study of Atmosphere and Ocean (JISAO) at the University of Washington.

2.2.2 Methodology

The synoptic climatological classification used in this research involves a two-step synoptic climatological method employing both principal component analysis (PCA), which serves to reduce multicollinearity in the data set (Wilks, 2011), and cluster analysis, in which different SLP pattern types are created (Yarnal, 1993). Daily SLP values for each of the 765 grid points are entered into an S-mode PCA without any seasonal adjustments, as it is desired to identify interseasonal variability in the identified patterns. From the 765 data points, 26 principal components meet the generally accepted threshold for retention (an eigenvalue of at least 1), explaining 98.2% of the total SLP dataset variability over the specified domain.

These 26 retained principal components are then entered into a two-step cluster technique (TSC), which classifies each day’s spatial SLP features into an appropriate cluster. The primary goal of cluster analysis is to minimize variability within each cluster while maximizing the variability between clusters, though some within cluster variability must be assumed (Yarnal, 1993). Spekat et al. (2010) has suggested saving roughly 10-15 clusters for 30 years of daily data. When seasonality is not removed, there is a tendency for a greater number of winter-dominant clusters than summer-dominant ones, therefore the decision was made to save 15 clusters in order to better distinguish among summer CPs in particular, especially since relationships between end-of-summer sea ice extent and melt season-dominant patterns are the focus of this paper. These patterns are created in SPSS
using the log-likelihood distance measure and Schwarz’s Bayesian information clustering criterion metrics (see Coleman and Rogers, 2007 for more TSC details).

Pearson’s bivariate correlations between annual sea ice minima at each longitude and monthly CP frequencies that precede it are evaluated in detail. Correlations are only calculated for CPs whose mean monthly frequency exceeds 5% of days in months preceding the minimum for a particular melt season (March-September; more description in Section 2.3). In all, 47 total months across the 15 clusters met this frequency threshold. The monthly CP frequencies are also introduced into a stepwise multiple linear regression (SMLR) model in order to evaluate their ability to hindcast mid-September sea ice extent along the 11 longitudes. SMLR involves a forward selection process whereby the initial predictor (CP frequency) explains the most predictand (sea ice) variance while subsequent predictors retained in the prediction equation must explain a significant amount of sea ice variance. All statistical results described are statistically significant at the 0.05 level unless noted otherwise.

2.3 Results

2.3.1 CP Descriptions and Relationships with Sea Ice

Due to the long seasonal lag evident in the Arctic, seasons are defined as follows: winter (JFM), spring (AMJ), summer (JAS), and autumn (OND). The 15 CPs are presented in Figure 2.2 and their monthly frequencies in Table 2.1. Patterns occurring more often from late winter (March) through the end of summer (September) are examined in greater detail as they occur during the melt season for each calendar year.
CPs 4, 6, 8, and 9 primarily occur during winter and exhibit fairly strong pressure gradients with pronounced low pressure centers over the North Pacific and high pressure areas over adjacent northeastern Siberia, eastern Alaska, Yukon and Northwest Territories and western Arctic Ocean. CPs 10 and 14 frequently occur during spring while patterns 11-13 mainly transpire during summer. These have much weaker pressure gradients as expected with a decreased thermal gradient. Of the three summer-dominant patterns: CP 11 exhibits a dome of high pressure over the Beaufort Sea (~1017 hPa), CP 12 has a broad, weak area of low pressure over that area (~1007 hPa) with slightly higher pressure just north of the Chukchi Sea (~1014 hPa), and CP 13 features a latitudinal pressure gradient, with fairly low pressure over the pole (~1001 hPa) increasing equatorward.

Correlations between the longitudinal ice extent minima and monthly CP frequencies are displayed in Table 2.2. Certain spring and summer circulation patterns show particularly consistent and statistically significant connections to the ice. Specifically, CP 11 June shows a statistically significant positive correlation to the ice minimum at all longitudes ($r = +0.42 - +0.59$) while CP 11 August also displays this relationship with the exception of L136 and L131. CP 13 June demonstrates a significant negative correlation across all longitudes ($r = -0.37 - -0.49$) while CP 13 August only displays this relationship at L176-L166. Overall, more monthly CP frequencies are correlated to the ice at the easternmost longitudes, near the Canadian Archipelago, versus those in the western part of the domain, though some of these have associations with autumn-dominant patterns.

Given a number of strong connections to the ice cover, it is of no surprise that the late spring/summer circulation patterns themselves also display a high level of predictability regarding the ice extent. The SMLR results (not shown) from all
aforementioned CP frequencies explain between 40% (L151) and 79% (L141) of the mid-September ice extent variance across all longitudes from 1979-2011, with the variance explained generally increasing eastward (Figure 2.3 a-c). September minimum extent in the eastern Beaufort is clearly less variable than in the western Beaufort and Chukchi seas, which may explain part of the reason why circulation is a better predictor of ice extent in this area. Figures 2.3a (L176) and 2.3b (L151) show large amounts of unexplained variance over the time series, especially in the initial 10 years of the time series when the regression model shows a positive bias and overpredicts the observed ice extent latitude by ~1-5°. During the last five years (2007-2011), there are large positive residuals from under prediction at L151, but surprisingly not at L176. The entire 33 year sea ice extent time series at L126 is well explained by circulation ($r^2=0.74$). In terms of individual patterns, CP 11 June and August are the initial predictors at 8 of the longitudes (L176-L141) accounting for between 22% and 32% of the explained variance.

2.3.2 Late Spring/Early Summer Circulation Pattern Evolution

Analysis in Section 3a reveals that late spring/summer patterns are particularly well related to the western Arctic ice extent minima from 1979-2011. An examination of recent years (since the dramatic decline of 2007) versus the climatology (1979-2006) is warranted in order to address the role certain patterns, especially in the months directly preceding the minimum, may be playing on recent, historic mid-September losses in the region. Figure 2.4 shows the monthly frequencies of CPs 11-13 for June, July and August over three temporal delineations, 1979-2006, 2007, and 2008-2011.
CP 11 occurred on roughly one-third of June days ($\mu = 9$ days mo$^{-1}$, $\sigma = 5.35$) from 1979-2006. However, during recent high ice loss years the June frequency of this pattern has increased and occurred on 16 occasions in 2007, which was $1.31\sigma$ above the 1979-2006 period mean, and transpired 17.3 times mo$^{-1}$ on average from 2008-2011. Prior to 2007, CP 11 had only occurred more than 14 times during June on two occasions (1983 and 1997), however from 2007-2011 no June month experienced less than 15 days of CP 11. Increased persistence of CP 11 June coincided with declines in other summer patterns during that month, particularly CP 13 June whose occurrence decreased $\sim$4 days mo$^{-1}$ from 2007-2011 relative to the 1979-2006 average ($\mu$=8.1 days mo$^{-1}$). CP 12 June frequency also declined from the (1979-2006) average during both 2007 ($\sim$2 days mo$^{-1}$) and 2008-2011 ($\sim$3 days mo$^{-1}$).

Though July frequencies for CPs 11-13 are minimally correlated with mid-September ice extent, their frequencies have changed substantially as well. During 2007, CP 11 July occurred 24 times or nearly 15 occasions more than the 1979-2006 average ($\mu=9.1$ days mo$^{-1}$, $\sigma=5.68$) and $2.62\sigma$ above the 1979-2006 mean. The CP 11 July pattern has continued to be more common in the recent period, occurring 11.5 days mo$^{-1}$ from 2008-2011 at the expense of the summer-dominant patterns that were climatologically more common.

The evolution of August frequencies of CPs 11-13 is also worth noting. In 2007, CP 11 August occurred 18 times, which is nearly 13 more days than the extended August climatology ($\mu=5.3$ days mo$^{-1}$, $\sigma=3.7$) and $3.44\sigma$ above the long-term mean. It is worth noting that before 2007, CP 11 had only manifested more than 10 days in August during two years (1987 and 1990). From 2008-2011, the persistence of CP 11 August was 9.3
days mo⁻¹, still in excess of 4 days from the long-term mean. Similar to previous results, anomalous increases in CP 11 August frequencies are associated with CP 12 and 13 August decreases. As a result, CP 13 August frequency decreased by more than 2 days in 2007 (occurred only 7 days) and 2008-2011 (µ=7.3 days mo⁻¹) relative to the base period (µ=9.8 days mo⁻¹).

2.3.3 Relationships to Teleconnections

As mentioned in Section 2.1, atmospheric and oceanic teleconnections have connections to the Arctic sea ice extent. In this section, monthly mean teleconnection indices are correlated to the monthly circulation pattern frequencies (meeting the >5% threshold over the melt season) in an attempt to evaluate whether the western Arctic synoptic circulation patterns are related to broader atmospheric and oceanic circulation influences. The correlations between CP frequency and the Arctic Dipole (AD), Arctic Oscillation (AO), Pacific-North American pattern (PNA) and Pacific Decadal Oscillation (PDO) are presented in Table 2.3.

The AD is significantly correlated with many warm-season dominant patterns. CP 11 frequencies are significantly negatively correlated with the AD index during June through August (r = -0.58 – -0.66), with a weaker correlation in September (r = -0.37). This suggests that increases (decreases) in the frequency of CP 11 follow decreases (increases) in the AD. To correspond with the CP11 relationship, the AD is positively correlated with both CP 12 and 13 in June (r = +0.52, +0.61) and July (r = +0.58, +0.35), which indicates that these pattern frequency increases (decreases) are tied to AD increases (decreases) during these months.
The AO and at least one CP are significantly correlated during all melt season
months except August. In particular, CPs 11-14 June are significantly correlated with the
AO during that month. CP 11 frequencies are negatively correlated with the AO during June
(r = -0.46) and July (r = -0.48) indicating that an increase (decrease) in the frequency of CP
11 mirrors a decline (rise) in the AO 1000 hPa pressure centers. CP 13 frequencies are
positively correlated with the AO during June (r = +0.55), July (r = +0.57), and September (r
= +0.64), which indicates that pattern frequency increases (decreases) are related to
increases (decreases) in pressure anomalies affiliated with the teleconnection during
summer.

While there were relatively few statistically significant PDO – CP correlations,
including none after April, correlations between the CPs and the PNA revealed a large
number of significant results. In late winter, CP 6 is significantly correlated with PNA in
March. Despite its summer dominance, CP 10 is well correlated with the PNA in March and
April. The positive correlation between CP 11 August and the August PNA index (r = +0.62)
indicates that an increase (decrease) in the 500 hPa height field potentially increases
(decreases) the frequency of CP 11. The negative correlation between CP 13 August (r = -
0.54) and September (r = -0.39) and those months’ PNA indices also indicates that an
increase (decrease) in the PNA tends to decrease (increase) the frequency of CP 13 over the
western Arctic.

2.4 Discussion

Whereas many studies have evaluated relationships between the summer sea ice
edge and mean summer season circulation over the western Arctic, our study associates
monthly CPs with the sea ice minimum and quantifies the number of days in which the patterns occurred in specified melt season months. Several circulation patterns are statistically connected to the ice extent minimum measured across the 11 longitudes, most significantly the June and August frequencies of CP 11, and to a lesser extent CP 13. The frequency of CP 11, the Beaufort High, during late spring (June) and to a lesser extent mid-summer (August) is especially notable examining its temporal evolution in recent years. While CP 11 July and August frequencies well exceeded their respective 1979-2006 mean values in 2007 then decreased (Figure 2.4), CP 11 has been occurring more frequently in June since 2007. As previously indicated, past studies have highlighted the importance of this SLP feature during summer in conjunction with summer sea ice extent variability (e.g., Rogers, 1978; Drobot and Maslanik, 2003; Ogi et al., 2008). It is important to note that as the thin, first-year ice cover has become more prevalent over time it has also become more susceptible to changes in atmospheric patterns (Stroeve et al., 2012).

The increase in frequency of the Beaufort High since 2007 has coincided with a strengthening of its pressure field in recent years. Moore (2012) found that this anticyclone has strengthened at its climatological core during summers since the late 1990s and regional cyclogenesis has decreased due to tropospheric warming over the western Arctic. Ogi and Wallace (2012) indicated that 925 hPa circulation has become more anticyclonic in summers since 2007, though the 2010 and 2011 wind fields were not quite as strong as those experienced in 2007. The enhanced lower tropospheric flow promotes warm air advection on the western side of the anticyclone off the adjacent northern Alaskan land surface, thermodynamic thinning, and ice export across the Arctic Basin and toward the Fram Strait. Overland et al. (2012) observed a marked change in
June meridional flow from 2007-2012 (versus previous decades) with a positive 700 hPa anomaly and corresponding pronounced MSLP field over the Beaufort Sea. The authors also suggest that these pressure anomalies over the western Arctic (i.e. the North American Ridge) are forcing the Arctic Dipole (AD) to persist and creating blocking patterns that are responsible for warming temperatures in parts of the Arctic, particularly downstream in Greenland. The findings in Section 2.3.3 suggest that an increase in June-August Beaufort High frequency is directly related to a decrease in the AD (i.e. negative values) during those months, which is consistent with the findings of Overland et al. (2012).

Increased persistence of the Beaufort High since 2007 combined with its strengthening during late spring (June) and into summer (July, August) may partially explain why other common synoptic patterns have diminished in recent years and why dramatic end-of-summer ice loss has transpired in the Arctic, especially in the Beaufort and Chukchi Seas. High pressure anomalies (both in terms of strength and persistence) over the Beaufort Sea have triggered a number of observed events that are conducive to substantial sea ice thinning and melt. For instance, the 2007 persistence of the Beaufort High produced widespread clear skies and allowed substantial increases in solar heating of the upper ocean during the summer season (Perovich et al., 2008), which manifested in SST anomalies of 0.5-2.5°C during that summer (Steele et al., 2008).

As stated in the introduction, the Beaufort High pressure signature is a staple in the winter and spring, but it is more centered and well-defined over the eastern Chukchi and Beaufort seas during the latter season as the pressure gradient weakens with changing seasonality and warmer temperatures (Serreze and Barrett, 2011). During summer, an even weaker surface pressure gradient is apparent and the Beaufort High shifts southward
in the Beaufort and Chukchi Seas and includes northern Alaska and broadens to the east to cover the western Archipelago and northern Yukon, Northwest Territories, and Nunavut (Serreze and Barrett, 2011). A roughly similar spatial pattern is resolved by the synoptic technique employed in this study. It is important to remember that the Beaufort High resolved as CP 11 is an amalgamation of similar flow regimes over the course of a year, not just one season even though it is summer-dominant, so its strength will differ from climatological plots of seasonal fields. This pressure pattern is also a “center of action” for the AO and to a lesser extent the AD and PNA during summer (Serreze and Barrett, 2011), which is not surprising given the significant correlations identified in Section 2.3.3 and shown in Table 2.3.

Aside from a strong connection between CP 11 and the ice in particular, there is a noticeable spatial association as the number of CPs connected to western Arctic ice extent increases eastward across the longitudes. The resulting SMLR also shows an increasing amount of ice extent variance is explained by circulation patterns in the eastern versus western half of the region. One reason why the westernmost ice variance (L176-L166) is relatively low \( r^2 \leq 0.49 \) for each of the three longitudes) could be their proximity to the Pacific Ocean via the Bering Strait. Warm water transfer from the Pacific into the western Arctic could be making substantial contributions to ice loss in recent years (Shimada et al., 2006), especially in the Chukchi Sea (Woodgate et al., 2010) where the westernmost ice longitude measurements of this study were derived. Aside from L151, the ice extent explained variance from L161-L126 is 61% or higher (not shown). The higher variance and improved model fit of circulation patterns to the ice extent data eastward toward the
Canadian Arctic Archipelago indicates that circulation plays a larger role in explaining the mid-September ice extent in that part of the domain.

2.5 Conclusions

The synoptic circulation pattern classification performed in this study identified 15 SLP patterns typical over the specified western Arctic domain from 1979-2011, and correlated the frequencies of these patterns with sea ice extent as well as teleconnection indices. Certain late spring/summer patterns, especially CP 11 (which resolved the Beaufort High) and CP 13 are found to be well correlated to the September ice retreat. Both of these CPs are also well correlated with the Arctic Dipole and Arctic Oscillation. The increasing frequency of CP 11 in June especially, but also during July and August, since 2007 has important consequences on the determining the extent of the ice retreat during the melt season, especially given continued thinning of the ice pack. Persistence of this anticyclonic feature corresponds to an enhanced wind field that promotes ice export out of the region while also yielding warm air advection off the land, clearing skies, positive solar radiation anomalies and coincident SST increases in the Beaufort/Chukchi region, which all perpetuate the ice-albedo feedback and stimulate further ice loss.

Aside from the limitations of the synoptic methodology outlined in Yarnal (1993), this study resolves a very concise spatial domain over the western Arctic. In doing so, other circulation patterns that may have influence over the western Arctic ice cover may only be partially resolved or excluded entirely from the analysis. Future work that spatially expands both the ice and SLP fields across a larger Arctic domain may yield some interesting, comparative results over a similar time scale.
Future synoptic methodological studies of circulation over the region will be valuable to identify further connections between ice extent and regional circulation, especially with respect to time. One specific advantage of the synoptic climatological approach, as shown in this study, is that each day of the study period is categorized into one of several patterns which easily allows for the temporal progression of any pattern to be tracked. Arctic climate is expected to continue to warm through the 21st century in conjunction with persistent sea ice decline and anticipated changes in extreme SLP values during summer (Vavrus et al., 2012). Identification of prominent monthly/seasonal circulation patterns and their temporal evolutions moving forward will play a key role in understanding past summer losses and potentially predicting sea ice minima in the future.

Acknowledgments

The authors appreciate the helpful comments provided during the review process that improved this manuscript. The authors would also like to thank Meng-Pai Hung (NOAA/NWS/NCEP/Climate Prediction Center) for his programming assistance, Cameron Lee and Meg Petroski (Kent State University, Department of Geography) for their help plotting the circulation patterns, and James Overland (NOAA/PMEL) and Muyin Wang (JISAO/University of Washington) for sharing the AD data. TJB was supported by NSF Integrated Graduate Education and Research Training Grant DGE 0904560.
References


Figure 2.1. Map of the western Arctic domain. The 11 longitudes from which the latitude of the mid-September sea ice retreat is measured are indicated.
Figure 2.2. The 15 SLP circulation patterns generated from NCEP/NCAR reanalysis.
Figure 2.3. Observed sea ice retreat latitudes (solid) and predicted values (dashed) based on the SMLR equation for (a) L176, (b) L151, and (c) L126. Total explained sea ice extent variance ($r^2$) by the melt season synoptic patterns is indicated.
Figure 2.4. Pie charts indicating the frequencies of circulation patterns 11-13 over June, July, and August for 1979-2006, 2007, and 2008-2011. The number indicates the average occurrence for 1979-2006, 2008-2011, while the 2007 number indicates the occurrence for that year.
### Tables

Table 2.1. Monthly frequencies (%) of the 15 circulation patterns from 1979-2011. The season in which they occur most frequently, winter (Wi), spring (Sp), summer (Su), and autumn (Au) are indicated in the final row.

<table>
<thead>
<tr>
<th>Season</th>
<th>Month</th>
<th>1</th>
<th>2</th>
<th>3</th>
<th>4</th>
<th>5</th>
<th>6</th>
<th>7</th>
<th>8</th>
<th>9</th>
<th>10</th>
<th>11</th>
<th>12</th>
<th>13</th>
<th>14</th>
<th>15</th>
</tr>
</thead>
<tbody>
<tr>
<td>Winter</td>
<td>J</td>
<td>7</td>
<td>13</td>
<td>8</td>
<td>11</td>
<td>7</td>
<td>13</td>
<td>5</td>
<td>6</td>
<td>17</td>
<td>6</td>
<td>&lt;1</td>
<td>&lt;1</td>
<td>&lt;1</td>
<td>1</td>
<td>4</td>
</tr>
<tr>
<td></td>
<td>F</td>
<td>8</td>
<td>13</td>
<td>7</td>
<td>11</td>
<td>7</td>
<td>11</td>
<td>4</td>
<td>9</td>
<td>14</td>
<td>8</td>
<td>&lt;1</td>
<td>1</td>
<td>1</td>
<td>1</td>
<td>4</td>
</tr>
<tr>
<td></td>
<td>M</td>
<td>8</td>
<td>6</td>
<td>7</td>
<td>6</td>
<td>6</td>
<td>10</td>
<td>6</td>
<td>11</td>
<td>8</td>
<td>13</td>
<td>1</td>
<td>2</td>
<td>2</td>
<td>5</td>
<td>7</td>
</tr>
<tr>
<td>Spring</td>
<td>A</td>
<td>6</td>
<td>3</td>
<td>5</td>
<td>4</td>
<td>4</td>
<td>2</td>
<td>9</td>
<td>15</td>
<td>2</td>
<td>15</td>
<td>5</td>
<td>3</td>
<td>5</td>
<td>17</td>
<td>5</td>
</tr>
<tr>
<td></td>
<td>M</td>
<td>2</td>
<td>1</td>
<td>2</td>
<td>1</td>
<td>1</td>
<td>1</td>
<td>4</td>
<td>9</td>
<td>1</td>
<td>13</td>
<td>17</td>
<td>12</td>
<td>11</td>
<td>22</td>
<td>4</td>
</tr>
<tr>
<td></td>
<td>J</td>
<td>1</td>
<td>&lt;1</td>
<td>&lt;1</td>
<td>&lt;1</td>
<td>&lt;1</td>
<td>&lt;1</td>
<td>1</td>
<td>1</td>
<td>1</td>
<td>&lt;1</td>
<td>2</td>
<td>34</td>
<td>25</td>
<td>25</td>
<td>10</td>
</tr>
<tr>
<td>Summer</td>
<td>J</td>
<td>&lt;1</td>
<td>&lt;1</td>
<td>&lt;1</td>
<td>&lt;1</td>
<td>&lt;1</td>
<td>&lt;1</td>
<td>1</td>
<td>1</td>
<td>&lt;1</td>
<td>1</td>
<td>32</td>
<td>35</td>
<td>27</td>
<td>2</td>
<td>1</td>
</tr>
<tr>
<td></td>
<td>A</td>
<td>1</td>
<td>1</td>
<td>1</td>
<td>1</td>
<td>&lt;1</td>
<td>5</td>
<td>1</td>
<td>1</td>
<td>2</td>
<td>20</td>
<td>31</td>
<td>30</td>
<td>3</td>
<td>4</td>
<td></td>
</tr>
<tr>
<td></td>
<td>S</td>
<td>5</td>
<td>4</td>
<td>3</td>
<td>2</td>
<td>6</td>
<td>1</td>
<td>10</td>
<td>3</td>
<td>3</td>
<td>6</td>
<td>12</td>
<td>13</td>
<td>15</td>
<td>6</td>
<td>13</td>
</tr>
<tr>
<td>Autumn</td>
<td>O</td>
<td>13</td>
<td>7</td>
<td>11</td>
<td>3</td>
<td>11</td>
<td>2</td>
<td>8</td>
<td>5</td>
<td>5</td>
<td>8</td>
<td>3</td>
<td>2</td>
<td>4</td>
<td>4</td>
<td>13</td>
</tr>
<tr>
<td></td>
<td>N</td>
<td>16</td>
<td>12</td>
<td>14</td>
<td>4</td>
<td>8</td>
<td>7</td>
<td>10</td>
<td>4</td>
<td>8</td>
<td>7</td>
<td>&lt;1</td>
<td>1</td>
<td>1</td>
<td>1</td>
<td>9</td>
</tr>
<tr>
<td></td>
<td>D</td>
<td>8</td>
<td>13</td>
<td>14</td>
<td>10</td>
<td>9</td>
<td>10</td>
<td>7</td>
<td>5</td>
<td>11</td>
<td>6</td>
<td>&lt;1</td>
<td>0</td>
<td>&lt;1</td>
<td>1</td>
<td>4</td>
</tr>
</tbody>
</table>

Table 2.2. Statistically significant correlations between September ice extent (along 11 longitudes from 176-126°W indicated at the top) and CP frequency from 1979-2011. Italicized values are significant at $\alpha = 0.05$ and bold values are significant at $\alpha = 0.01$. Positive and negative correlation coefficients are indicated by the appropriate sign after the corresponding month. CPs 2, 4, 5, 6, and 9 had no significant correlations during the melt season and were therefore omitted from the table.

<table>
<thead>
<tr>
<th>CP</th>
<th>L176</th>
<th>L171</th>
<th>L166</th>
<th>L161</th>
<th>L156</th>
<th>L151</th>
<th>L146</th>
<th>L141</th>
<th>L136</th>
<th>L131</th>
<th>L126</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>3*</td>
<td>3*</td>
<td>3*</td>
<td>3*</td>
<td>3*</td>
<td>3*</td>
<td>3*</td>
<td>3*</td>
<td>3*</td>
<td>3*</td>
<td>3*</td>
</tr>
<tr>
<td>3</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>3*</td>
<td>3*</td>
<td>3*</td>
<td>3*</td>
<td>3*</td>
<td>3*</td>
<td>3*</td>
</tr>
<tr>
<td>7</td>
<td></td>
<td></td>
<td></td>
<td>4*</td>
<td>4*</td>
<td>4*</td>
<td>4*</td>
<td>4*</td>
<td>4*</td>
<td>4*</td>
<td>4*</td>
</tr>
<tr>
<td>8</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>4*</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>10</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>9*</td>
<td>9*</td>
<td>9*</td>
<td>9*</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>11</td>
<td>6*</td>
<td>6*</td>
<td>6*</td>
<td>6*</td>
<td>6*</td>
<td>6*</td>
<td>6*</td>
<td>6*</td>
<td>6*</td>
<td>6*</td>
<td>6*</td>
</tr>
<tr>
<td>12</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>6*</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>5*</td>
</tr>
<tr>
<td>13</td>
<td>6*</td>
<td>6*</td>
<td>6*</td>
<td>6*</td>
<td>6*</td>
<td>6*</td>
<td>6*</td>
<td>6*</td>
<td>6*</td>
<td>6*</td>
<td>6*</td>
</tr>
<tr>
<td>14</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>5*</td>
<td>5*</td>
<td>5*</td>
<td>5*</td>
<td></td>
</tr>
<tr>
<td>15</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>4*</td>
</tr>
</tbody>
</table>
Table 2.3. Same as Table 2.2, but for monthly teleconnections and CP frequency. CPs 2, 4, and 15 have no significant correlations over the melt season.

<table>
<thead>
<tr>
<th>CP</th>
<th>AD</th>
<th>AO</th>
<th>PDO</th>
<th>PNA</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>4+</td>
<td>4-</td>
<td>4+</td>
<td></td>
</tr>
<tr>
<td>3</td>
<td>4-</td>
<td>4-</td>
<td></td>
<td></td>
</tr>
<tr>
<td>5</td>
<td></td>
<td></td>
<td>9-</td>
<td></td>
</tr>
<tr>
<td>6</td>
<td></td>
<td></td>
<td></td>
<td>3-</td>
</tr>
<tr>
<td>7</td>
<td></td>
<td>3-</td>
<td>3-</td>
<td>8-</td>
</tr>
<tr>
<td>8</td>
<td>3-</td>
<td>4-</td>
<td>5-</td>
<td>4+</td>
</tr>
<tr>
<td>9</td>
<td>3+</td>
<td>3-</td>
<td></td>
<td></td>
</tr>
<tr>
<td>10</td>
<td></td>
<td>3-</td>
<td></td>
<td>3-</td>
</tr>
<tr>
<td>11</td>
<td>3-</td>
<td>3-</td>
<td>3-</td>
<td>8+</td>
</tr>
<tr>
<td>12</td>
<td>5-</td>
<td>6-</td>
<td>6-</td>
<td>5-</td>
</tr>
<tr>
<td>13</td>
<td>6-</td>
<td>7-</td>
<td>4-</td>
<td>8-</td>
</tr>
<tr>
<td>14</td>
<td>4-</td>
<td>5-</td>
<td>3-</td>
<td>3-</td>
</tr>
</tbody>
</table>
CHAPTER 3

REGIONAL ATMOSPHERIC PATTERNS AND THE DELAYED SEA-ICE FREEZE-UP IN THE WESTERN ARCTIC

Abstract

The western Arctic sea ice cover has dramatically changed since the late 1970s, particularly the timing of the autumn freeze-up. While atmospheric dynamic and thermodynamic processes associated with synoptic-scale weather patterns largely impact the onset of regional ice formation, linkages between the subseasonal occurrences of these patterns, across interannual to multidecadal time scales, and the freeze-up are not well understood. This manuscript takes a synoptic climatological atmospheric pattern (AP) classification approach to evaluate the role of warm season-dominant (i.e. May-October) mean sea-level pressure (MSLP) and 1000-500 hPa thickness APs, derived from daily NCEP/NCAR reanalysis fields, on the passive microwave-derived freeze-up dates for the marginal Beaufort/Chukchi Seas and western Arctic Ocean from 1979-2013. Analysis of the respective classifications’ frequencies and their relationships to the freeze-up reveals that approximately one-third of freeze-up variance may be explained by early/middle warm season Beaufort Sea High surface pressure pattern frequency changes. A similar amount of

freeze-up variance is explained by the occurrence of mid-warm season dominant thermal patterns, either earlier or later than their predominant season. Both results suggest that pattern changes may be associated with changing ocean-atmosphere heat exchanges affiliated with lengthened periods of melt conditions.

3.1 Introduction

Satellite observations since the late 1970s have confirmed substantial changes to the Arctic sea-ice cover, including extent declines during all months, highlighted by strong, negative trends during the September minimum (e.g. Simmonds 2015), decreases in thickness and volume (e.g. Stroeve et al. 2014a), and earlier melt and later freeze-up yielding expansion of the melt season (e.g. Stroeve et al. 2014b). While trend analyses on the pan-Arctic scale broadly summarize the rate of change of the ice edge, the regional (i.e. the marginal Arctic seas) ice cover behaviors throughout the annual cycle are not uniform. For instance, Comiso (2012) noted the largest multiyear ice concentration declines are found in the western Arctic Ocean, namely the Beaufort and Chukchi Seas, which is where Xia et al. (2014) also reported the largest ice extent retreat from August to October relative to all other marginal Arctic seas.

There has been an abundance of research tracking summertime western Arctic ice edge variability and the concomitant causal mechanisms (e.g. Rogers 1978; Drobot and Maslanik 2003; Ballinger and Rogers 2013). However, much less research has been conducted regarding the factors connected to Arctic sea ice melt season changes, especially as it relates to the temporal patterns of the autumn ice cover freeze-up. Stroeve et al. (2014b) found statistically significant negative trends in melt onset dates in the Beaufort (-
2.3 days decade\(^{-1}\) and Chukchi (-2.7 days decade\(^{-1}\)) Seas from 1979-2013, with resulting changes in albedo due to ice melt extending the period of significant solar heating of the region’s open ocean. Increased heat input into the ocean during the warm-season delays the freeze-up (Vihma 2014), and partially explains the observed sea surface temperature (SST) and surface air temperature (SAT) increases in the western Arctic region, which act to slow the rate of sea-ice formation as the open ocean gradually cools with diminishing insolation through autumn (e.g. Wood et al. 2013; Stroeve et al. 2014b). Changes in ocean-to-atmosphere heating due to later ice cover formation have been recognized as a substantial culprit for recent autumn lower-tropospheric warming (e.g. Overland 2009; Serreze et al. 2009) and may be connected to the amplified (blocking) nature of large-scale atmospheric flow and persistence of synoptic scale weather systems in the Northern Hemisphere high and middle latitudes in recent years due to a reduction in the high/middle latitude temperature gradient (e.g. Francis and Vavrus 2012; Tang et al. 2014).

There are ultimately several dynamic and thermodynamic mechanisms that can explain some of the interannual to multidecadal variability of the western Arctic sea ice pack throughout different portions of the annual cycle. Wu et al. (2014) noted increasing trends in MSLP magnitudes and associated strengthening summer and autumn anticyclonic surface winds over the western Arctic from 1979-2012. Ogi and Wallace (2012) and Overland et al. (2012) linked early summer 925-hPa anticyclonic wind field anomalies and positive 700-hPa geopotential height anomalies, respectively, over the western Arctic to abrupt summertime Arctic sea-ice losses since 2007. Candlish et al. (2014) explored relationships between wind speed/direction and SAT to western Canadian Arctic sea-ice
concentration and found a statistically significant, negative ($r = -0.88$, $p<0.05$) 1-month lagged correlation between summer/autumn SATs and Beaufort sea ice from 1981-2010, indicating that a rise in air temperature forces the subsequent month’s decline in sea ice. However, wind speed and direction were secondary predictors of ice concentration variability in their analyses.

While deviations and anomalies from mean seasonal atmospheric circulation fields are more commonly used, synoptic climatological methods, particularly atmospheric pattern (AP) classification, have been effectively utilized to holistically understand the spectrum of common, daily weather patterns and their temporal variability related to regional cryospheric change in the high northern latitudes (e.g. Mote 1998; Maslanik et al. 2007; Fettweis et al. 2011, 2013; Bezeau et al. 2014). Most of the western Arctic-centric synoptic-typing schemes have been dynamical in nature, relating the region’s ice cover variables to the seasonal and subseasonal occurrence of mean sea-level pressure (MSLP) patterns. Asplin et al. (2009) associated seasonal mean MSLP AP frequency composites, constructed via k-means cluster analysis, to examine sea-ice motion in the Beaufort Sea from 1979-2006. Ballinger and Sheridan (2014) used a two-step cluster (TSC) analysis technique to generate a MSLP AP catalogue and statistically relate the monthly occurrence of melt season-dominant patterns against September sea-ice extent variability across the western Arctic Ocean from 1979-2011.

While variables such as ice extent and motion have been studied with respect to atmospheric pattern variability, little research has evaluated links between sea ice freeze-up and atmospheric changes. Better understanding of atmospheric conditions influencing the freeze-up onset is critical to improved modeling of high-latitude ocean-atmosphere
heat exchanges and assessment of cryospheric impacts on high/middle latitude weather conditions. To explore these connections, this research takes a synoptic-typing approach to better understand the temporal role that dynamic (i.e. MSLP) and thermodynamic (i.e. 1000-500 hPa thickness) atmospheric patterns are having on changes observed of the western Arctic sea ice freeze-up from 1979-2013. These relationships are analyzed over different time periods to evaluate how the interannual and multidecadal monthly temporal variability of the pressure and thickness patterns relate to the onset of the continuous sea ice formation. This manuscript is subdivided as follows; Section 2 discusses the sea ice and reanalysis datasets used in the study. Section 3 outlines the synoptic climatological methods employed to construct the respective atmospheric pattern classifications. Section 4 describes the results, which include a monthly climatology of the classified patterns, temporal composites that highlight the respective patterns' frequency changes relative to the timing of sea ice formation, and statistical associations between the patterns' occurrences and the freeze-up dates. Section 5 discusses the results, while Section 6 provides brief, concluding remarks on the analyses, and suggests future directions for integrated western Arctic climate and sea ice freeze-up research.

3.2 Data
3.2.1 Reanalysis Fields

Daily-averaged, raw fields of mean sea-level pressure (MSLP) and geopotential-height data used to calculate the 1000-500 hPa thicknesses are obtained from the National Centers for Environmental Prediction/National Center for Atmospheric Research (NCEP/NCAR) first generation reanalysis (Kalnay et al. 1996) for all calendar days over the
1979-2013 period (n = 35 years). The gridded data are analyzed at a 2.5° latitude/longitude horizontal resolution and two domains are utilized; MSLP fields are mapped from 50-90°N and 150°E-100°W (721 gridpoints), while the 1000-500 hPa thickness data are displayed in a smaller domain across 70-85°N and 150°E-100°W (315 gridpoints). Justification for the use of these disparate domains is provided in the Methodology section. NCEP/NCAR MSLP and geopotential-height fields are selected because these model-derived fields closely follow available observational data (Kalnay et al. 1996) and these respective fields yield similar atmospheric flow patterns as other reanalysis products as corroborated by recent western Arctic research (e.g. Serreze and Barrett 2011; Moore 2012). Further, the respective datasets may be used to infer numerous atmospheric conditions, including near-surface wind speed and direction and cloud cover (MSLP), and thermal conditions and associated atmospheric flow of the middle/lower portion of troposphere (1000-500 hPa thickness), respectively.

3.2.2 Sea Ice Data

The western Arctic sea ice freeze-up, referencing the date of continuous freeze conditions observed in a particular pixel, is calculated from passive microwave satellite-derived data for the 1979-2013 period. The satellites employed in constructing the freeze-up record include the Nimbus 7 Scanning Multichannel Microwave Radiometer, and the Defense Meteorological Satellite Program “F family” satellite platforms hosting the Special Sensor Microwave/Imager and the Special Sensor Microwave Imager/Sounder. The algorithm used to process the data is summarized in both Markus et al. (2009) and Stroeve et al. (2014b). Data are obtained at a 25 x 25 km horizontal resolution across the Chukchi
(66.5-77.3°N, 177°E-156.5°W) and Beaufort Seas (71.2-77.3°N, 156.5°W-125°W) in the western Arctic Ocean. The data is quality checked and pixels not registering a date are removed. All pixels that record a freeze-up date are then averaged across the domain of each marginal Arctic sea to construct the annual mean freeze-up date for each year in the study period. The annual mean western Arctic freeze-up date is also constructed and represents the average of all pixel-registered freeze-up dates in the Beaufort and Chukchi Seas for a particular year. Since the Pearson bivariate correlation coefficient between the freeze-up datasets for these marginal seas reveals slight divergence in interannual variability (r = +0.77, p < 0.01), particularly during the mid-2000s (Figure 3.1), associations between the synoptic types and the Beaufort, Chukchi, and areal western Arctic ice cover are all separately explored.

While sea ice freeze-up represents the primary environmental variable of interest in this synoptic climatological study, a time series of melt onset dates (i.e. marking the beginning of continuous melt) is also created from the aforementioned data sources and methods (Figure 3.2). The difference between the respective onset of continuous melt and freeze-up conditions represents the temporal boundaries of the melt season (synonymous with “warm season” throughout this manuscript) and are used to constrain the forthcoming analysis described in Section 3.4.

3.3 Methodology

3.3.1 General Overview of the Two-Step Cluster Analysis

Synoptic climatological classification often involves an atmospheric pattern (AP) methodological approach, which describes the range of typical synoptic scale weather
patterns observed at a particular atmospheric level (Yarnal 1993). This study employs the AP approach to relate MSLP and 1000-500 hPa thickness synoptic types to the freeze-up of the western Arctic Ocean. Since it is desired to analyze the magnitude and seasonality of all patterns with respect to the aforementioned freeze-up, no adjustments were made to the raw reanalysis data prior to initiating the classifications. The synoptic AP classifications, conducted using SPSS Statistics Version 19.0, are initiated by subjecting all gridpoints in each domain to an unrotated S-mode correlation matrix-based principal component analysis (PCA), which reduces the influence of spatial autocorrelation amongst neighboring gridded values and simplifies the dataset while maintaining the bulk of its original explained variance. The retained principal components (PCs) are then entered into a two-step cluster (TSC) analysis in which the first step classifies the initial day's synoptic field (1 January 1979) as the first AP, while the second step initiates a hierarchical clustering procedure that either classifies the second day with the original AP or creates a new AP. This process continues until all days in the dataset are classified into a stipulated number of APs of similar magnitude and spatial orientation that are distinct from other patterns created. More descriptive details of the TSC analysis for synoptic climatological research may be found in Coleman and Rogers (2007).

3.3.2 Construction of the MSLP AP Classification

Statistical techniques employed by Ballinger and Sheridan (2014; abbreviated B&S hereafter in this section) revealed robust correlations between monthly MSLP pattern frequencies and the western Arctic summer sea-ice extent minima, 1979-2011. For use in this research, the addition of 2012 and 2013 MSLP data is necessary. This integration is
carried out using a multi-step procedure similar to that outlined in Lee and Sheridan (2012). The B&S classification details are briefly summarized first. PCA was conducted on all MSLP gridpoints and generated 26 PCs with eigenvalues of at least 1.0, accounting for 98.15% of the original dataset variance. Those retained PCs were then subjected to TSC analysis (with the log-likelihood distance measure selected) and 15 APs were generated.

A stepwise multiple linear regression model is then used to merge the 2012 – 2013 data, where all 26 PCs from the B&S classification are separately regressed onto each of the 721 gridpoints of the complete 1979-2013 MSLP time series. The unstandardized predicted values (i.e. virtual PCs) derived from the regression analysis are saved in order to maintain the spatial properties of the 15 APs previously created. Discriminant function analysis (DFA) is then employed; the 15 APs from the B&S classification are collectively defined as the grouping variable while the virtual PCs represent the independent values for training the new classification with the 2012 and 2013 MSLP data. After these procedures are executed, a new, consistent classification comprised of 15 mean sea-level pressure patterns (abbreviated P1, P2,...P15 in forthcoming sections) is generated that is spatiotemporally similar to B&S in terms of the appearance of the patterns and their co-occurrence (82.94%) during overlapping time periods (i.e. 1979-2011).

3.3.3 Creation of the 1000-500 hPa Thickness Classification

Initially, a 1000-500 hPa thickness classification was created using the same domain as the MSLP classification described previously. However, preliminary analysis revealed no statistically significant (p≤0.05) Pearson bivariate correlations between May-October monthly AP frequencies (i.e. during the western Arctic melt season) and the western Arctic
freeze-up dates. Therefore, several subdomains within the MSLP domain were created and similar monthly frequencies were correlated with the freeze-up data before the final domain was selected. Ultimately, the 70-85°N, 150°E-100°W domain is chosen for forthcoming analyses because this domain yields the largest number of statistically significant AP correlations with the freeze-up date time series.

The 1000-500 hPa thickness classification is produced in a manner similar to the MSLP classification. An unrotated, correlation matrix-based PCA is conducted on all gridpoints in the analysis and yields 7 PCs with eigenvalues of at least 1.0, explaining 98.89% of the original dataset variance. Those 7 PCs are subjected to TSC analysis (with the log-likelihood distance measure designated) and 15 thickness patterns (abbreviated T1, T2,...T15 in forthcoming sections) are generated.

3.4 Results

3.4.1 Frequency Analysis

3.4.1.1 MSLP Classification

The monthly mean frequencies of 15 sea-level pressure patterns are presented in Table 3.1, while all synoptic pattern maps are displayed in Figure 3.3. Pattern frequencies are roughly distributed throughout the year and aside from P9 during January (i.e. P9 January) and P1 November, which both occur 17% of days in those months, no other pressure patterns occur at least 13% days month$^{-1}$ outside of warm-season months May-September. During these months, P11-14 (Figure 3.4) occur much more frequently relative to the other 11 patterns during these months or over the remaining months of the year (i.e. November-April). P11 occurs 37% (33%) of all June (July) days. P12 and P13 have similar
seasonality, occurring most often during August (35% and 31% of days respectively), followed by slightly lesser frequencies in July and June. P14 is early warm season-dominant and occurs most often during May (21% of days).

All remaining analyses are constrained to the warm season-dominant P11-14 for several reasons. The mean absolute correlation coefficients of the western Arctic freeze-up dates and all AP frequencies aggregated by month reveal a general peak during May-October (see Table 3.2). This window of months typically frames the western Arctic melt season (i.e. sea ice melt onset until freeze-up) as melt onset dates have commenced from late May (e.g. Day of Year = 148; Figure 3.2), while freeze-up has often transpired during October (e.g. Day of Year$_{1979-2013}$ ≈ 277; Figure 3.1). Further, of the 8 statistically significant correlation coefficients between May-October MSLP pattern frequencies and the western Arctic freeze-up record, 5 involve P11-14 (Figure 3.4; more statistical details are offered in Section 3.4.3).

3.4.1.2 1000-500 hPa Thickness Classification

Monthly mean frequencies of the thickness patterns are displayed in Table 3.3 and the complete thickness pattern classification is shown in Figure 3.5. These 15 patterns display a more distinctive seasonality, compared to the MSLP patterns, with several patterns predominantly occurring during the cold (i.e. November-April) or warm seasons. For instance T1 January and February occur 17% and 19% of all days in those respective months, while the pattern’s occurrence during June-September is negligible. However, no cold season-dominant patterns occur more than 19% of days month$^{-1}$. With the exception of T6, whose maximum frequency of occurrence is 15% of July days, all other warm season-
dominant patterns (T6, T9, and T13-15; Figure 3.6) occur on at least 1 month during 22% of days or more.

Similar to the MSLP analyses, remaining 1000-500 hPa thickness pattern analyses focus on warm season-dominant T6, T9, and T13-15. In particular, mean absolute correlation coefficients among the monthly thickness pattern frequencies and western Arctic freeze-up dates are typically highest during May-October (Table 3.2), matching the western Arctic melt season temporal domain. Statistically significant AP correlations with the western Arctic sea ice freeze-up dates also largely transpire during these months, with 6 of the 11 total significant correlations observed with T6, T9, T13 and T15 (more statistical details may also be found in Section 3.4.3).

3.4.2 Assessing AP Temporal Differences through Compositing

3.4.2.1 MSLP

As seen in Figure 3.1, western Arctic freeze-up has been dramatically delayed since the late 1990s relative to the previous years dating back to 1979. Warm-season dominant pressure patterns P11-14 are sequentially composited in seven equal quintiles (i.e. Early, 1979-1985; Early-Middle, 1986-1992; Middle, 1993-1999; Middle-Late, 2000-2006; Late, 2007-2013) to evaluate if their monthly occurrences in successive periods throughout the time series may be associated with changes in the date of the freeze-up (Figure 3.7). Interquintile comparisons reveal substantial changes between periods, particularly more recently. P11, which resolves a Beaufort Sea High (BSH) pattern, has exhibited a notable increase in frequency during the Late versus Middle-Late periods for June (+4.14 days year\(^{-1}\)) and July (+4.43 days year\(^{-1}\)), while P11 August frequencies also occur much more
frequently during the later freeze-up years since 2000 relative to the previous three periods. To assess intramonthly changes across the warm season, the cumulative frequency of P11 classified days (May-October) within each quintile, for instance, is calculated and a 14-day centered moving average is applied to smooth the time series and assess the frequency variability between the Early, Middle, and Late periods (Figure 3.8). During June, July, and August (JJA) P11 occurs much more frequently during the last seven years of the study period relative to those same months during the Middle and Early portions of the record, which may be associated with the substantial summertime retreat of the region’s ice edge (Ballinger and Sheridan 2014) and may potentially contribute to the recent persistence shown in the freeze-up record.

As shown in Figure 3.7, the increase in the number of P11 classified days during the core of the warm season (JJA) is offset by the sharp decrease of P12 during June and July and more subtle declines of P13 for those same months during the Late period. P13 August frequencies also drop considerably from the Middle to Late periods (-4.57 days year\(^{-1}\)), which compensate for P11 frequency increases during this month.

P14, a broader, stronger high pressure pattern compared to P11, has also occurred much more often during May from the Middle-Late to Late quintiles (+3.71 days year\(^{-1}\)), and then occurs infrequently from June-October relative to the other three main MSLP patterns (see Figure 3.7). This particular pressure pattern may represent a more spatially-extensive BSH pattern and relate to the mid-tropospheric anticyclonic circulation that has been observed (with increasing frequency) stretching across the western Arctic to Greenland over a similar time period (Bezeau et al., 2014).
3.4.2.2 1000-500 hPa Thickness

The thickness pattern frequencies are aggregated similarly to the pressure patterns discussed previously, and substantial frequency differences observed amongst the different periods are discussed here (and seen in Figure 3.9). During the Early, Early-Middle, and Middle-Late periods T6 May occurs relatively infrequently compared to the Middle and Late Periods and these frequency spikes are roughly coincident with periods of later freeze-up. T9 also exhibits an increase from Middle and Middle-Late to Late periods during May (+1.14 days year\(^{-1}\)) and September (+1.57 days year\(^{-1}\)). These frequency increases from the Early and Middle periods to the Late period are largely manifested in the latter halves of those respective months (Figure 3.10).

Increased occurrence of the T13 thermal ridge over the western Arctic Ocean from the Middle-Late to Late periods for June (+2.43 days year\(^{-1}\)), July (+2.57 days year\(^{-1}\)), and August (+3.43 days year\(^{-1}\)) may collectively explain some of the persistence of the freeze dates over these periods (e.g. Middle-Late Day of Year (DOY) Freeze-up\(_{\mu} = 279\) versus Late DOY Freeze-up\(_{\mu} = 295\)). The T13 increases are balanced by declines of T14 during the aforementioned months (Figure 3.9). The late warm-season increases of T15 September and October from the Middle-Late to Late periods, represented by a strong thickness gradient over the eastern half of the domain, may also offer physical explanation for the recent expansion of the western Arctic freeze-up period.
3.4.3 Statistical Analyses

3.4.3.1 MSLP

The multiyear/short-term (i.e. 7-year) frequency changes of atmospheric patterns assessed in Section 3.4.2 are likely playing a role in freeze-up change as the ice edge continues to retreat poleward through summer, but the low-frequency (i.e. multidecadal) behaviors of the APs are also likely fueling freeze date persistence over time. To test this notion, Pearson bivariate correlations and stepwise multiple linear regression analyses involving the frequencies of the warm season MSLP patterns and Beaufort, Chukchi, and western Arctic freeze-up time series are performed in order to statistically evaluate individual and collective pressure pattern associations with the variability of the annual freeze dates.

As shown in Table 3.4, there are several patterns whose monthly frequencies are statistically connected to the timing of the freeze-up in the Beaufort, Chukchi, and western Arctic and these significant associations often occur in May, June, August, and September. Increases in P14 May, reflected by prevalence of high pressure stretching across most of the northernmost portion of the region, yields a tendency for delayed freeze-up across all three ice domains. Moreover, P11 June is positively correlated with freeze dates of the Beaufort (r = +0.34, p = 0.044), Chukchi (r = +0.42, p = 0.012), and western Arctic (r = +0.42, p = 0.013) domains. P11 August frequencies also exhibit similar multidecadal covariability with the Chukchi Sea (r = +0.40, p = 0.018) and western Arctic (r = +0.36, p = 0.036), suggesting that early and middle warm season increases of BSH patterns also facilitate later freeze dates.
The monthly pressure patterns significantly correlated with the respective ice data, displayed in Table 3.4, are used as predictors against those freeze-up time series in three separate stepwise multiple linear regression models. The freeze-up explained variance (i.e. adjusted $r^2$) by the significant frequencies is similar across the western Arctic Ocean and Beaufort and Chukchi Seas. P11 June and P13 August account for 30% (31%) of the cumulative freeze-up variance across the western Arctic (Chukchi Sea) and offer some statistical explanation between early/middle warm season-dominant regional high pressure patterns’ variability and delayed ice formation in the referenced domains. MSLP pattern predictors for the Beaufort Sea freeze-up include P12 June and P14 May (cumulative $r^2 = 28$%), and excludes the P11 BSH pattern that explains some freeze-date variance of the larger sea-ice domains.

3.4.3.2 1000-500 hPa Thickness

The significant correlations between the thickness pattern frequencies and the sea ice freeze-up share similar subseasonal associations with the MSLP results described in the previous section, and there is a clear seasonal signal derived from early and middle/late warm season thickness pattern occurrences (Table 3.4). May occurrences of T6 and T9 are positively correlated with freeze-up across most of the domains, indicating that the increase of these patterns during the early season from 2007-2013 potentially prompt later freezes through earlier melting of the ice pack.

T13 exhibits a warm ridge over the Beaufort Sea and its August and September frequencies are positively correlated with the sea ice in the Chukchi and greater western Arctic Ocean. In particular, T13 August frequencies (Figure 3.9) dramatically increase with
freeze-date persistence during recent years, while T13 September frequencies have generally been much higher since the Middle period (i.e. 1993-1999). T15 August, and its associated weak thickness gradient across the domain, shows a negative linear relationship with the ice freeze-up in the Chukchi Sea and western Arctic Ocean, likely since its decline has been associated with the noted T13 frequency increase. In contrast, when T15 occurs more frequently during October, as has been observed over the Late period, freeze dates tend to be later. The correlation coefficient between this pattern’s monthly frequency and the freeze-up is highest in the Chukchi Sea (r = +0.52, p = 0.002) and western Arctic (r = +0.47, p = 0.004) relative to all thickness patterns explored via correlation analysis.

Stepwise multiple linear regression modeling using the significant, monthly thickness patterns linked to the freeze-up dates as predictors reveals similar variance contributions for the ice domains as found in the MSLP regression analyses (Table 3.4). Nearly one-third of the western Arctic freeze-up is collectively represented by T15 October and T9 May frequencies (r² = 29%). Similarly, T15 October is the initial predictor of Chukchi freeze-up, and combines with T13 September to explain 33% of the variance. The Beaufort Sea ice freeze dates are only weakly explained by T9 May (r² = 12%). While few thickness patterns explain roughly one-third or less of the respective freeze date variability since 1979, regression fits suggest the cumulative importance of early and late-warm season thermal patterns to explaining some freeze-up variability across much of the western Arctic region.
3.4.3.3 MSLP and 1000-500 hPa Thickness Contributions

The monthly frequency variability of MSLP patterns or 1000-500 thickness patterns discussed previously are linked to greater ice loss at certain times through the warm season than others. For instance, P11 June frequency increases coinciding with the maximization of solar insolation during that month aid the summer ice pack melt and its dynamical movement. However, the collective influence of MSLP and 1000-500 hPa thickness pattern occurrences during the melt season may also influence the autumn ice formation over time.

In order to evaluate this collective contribution by both disparate synoptic catalogues, all statistically significant patterns shown in Table 3.4 for each domain are entered into stepwise multiple linear regression models to evaluate their ability to hindcast the respective regions’ freeze-up dates. The amount of collective variance explained by the predictors is consistent amongst the 3 domains (total $r^2 = 38 - 41\%$). P12 June, P14 May, and T14 May collectively account for 39% of the Beaufort Sea freeze-up variability. Chukchi Sea freeze-up can be explained by three predictors (total $r^2 = 41\%$), which include T15 October, T9 May, and P13 September, while the western Arctic cover (total $r^2 = 38\%$) can also partially be explained by those same predictors.

Despite the omission of P11, particularly during June, as a statistically significant predictor in the collective regression model runs, its recent quintile frequency increases (i.e. 2007-2013, see Figure 3) are substantial relative to earlier time periods, and the multidecadal statistical links to the freeze-up (Table 3.4) are relatively strong. Recent AP classifications have similarly corroborated the summertime increase of the BSH pattern (e.g. Ballinger et al. 2014; Belleflamme et al. 2015), and a statistical connection between its
early warm season occurrence and the observed September Beaufort and Chukchi sea ice extent minima (Ballinger and Sheridan 2014).

3.5 Discussion

3.5.1 MSLP

BSH pressure patterns have increased across the early/middle warm season months in recent years. These substantial increases in the BSH frequencies (e.g. P11 JJA or P14 May) create a dynamic environment ideal for sea-ice movement with a synoptic setup promoting large-scale subsidence and southerly flow over the Beaufort and Chukchi Seas. Melt conditions would be particularly optimal when these pressure patterns persist during the early and middle portions of the warm season as clear skies allow diabatic heating of the sea ice surface and shortwave radiation absorption into the open ocean adjacent to the ice edge, which enhances the ice-albedo feedback.

Decreases in P12 and P13 JJA frequencies compensate for some of the observed BSH (i.e. P11) increases during those months. P12 represents a relatively weak pressure gradient across the western Arctic and it is no surprise that a decrease in its frequency coincides with a later freeze-up (negative correlation coefficient) since this pattern’s general frequency decreases over time (as witnessed in Figure 3.7) are paralleled by P11 increases. This sharp change in occurrence of low-pressure patterns (e.g. P13) to high pressure patterns in the western Arctic (e.g. P11 and P14) may partly explain the freeze-up delays into later autumn. Moreover, correlation coefficients between P13 August and September frequencies and all three domains’ freeze-up are also negative; northerly, cold air advection fueled by the low-pressure cell located in the central Arctic Ocean would yield
conditions more conducive to earlier ice-cover formation. P11 has increased during August since 2000 and offsets some observed declines in P13 frequencies during that month, which may also partially explain why freeze-up dates are transpiring later in the year over recent years.

3.5.2 1000-500 hPa thickness

T6 May and T9 May/September thickness pattern frequency increases in recent years are likely playing a role on sea-ice conditions. Thicknesses of ~5300-5450 m are prevalent across the domain of both patterns. Coastal areas, including northern Siberia (T6) and northern Alaska and Siberia (T9), are encompassed by the 5450-m contour, with lower thickness values stretching poleward. This particular thickness value represents a proxy for near-freezing surface air temperatures, and its prevalence over the ocean may aid changes in freeze-up either through enhanced sea-ice melting early in the warm season (i.e. during May), delayed ice formation during the latter portion of the warm season (i.e. during September), or a combination of both mechanisms. It should also be noted that the relatively meridional orientation of both thickness patterns’ isopachs yields a thermal wind regime that advects warmer, southerly air off the adjacent landmasses toward the open ocean and adjacent ice pack thereby promoting the thermodynamic melting and slowing of the subsequent formation of ice cover described.

Increases in T13 August and September frequencies are statistically linked with Chukchi and western Arctic freeze dates. Near-freezing air temperatures inferred from the thermal ridge of 5450-m thickness values in the south-central portion of the domain and the associated off-continent, southwesterly flow may act to postpone sea-ice formation.
The inverse association between T15 August frequencies and the freeze dates may largely be the result of recent, coincident increases in T13 (see Figure 3.9) that exacerbate sea ice melt and ultimately delay freeze onset.

The relationship between T15 and freeze-up timing appears to change as the warm season progresses. Whereas decreases (increases) of T15 August are tied to later (earlier) western Arctic and Chukchi Sea freeze-ups, the correlation coefficient flips sign and strengthens during October and direct associations can be made between T15 occurrences and the aforementioned domains’ freeze dates. T15 October’s increase from pre-1999 to the years since 2007 is roughly by a factor of two (Figure 3.9). This substantial frequency change may be driven by long-term changes to the ice cover. From 1979-2012, the October open water fraction (inversely related to sea-ice concentration trends) increased 44% and 46% in the Beaufort and Chukchi Seas, respectively; this generates increases in oceanic sensible heat release to the atmosphere and lower tropospheric warming as evidenced by strong, positive linear trends in Barrow, Alaska’s surface air temperatures over the period (Wendler et al. 2014). This could manifest in T15’s prolonged occurrence during October relative to previous years. Hence the ice cover changes through October, including the observed freeze-date persistence in recent years, may be expanding the seasonality of the typical warm-season weather patterns in the western Arctic and altering thermodynamic associations with the sea ice cover over time.

3.6 Conclusions

This study utilizes two synoptic climatological classifications, derived from MSLP and 1000-500 hPa thickness data, to explore the observed changes in sea ice freeze-up
dates in the western Arctic. Both classifications show clear changes in their respective patterns over the 7-year windows, particularly in the Late period (i.e. 2007-2013) period for the patterns exhibiting statistically significant associations with the ice dates. Moreover, the frequency variability of the early/middle warm season MSLP patterns and early/late warm season thickness patterns are largely related to the timing of the seasonal sea-ice cover formation during autumn (i.e. September-November). Correlation analyses and multiple linear regression models applied to assess the cumulative impacts of both classifications' pattern frequencies on freeze-up timing suggest that the influence of thermal patterns may be more important than the wind-driven forcing, especially for those patterns occurring late in the warm season with high thickness values and associated air temperatures near the freezing point extending over the ocean.

There are undoubtedly a multitude of atmospheric mechanisms that contribute to the interannual and multidecadal degradation and thinning of the western Arctic summer ice cover (e.g. Ballinger and Rogers 2013), leading to freeze-up persistence. Factors including Northern Hemisphere temperature increases and warm season atmospheric teleconnection phases involving the Arctic Dipole, Arctic Oscillation, and Pacific-North American Pattern have also been shown to account for a significant portion of the region's summer sea-ice extent variance over the last two decades (e.g. Ballinger and Rogers 2014) and may further elucidate some of the residual regression model freeze-up variance presented in this manuscript. Substantial thinning of the ice pack during the last decade, particularly in the western Arctic (e.g. Kwok and Cunningham 2010), suggests that anomalous frequencies of atmospheric patterns may play an increasingly significant role in massive seasonal sea-ice losses and therefore these relationships should continue to be
monitored, particularly as linkages between freeze-up and weather pattern changes become elucidated through continued cryospheric research.

This limited-domain synoptic typing study has been shown to be effective at capturing typical, synoptic scale patterns and their recent frequency increases (i.e. the BSH) that have also been resolved through different classification approaches, alternative reanalysis products, and larger domains (e.g. Bezeau et al. 2014; Belleflamme et al. 2015). Future work exploring various atmospheric pressure levels' geopotential height and wind field associations to the freeze-up may be helpful, for instance, at further explaining the residual variance of the linear best fit models presented in this manuscript. A spatial expansion of both the sea ice and atmospheric domains may also be useful to assess and compare the impacts of high latitude atmospheric pattern variability, and related dynamic and thermodynamic processes, to freeze-up changes observed in other marginal Arctic seas.

Acknowledgments

The authors would like to thank Jeffrey Miller (NASA Cryospheric Sciences Laboratory/Wyle, Inc.), Stephen Howell (Environment Canada), Mary Haley (NCAR/CISL), and Cameron Lee (Kent State University, Department of Geography) for their assistance with aspects of data acquisition/processing and figure creation.

The NCEP/NCAR reanalysis data is obtained from the NOAA/ESRL Physical Science Division (http://www.esrl.noaa.gov/psd/data/reanalysis/), while the freeze-up data is acquired from the NASA Cryosphere Science Research Portal.
References


Figure 3.1. Freeze-up dates for the Beaufort Sea, Chukchi Sea, and Western Arctic Ocean for the 1979-2013 study period.
Figure 3.2. Same as Figure 3.1, but for melt onset dates.
Figure 3.3. Full catalogue of the 15 MSLP patterns.
Figure 3.4. Warm season-dominant MSLP patterns P11-14. See Figure 3.3 for the full catalogue.
Figure 3.5. Full catalogue of the 15 1000-500 hPa thickness patterns.
Figure 3.6. Warm season-dominant 1000-500 hPa thickness patterns T6, T9, and T13-15.

See Figure 3.5 for the full catalogue.
Figure 3.7. Quintiles of P11-14 frequencies by warm season months, May-October. Each quintile represents a consecutive 7-year period across the entire time series (1979-2013).
Figure 3.8. Time series of P11 across the warm season months for the Early, Middle, and Late periods described in Figure 3.7. The frequency represents the total number of days classified for each day (May-October) within each period. A 14-day centered moving average is then applied to smooth the time series across all months.
Figure 3.9. Quintiles of T6, T9, and T13-15 frequencies by warm season months, May-October. Each quintile represents a consecutive 7-year period across the entire time series (1979-2013).
Figure 3.10. Time series of T9 across the warm season months for the Early, Middle, and Late periods described in Figure 3.9. The frequency represents the total number of days classified for each day (May-October) within each period. A 14-day centered moving average is then applied to smooth the time series across all months.
Tables

Table 3.1. Mean sea-level pressure pattern (P) classification monthly mean frequency of occurrence (%) from 1979-2013.

<table>
<thead>
<tr>
<th>Month/P</th>
<th>1</th>
<th>2</th>
<th>3</th>
<th>4</th>
<th>5</th>
<th>6</th>
<th>7</th>
<th>8</th>
<th>9</th>
<th>10</th>
<th>11</th>
<th>12</th>
<th>13</th>
<th>14</th>
<th>15</th>
</tr>
</thead>
<tbody>
<tr>
<td>January</td>
<td>6</td>
<td>12</td>
<td>8</td>
<td>9</td>
<td>8</td>
<td>13</td>
<td>5</td>
<td>6</td>
<td>17</td>
<td>8</td>
<td>1</td>
<td>&lt;1</td>
<td>1</td>
<td>1</td>
<td>4</td>
</tr>
<tr>
<td>February</td>
<td>8</td>
<td>12</td>
<td>10</td>
<td>6</td>
<td>10</td>
<td>5</td>
<td>11</td>
<td>13</td>
<td>10</td>
<td>1</td>
<td>1</td>
<td>2</td>
<td>2</td>
<td>2</td>
<td>4</td>
</tr>
<tr>
<td>March</td>
<td>9</td>
<td>5</td>
<td>7</td>
<td>5</td>
<td>6</td>
<td>9</td>
<td>6</td>
<td>11</td>
<td>8</td>
<td>14</td>
<td>3</td>
<td>2</td>
<td>4</td>
<td>5</td>
<td>6</td>
</tr>
<tr>
<td>April</td>
<td>7</td>
<td>3</td>
<td>3</td>
<td>3</td>
<td>4</td>
<td>2</td>
<td>9</td>
<td>14</td>
<td>2</td>
<td>13</td>
<td>8</td>
<td>5</td>
<td>7</td>
<td>18</td>
<td>4</td>
</tr>
<tr>
<td>May</td>
<td>3</td>
<td>1</td>
<td>1</td>
<td>&lt;1</td>
<td>1</td>
<td>1</td>
<td>4</td>
<td>7</td>
<td>&lt;1</td>
<td>10</td>
<td>22</td>
<td>15</td>
<td>12</td>
<td>21</td>
<td>3</td>
</tr>
<tr>
<td>June</td>
<td>&lt;1</td>
<td>&lt;1</td>
<td>&lt;1</td>
<td>&lt;1</td>
<td>&lt;1</td>
<td>&lt;1</td>
<td>1</td>
<td>1</td>
<td>&lt;1</td>
<td>1</td>
<td>37</td>
<td>24</td>
<td>26</td>
<td>9</td>
<td>&lt;1</td>
</tr>
<tr>
<td>July</td>
<td>&lt;1</td>
<td>&lt;1</td>
<td>&lt;1</td>
<td>&lt;1</td>
<td>&lt;1</td>
<td>&lt;1</td>
<td>1</td>
<td>1</td>
<td>&lt;1</td>
<td>&lt;1</td>
<td>33</td>
<td>34</td>
<td>29</td>
<td>1</td>
<td>1</td>
</tr>
<tr>
<td>August</td>
<td>1</td>
<td>1</td>
<td>1</td>
<td>&lt;1</td>
<td>1</td>
<td>&lt;1</td>
<td>1</td>
<td>4</td>
<td>&lt;1</td>
<td>&lt;1</td>
<td>1</td>
<td>20</td>
<td>35</td>
<td>31</td>
<td>2</td>
</tr>
<tr>
<td>September</td>
<td>5</td>
<td>4</td>
<td>3</td>
<td>1</td>
<td>5</td>
<td>&lt;1</td>
<td>9</td>
<td>2</td>
<td>2</td>
<td>4</td>
<td>13</td>
<td>17</td>
<td>17</td>
<td>5</td>
<td>12</td>
</tr>
<tr>
<td>October</td>
<td>12</td>
<td>8</td>
<td>9</td>
<td>2</td>
<td>9</td>
<td>2</td>
<td>10</td>
<td>4</td>
<td>6</td>
<td>7</td>
<td>5</td>
<td>3</td>
<td>5</td>
<td>4</td>
<td>12</td>
</tr>
<tr>
<td>November</td>
<td>17</td>
<td>12</td>
<td>12</td>
<td>3</td>
<td>8</td>
<td>7</td>
<td>10</td>
<td>4</td>
<td>8</td>
<td>6</td>
<td>1</td>
<td>2</td>
<td>2</td>
<td>1</td>
<td>8</td>
</tr>
<tr>
<td>December</td>
<td>6</td>
<td>11</td>
<td>13</td>
<td>8</td>
<td>9</td>
<td>11</td>
<td>8</td>
<td>6</td>
<td>12</td>
<td>7</td>
<td>1</td>
<td>1</td>
<td>1</td>
<td>1</td>
<td>6</td>
</tr>
</tbody>
</table>
Table 3.2. Mean absolute correlation coefficients (|r|) between the western Arctic freeze-up dates and the MSLP and 1000-500 hPa thickness AP frequencies by month. The total number of patterns which occur at least one day during half of the respective months of the time series (e.g. ≥18 Januaries) are used to calculate the |r| and listed accordingly (#P/T\textsubscript{patterns}).

<table>
<thead>
<tr>
<th>APs by Month</th>
<th>MSLP</th>
<th>1000-500 hPa thickness</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>J\text{anuary}</td>
<td>0.15</td>
<td>11</td>
</tr>
<tr>
<td>F\text{ebruary}</td>
<td>0.14</td>
<td>11</td>
</tr>
<tr>
<td>M\text{arch}</td>
<td>0.20</td>
<td>14</td>
</tr>
<tr>
<td>A\text{pril}</td>
<td>0.13</td>
<td>10</td>
</tr>
<tr>
<td>M\text{ay}</td>
<td>0.21</td>
<td>7</td>
</tr>
<tr>
<td>J\text{une}</td>
<td>0.33</td>
<td>4</td>
</tr>
<tr>
<td>J\text{uly}</td>
<td>0.09</td>
<td>3</td>
</tr>
<tr>
<td>A\text{ugust}</td>
<td>0.27</td>
<td>4</td>
</tr>
<tr>
<td>S\text{eptember}</td>
<td>0.23</td>
<td>9</td>
</tr>
<tr>
<td>O\text{ctober}</td>
<td>0.14</td>
<td>12</td>
</tr>
<tr>
<td>N\text{o\l\v e\m\b\e\r\n\e}</td>
<td>0.13</td>
<td>11</td>
</tr>
<tr>
<td>D\text{e\c\e\m\b\e\r\n\e}</td>
<td>0.09</td>
<td>10</td>
</tr>
</tbody>
</table>
Table 3.3. Same as Table 3.1, but for the 1000-500 hPa thickness pattern (T) classification.

<table>
<thead>
<tr>
<th>Month/T</th>
<th>1</th>
<th>2</th>
<th>3</th>
<th>4</th>
<th>5</th>
<th>6</th>
<th>7</th>
<th>8</th>
<th>9</th>
<th>10</th>
<th>11</th>
<th>12</th>
<th>13</th>
<th>14</th>
<th>15</th>
</tr>
</thead>
<tbody>
<tr>
<td>January</td>
<td>17</td>
<td>7</td>
<td>6</td>
<td>14</td>
<td>7</td>
<td>1</td>
<td>10</td>
<td>8</td>
<td>&lt;1</td>
<td>11</td>
<td>8</td>
<td>11</td>
<td>&lt;1</td>
<td>&lt;1</td>
<td>&lt;1</td>
</tr>
<tr>
<td>February</td>
<td>19</td>
<td>7</td>
<td>6</td>
<td>14</td>
<td>6</td>
<td>&lt;1</td>
<td>9</td>
<td>8</td>
<td>&lt;1</td>
<td>9</td>
<td>8</td>
<td>13</td>
<td>&lt;1</td>
<td>&lt;1</td>
<td>&lt;1</td>
</tr>
<tr>
<td>March</td>
<td>13</td>
<td>7</td>
<td>7</td>
<td>13</td>
<td>10</td>
<td>&lt;1</td>
<td>9</td>
<td>10</td>
<td>&lt;1</td>
<td>9</td>
<td>8</td>
<td>12</td>
<td>&lt;1</td>
<td>&lt;1</td>
<td>&lt;1</td>
</tr>
<tr>
<td>April</td>
<td>13</td>
<td>11</td>
<td>10</td>
<td>10</td>
<td>8</td>
<td>1</td>
<td>9</td>
<td>10</td>
<td>&lt;1</td>
<td>6</td>
<td>10</td>
<td>7</td>
<td>1</td>
<td>1</td>
<td>2</td>
</tr>
<tr>
<td>May</td>
<td>4</td>
<td>6</td>
<td>3</td>
<td>5</td>
<td>6</td>
<td>10</td>
<td>5</td>
<td>2</td>
<td>9</td>
<td>3</td>
<td>4</td>
<td>4</td>
<td>16</td>
<td>10</td>
<td>14</td>
</tr>
<tr>
<td>June</td>
<td>&lt;1</td>
<td>1</td>
<td>&lt;1</td>
<td>&lt;1</td>
<td>1</td>
<td>12</td>
<td>&lt;1</td>
<td>&lt;1</td>
<td>22</td>
<td>&lt;1</td>
<td>1</td>
<td>&lt;1</td>
<td>27</td>
<td>18</td>
<td>18</td>
</tr>
<tr>
<td>July</td>
<td>&lt;1</td>
<td>&lt;1</td>
<td>&lt;1</td>
<td>&lt;1</td>
<td>&lt;1</td>
<td>15</td>
<td>&lt;1</td>
<td>&lt;1</td>
<td>18</td>
<td>&lt;1</td>
<td>&lt;1</td>
<td>&lt;1</td>
<td>25</td>
<td>23</td>
<td>18</td>
</tr>
<tr>
<td>August</td>
<td>&lt;1</td>
<td>1</td>
<td>&lt;1</td>
<td>&lt;1</td>
<td>1</td>
<td>14</td>
<td>&lt;1</td>
<td>&lt;1</td>
<td>14</td>
<td>&lt;1</td>
<td>&lt;1</td>
<td>&lt;1</td>
<td>23</td>
<td>24</td>
<td>22</td>
</tr>
<tr>
<td>September</td>
<td>&lt;1</td>
<td>5</td>
<td>4</td>
<td>2</td>
<td>5</td>
<td>11</td>
<td>1</td>
<td>1</td>
<td>10</td>
<td>1</td>
<td>4</td>
<td>1</td>
<td>20</td>
<td>16</td>
<td>19</td>
</tr>
<tr>
<td>October</td>
<td>11</td>
<td>11</td>
<td>10</td>
<td>8</td>
<td>14</td>
<td>2</td>
<td>6</td>
<td>4</td>
<td>1</td>
<td>6</td>
<td>12</td>
<td>6</td>
<td>3</td>
<td>1</td>
<td>5</td>
</tr>
<tr>
<td>November</td>
<td>12</td>
<td>7</td>
<td>10</td>
<td>13</td>
<td>13</td>
<td>&lt;1</td>
<td>7</td>
<td>8</td>
<td>&lt;1</td>
<td>10</td>
<td>10</td>
<td>11</td>
<td>&lt;1</td>
<td>&lt;1</td>
<td>&lt;1</td>
</tr>
<tr>
<td>December</td>
<td>15</td>
<td>7</td>
<td>5</td>
<td>15</td>
<td>8</td>
<td>&lt;1</td>
<td>11</td>
<td>8</td>
<td>&lt;1</td>
<td>9</td>
<td>12</td>
<td>10</td>
<td>&lt;1</td>
<td>&lt;1</td>
<td>&lt;1</td>
</tr>
</tbody>
</table>
Table 3.4. Statistically significant Pearson bivariate correlations between the respective warm season patterns and sea ice freeze-up dates for the Beaufort, Chukchi, and Western Arctic (p ≤ 0.05). Numbers in columns 2-4 represent calendar months and the superscripts indicate the sign of the correlation coefficient.

<table>
<thead>
<tr>
<th>Pattern Catalogue</th>
<th>Beaufort</th>
<th>Chukchi</th>
<th>Western Arctic</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Pressure (P)</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>11</td>
<td>6*</td>
<td>6*</td>
<td>6*</td>
</tr>
<tr>
<td></td>
<td>8*</td>
<td></td>
<td>8*</td>
</tr>
<tr>
<td>12</td>
<td>6*</td>
<td></td>
<td></td>
</tr>
<tr>
<td>13</td>
<td>8*</td>
<td>6*</td>
<td>8*</td>
</tr>
<tr>
<td></td>
<td>9*</td>
<td>8*</td>
<td>9*</td>
</tr>
<tr>
<td>14</td>
<td>5*</td>
<td>5*</td>
<td>5*</td>
</tr>
<tr>
<td><strong>Thickness (T)</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>6</td>
<td>5*</td>
<td></td>
<td>5*</td>
</tr>
<tr>
<td>9</td>
<td>5*</td>
<td>5*</td>
<td>5*</td>
</tr>
<tr>
<td>13</td>
<td>8*</td>
<td></td>
<td>8*</td>
</tr>
<tr>
<td></td>
<td>9*</td>
<td>9*</td>
<td></td>
</tr>
<tr>
<td>15</td>
<td>8*</td>
<td>10*</td>
<td>8*</td>
</tr>
<tr>
<td></td>
<td></td>
<td>10*</td>
<td></td>
</tr>
</tbody>
</table>
CHAPTER 4

SEA ICE IMPACTS ON POLAR SURFACE WEATHER TYPES IN THE NORTH AMERICAN ARCTIC

Abstract

The Arctic cryosphere has rapidly changed, especially in terms of diminishing summer and autumn sea ice cover, during the last three decades. During that period, increases in lower tropospheric air temperature and moisture, especially at high latitudes, have also been observed across autumn (October – December) and winter (January – March) months. However, the spatiotemporal effects of the observed ice losses on North American weather conditions are not well understood. This study employs a synoptic climatological weather typing scheme known as the Spatial Synoptic Classification (SSC) to holistically evaluate temperature and moisture changes across terrestrial North American Arctic (NAA) coincident with western Arctic sea ice freeze-up dates from 1979-2013. Monthly climatologies and linear trends of autumn/winter-dominant SSC weather types, Moist Polar (MP) and Dry Polar (DP), are assessed and statistically linked to the sea ice freeze dates. Results suggest that the MP weather types are increasing, at the expense of DP types, across much of the domain. These frequency changes are positively correlated with the timing of the western Arctic freeze, especially across much of Alaska and the Yukon Territory. Comparisons between late-and early-freeze years reveal that MP
frequencies occur much more often during late-freeze versus early-freeze years, especially at the highest northern latitude stations. The results suggest strong, multidecadal connections between western Arctic sea ice freeze onset and monthly weather type variability across much of the NAA. Further research is needed to explore the spatial extent of regional sea ice impacts on surface weather conditions throughout the Northern Hemisphere middle latitudes.

4.1 Introduction

Arctic amplification, the augmented warming of air temperatures in the Northern high latitudes relative to the entire Northern Hemisphere, has been observed for decades (e.g. Serreze and Barry, 2011). Increases in lower tropospheric temperature and moisture have been noted in observational and reanalysis datasets across much of the North American Arctic (NAA) during the autumn (October – December) and winter (January – March) seasons since 1979 (e.g. Bekryaev et al., 2010; Serreze et al., 2012; Cohen et al., 2014). One of the factors linked to the low-level thermal and humidity changes has been massive Arctic sea ice loss during summer, and associated autumn freeze-up delays, which have allowed increases in Arctic Ocean-to-atmosphere heat fluxes during the cold season months (e.g. Serreze et al., 2009). These anomalous heat fluxes may be decreasing the high/middle latitude temperature gradient, promoting wavier jet stream patterns, and potentially promoting anomalous spells of extreme weather across North America in recent years (e.g. Kumar et al., 2010; Overland et al., 2011; Francis and Vavrus, 2012; Cassano et al., 2014).
While research suggests a changing boreal cryosphere is impacting large-scale weather and climate, the geographic extent of Arctic change impacts across the Northern Hemisphere is generally vague. Budikova (2009) noted that surface air temperatures (SATs) tend to be locally amplified by sea ice retreat, while temperatures tend to cool rapidly with increasing distance from the ice cover throughout the middle latitudes. Overland et al. (2011) used reanalysis fields to show the localized/regionalized NAA December lower tropospheric warming anomalies along with cooling over parts of the North American middle latitudes during high ice loss years, 2008-2010. In contrast, Francis et al. (2009) found low summer sea ice extent observations linked to higher winter temperatures and atmospheric moisture across North America (November – January). As sea ice is expected to continue to diminish through at least the middle of the 21st century (e.g. Overland et al., 2014), future impacts of these losses on weather patterns is also anticipated. Screen et al. (2015) contend, through climate model simulations, that a continued decline in summer sea ice may decrease the likelihood of cold extremes across North America, especially if greenhouse gas emissions continue to rise in the future.

Studies linking cryospheric changes to weather and climate responses often do so through analysis of disparate observational and model-derived meteorological and climatological datasets, such as SAT and precipitation (i.e. rain and snow measurements). Synoptic climatological weather type classifications, which holistically classify daily weather conditions based on variables such as temperature and moisture that are measured at high frequencies at land-surface meteorological stations (Yarnal, 1993), provide a unique opportunity to summarize thermal and moisture conditions potentially related to Arctic change. One particular weather type classification, the Spatial Synoptic
Classification (SSC; Sheridan, 2002), has been used in continental-scale cryospheric studies to assess the impacts of snow cover on air mass (i.e. weather type) frequencies (Leathers et al., 2002) and to better understand surface weather conditions associated with multidecadal snow cover changes across much of North America (Dyer and Mote, 2007).

To the author's knowledge, cryospheric research has yet to explore potential associations between the frequencies of SSC types and sea ice freeze-up timing over extended periods or anomalous years, particularly at the regional scale. Studies analyzing links between freeze onset and land surface weather characteristics are crucial to not only advance the understanding of autumn and winter sea ice-atmosphere-land surface interactions, but also improve short and long-term mesoscale and synoptic-scale forecasts. In order to better understand the potential role that changing Arctic sea ice conditions may be having on high latitude autumn and winter surface weather, this study assesses the dominant NAA monthly SSC type frequencies (October – March) and their relationships to western Arctic freeze-up variability, 1979-2013. Some of the largest positive trends (i.e. delays) of autumn freeze-up dates in the Arctic marginal seas since 1979 have been measured in the Beaufort Sea and Chukchi Sea (Stroeve et al. 2014), which together comprise the westernmost portion of the Arctic Ocean and border some northern coasts of NAA lands. As such, the sea ice domain used in this study is constrained to the western Arctic Ocean to better understand how regional freeze-up variability relates to NAA-dominant weather types. The manuscript is composed as follows: Section 4.2 summarizes the data and methods. Section 4.3 describes the results, including temporal evaluations of the dominant SSC types’ frequencies, their multidecadal links to western Arctic freeze-up variability, and their observed changes in the most extreme freeze-up years. Section 4.4
contextualizes the relationships between SSC types and the freeze-up data, and Section 4.5 offers conclusions and potential paths for future research.

4.2 Data and Methods

4.2.1 Freeze-up Data

The western Arctic sea ice freeze date time series, 1979-2013 (n = 35 years), is calculated from passive microwave satellite-derived data. This specific dataset is composed of continuous freeze-up measurements garnered from the Nimbus 7 Scanning Multichannel Microwave Radiometer, and the Defense Meteorological Satellite Program “F family” satellite platforms hosting the Special Sensor Microwave/Imager and the Special Sensor Microwave Imager/Sounder. The details of the algorithm used to process the freeze data are described in Markus et al. (2009).

Freeze-up data are obtained at a 25 x 25 km horizontal resolution across both the Chukchi (66.5-77.3°N, 177°E-156.5°W) and Beaufort Seas (71.2°N-77.3°N, 156.5°W-125°W), which together geographically comprise the western Arctic Ocean. Pixels not registering a freeze date or that intersect a portion of the adjacent land surface are removed from the analysis. All pixels that register a freeze-up date are then averaged across the respective seas to create mean freeze-up date for each calendar year. The Beaufort and Chukchi freeze date time series are then averaged for each year to produce a cohesive set of western Arctic freeze-up dates, which vary between mid-September and late October throughout the study period. Freeze-up anomalies (compared to the 1981-2010 mean; Figure 4.1) are then constructed to highlight the surface weather responses
that occur during the extreme late and early freeze years (see Results subsections 4.3.3 and 4.3.4).

4.2.2 The Spatial Synoptic Classification

There are two primary methods inherent to synoptic climatological studies, weather typing and map (i.e. circulation/atmospheric) pattern classification (Yarnal, 1993). Weather typing approaches evaluate surface weather conditions at specific locations and the SSC, in particular, is exclusively used in this study. This weather typing scheme is employed to holistically assess near-surface, in situ thermal and moisture observations across Arctic North America that may be related to western Arctic sea ice freeze-up variability. The SSC takes into account 6-hourly SAT, dew point, sea-level pressure, wind speed and direction, and cloud cover as measured at first-order weather stations to classify each day, spanning the 1979-2013 study period, into one of seven different weather types according to their temperature and moisture characteristics. These weather types include Dry Moderate, Dry Polar, Dry Tropical, Moist Moderate, Moist Polar, Moist Tropical, and Transition. As described in more detail in the next section, the primary weather types classified in the NAA throughout most of the year, especially during the autumn and winter months, are Dry Polar (DP) and Moist Polar (MP), and these two types represent the focus of forthcoming analyses. In total, 27 stations (see Figure 4.2 and Table 4.1) north of 60°N in North America, including two newly created SSC stations in Greenland (i.e. THU and SFJ), encompass the NAA domain used for this analysis.

These weather stations have data records complete up to at least 1979-2010...2013 (i.e. data terminating between 2010 and 2013), which are evaluated separately and against
the sea ice freeze-up dataset. Analyses of the October – December SSC trends and climatologies span 1979-2013, while the January – March trends are calculated over the 1980-2013 period. Further, October – December SSC-freeze correlations of the respective datasets overlap similar years, 1979-2010...2013, while the January – March correlations take into account the offset between the previous calendar year’s freeze date, 2009...2012, and the SSC data concluding the following year. Composite and individual extreme year analyses (Sections 4.3.3 and 4.3.4) are limited to 22 NAA stations and exclude YRB, YFB, YZF, BET, and YXY because their records terminate prior to 2013, following the final extreme late freeze year of 2012 (more details in Section 4.3.3).

4.2.3 Analysis of SSC and Freeze-up Datasets

Associations between monthly frequencies of DP and MP weather types and the freeze-up dates are conducted using correlation analysis and compositing procedures and subsequently mapped and described throughout the forthcoming section. The statistical strength of the relationships between the respective freeze-up data and individual month DP and MP frequency time series (from each NAA station) is assessed using Pearson bivariate correlations (Section 4.3.2). Composites of weather type frequencies by freeze-up date extremes, defined as anomalous dates lying outside ±1 standard deviation (σ) from the 1981-2010 climatological freeze date, is also carried out to see if there is a preponderance of types that occur in late versus early years across the domain (Section 4.3.3). More details of the analyses are described at the outset of each subsection of results presented in Section 4.3.
4.3 Results

4.3.1 MP/DP Climatology and Trends

Monthly climatologies of the NAA MP and DP weather types for October – March are presented in Figures 4.3 and 4.4. MP weather types comprise ≥50% of classified days during October for many of the western and southernmost stations in the domain (Figure 4.3). Pockets of western and southeastern Alaska, southwest Yukon Territory, and the southern Northwest Territories show ~50% of days classified as MP in November and December as well. The spatial frequency patterns of MP occurrences are relatively similar for January-March with the majority of the Canadian Archipelago experiencing MP conditions on ~20% of all days with increasing frequencies of up to 40% extending toward the western and easternmost edges of the domain toward Alaska and the west coast of Greenland.

DP classified days are more common relative to the MP type as shown by the monthly climatologies presented in Figure 4.4. Except during October, DP occurs between ~50-80% of all classified days for the majority of the stations across the core of the NAA. Peripheral areas, including much of continental and the extreme southern and eastern stations show 20-40% of days classified as the DP type, analogous to MP frequency values for similar months of the year.

As shown by the MP and DP climatologies described above, these polar weather types are the most common types across the Arctic environment from October-March. With evidence of Arctic climate change over the last several decades, it would be expected that the SSC weather types would not remain static in their frequency of occurrence across time, but similarly change as near-surface temperature and moisture conditions vary. In
order to evaluate frequency changes of the classified types, a linear regression is performed and the trends (indicated by the slopes of the respective regression lines representing changes in MP/DP classified days year\(^{-1}\)) are displayed in Figures 4.5 and 4.6.

Positive trends in MP classified days (0.10 – 0.40 days year\(^{-1}\)) are apparent in October and November for much of the NAA domain (Figure 4.5). These trends are strongest during October and November across parts of northern Alaska, the Archipelago, and northern Greenland. Stations sampled in Alaska indicate that MP is occurring more frequently during all months except March. MP trends appear to weaken from December to January and flip sign across the Archipelago, signifying an overall decrease in MP classified days over time. However, greater, positive MP trends are observed for February, especially in the core of continental Alaska, before again weakening or changing to slightly negative (-0.10 days year\(^{-1}\)) for March across much of the NAA.

Observed decreases in DP trends compensate for much of the increase in MP frequencies since 1979. October and November display negative DP trends (-0.10 - -0.30 days year\(^{-1}\)) around the northern portion of the NAA in particular, while this type seems to increase in November across central and southern Alaska (Figure 4.6). The negative DP trend largely weakens from November to December before becoming negligible to weakly positive during January. Negative February trends in DP frequencies (-0.20 - -0.40 days year\(^{-1}\)) are witnessed over much of Alaska before switching sign to positive trends of comparable magnitudes across both Alaska and western Greenland during March.
4.3.2 Long-term MP/DP Associations with Freeze-up

Multidecadal interactions between MP/DP weather type variability and western Arctic Ocean freeze-up dates are evaluated to better understand how these relationships co-vary during October – March throughout the study period. Initially, the Pearson bivariate correlations were calculated separately between the Chukchi Sea, Beaufort Sea, and western Arctic Ocean freeze-up dates and the frequencies of the SSC types to see if results significantly vary depending on the ice domain selected. An example, using November MP frequencies versus the different freeze dates (1979-2013), is shown in Figure 4.7. Little differences in the strength of the correlations are observed, while the sign of the coefficients is consistent with each domain throughout the NAA. Therefore, the western Arctic domain, comprising both marginal Arctic seas, is chosen for analyses in the remaining results sections.

Correlation coefficients between the monthly MP frequencies and western Arctic freeze-up are presented in Figure 4.8. The highest, positive correlations consistently appear during October – December across much of northern and western areas of the NAA. Specifically, the most positive coefficients are found over northern Alaska during November (r ~ +0.50) suggesting a strong, localized relationship between the freeze onset of the adjacent sea ice cover and MP classified days at Barrow in particular. The correlations generally weaken to the south and east from Barrow toward the Archipelago where coefficients are negligible to weakly negative (r = 0.00 – -0.20) during all months except October when the coefficients are slightly positive. The majority of Alaskan and Yukon Territory stations exhibit positive correlations, while western Greenland (i.e.
Kangerlussuaq) MP frequencies versus the freeze-ups are negative ($r = -0.10 - -0.40$) during the months analyzed.

DP weather types are predominantly anti-correlated with the freeze-up dates (Figure 4.9). Negative DP correlations are strongest during October and November for the northernmost NAA stations ($r \approx -0.40 - -0.50$) and weaken south, especially across the Archipelago. Central and southern Alaska show weakly positive November, January, and March DP frequency-freeze correlations and these areas largely flip sign during October, December and February. Positive DP-ice correlations at western Greenland, of similar magnitude to the MP-ice associations previously mentioned, are observed for October – March and suggest the surface environment is becoming cooler and drier in this area of the NAA in contrast to the warming and moistening conditions witnessed throughout much of high latitude North America.

**4.3.3 Late versus Early Freeze Year Composites**

Multidecadal relationships between the freeze date time series and MP/DP frequencies are further investigated by the exploring differences in these weather types during extreme freeze-up years. These extreme years are characterized as the freeze date outliers that are $\pm 1\sigma$ from the 1981-2010 mean freeze-up date ($\pm 12.88$ days versus the 30-year climatology). As such, there are seven years exceeding $+1\sigma$ (1998, 2006-2008, 2010-2012; hereafter *late* years) and seven years less than $-1\sigma$ (1980, 1981, 1983, 1985, 1988, 1992, 1994; hereafter *early* years). The respective weather type frequencies in the late years and early years are averaged and subtracted (late – early) to assess differences in MP and DP classified days during these extreme year composites.
Figure 4.10 shows the mean differences in MP classified days between the late and early freeze years. Positive, mean differences in MP frequencies are most substantial (+2 – 8 days year\(^{-1}\)) and spatially consistent across the Alaskan and Yukon Territory stations relative to the rest of the NAA during most months. The positive differences are strongest during November, peaking at +8 MP days year\(^{-1}\) on average during the late versus early years at Barrow. Northern stretches of the Northwest and Yukon Territories and north/central Alaska also exhibit MP changes of +2 – 6 days year\(^{-1}\) on average during November with the number of MP classified days increasing toward the western Arctic Ocean. Little to no change in MP days is observed during March except that 8 fewer MP days are found at Kangerlussuaq during the late versus early years, suggesting colder, drier conditions are more typical at that station following late western Arctic freeze-ups.

Differences in DP classified days between the late and early freeze years generally contrast those of the MP results previously described. For instance, DP classified days occur much less frequently in late versus early freeze years for northern Alaska (i.e. Barrow) and the southeastern Archipelago during October and especially November (Figure 4.11). There is a substantial gradient of DP differences observed during November stretching from northern Alaska (-12 days year\(^{-1}\)) to the central part of the state (+4 days year\(^{-1}\)), suggesting an abrupt transition of higher to lower temperature and moisture conditions with increasing distance from the ocean during this month that is not realized as strongly in the other autumn and winter months. Much of interior Alaska and Yukon and Northwest Territories show a positive difference in DP classified days in November as well as March (+2 – 10 days year\(^{-1}\)), indicating cooler, drier weather types prevail in these continental locales following later freeze of the adjacent ocean surface.
4.3.4 Highlighting Individual Extreme Freeze Years

As shown in Figure 4.1, there are several years that exceed ±1σ from the freeze date climatology. However, three late (2007, 2011, 2012) and three early freeze years (1983, 1985, and 1994) are clearly most extreme in their respective freeze-up timing compared to the remaining years of the time series. Next, comparisons assess the extent to which the MP and DP weather type frequencies are anomalous during the months of these extreme early and late freeze years (October – December), and those months immediately following (January – March), with respect to mean DP/MP monthly occurrences from the 1981-2010 base period.

Overall, the monthly MP and DP frequencies observed across the NAA stations are not spatially consistent across the most extreme early and late freeze years (Figures 12-23). However, the late freeze year of 2007 and the early freeze year of 1983 do exhibit some regional similarities in weather type anomalies, albeit of opposite sign. In 2007, MP frequencies are consistently at or above the climatological mean (≥0 days) across most of Alaska for all months except February (see Figures 4.12, 4.14, 4.16, 4.18, 4.20, 4.22). The largest anomalies manifest during November (Figure 4.14) across central/northern Alaska (+4-12 days versus the climatology), increasing poleward toward the western Arctic Ocean. Pockets of positive MP anomalies (~+4 days) are also found during October (central Alaska and most of the Yukon Territory), December (central/northern Alaska and northern Yukon Territory), January (southern Alaska and Yukon Territory), and March (western Alaska). Negative spatial patterns in MP classified days are witnessed for similar geographic areas (as seen in 2007) in October and December during the early freeze year of 1983.
Negative DP anomalies in 2007 (Figures 4.13, 4.15, 4.17, 4.19, 4.21, and 4.23) compensate for positive MP anomalies during select months, especially November across north/central Alaska (~8 days; Figure 4.15). However, the DP gradient in classified days is not as steep as the one observed in the MP plot (Figure 4.14) toward the western Arctic Ocean indicating that frequency changes of other SSC weather types are likely offsetting the MP increases. Positive DP anomalies are relatively pronounced across central and northern Alaska in most months, except November, surrounding the 1983 extreme early freeze-up as well. DP occurs at least 4-16 days more than average in central/southern Alaska and most of the Yukon Territory during December. Positive anomalies over similar magnitude (+4-16 days) are also witnessed across central/northern Alaska and the Yukon Territory during the month of February.

4.4 Discussion

In this analysis, NAA MP and DP weather type frequencies are evaluated for their climatological and linear trend characteristics and subsequently associated with the western Arctic freeze-up time series and the most extreme freeze-up years over the 1979-2013 study period. The results indicate that October and November MP (DP) weather type frequencies are increasing (decreasing) most substantially, compared to the other months analyzed at most NAA stations in Figure 4.5 (Figure 4.6), especially across the northernmost portion of the domain. Observed MP weather type increases at coastal and near-coastal stations during the autumn months suggest that coincident persistence in regional Beaufort and Chukchi sea ice cover are creating a warmer, more humid climate across the North Slope (e.g. Wendler et al., 2010, 2014)
Noteworthy Pearson correlations ($r > +0.20$) between the sea-ice freeze dates and the monthly MP frequencies extend beyond stations adjacent to the ocean during some months (Figure 4.8); this link suggests that the delays in ice cover formation impact not only coastal locales, but also synoptic-scale weather conditions at continental sites. For most months, the positive correlation coefficients are the highest for the western half of the domain, encompassing both Alaska and the Yukon Territory. Peak values in correlation coefficients spatially vary by month within this portion of the NAA, but are strongest overall during November across northern Alaska (i.e. Barrow; Figure 4.8).

Observed increases in moisture and temperature across the western NAA, possibly driven by sea ice cover change, have been documented in recent research and likely explain increases in autumn MP classified days over time. Positive trends in SATs (1981-2012) have been observed across much of Alaska during October and November with the most significant warming taking place over the North Slope and the warming signal generally diminishing toward the Gulf of Alaska (Bieniek et al., 2014). Serreze et al. (2012) document positive trends in precipitable water from the surface to the 500 hPa layer above Barrow during October – December, 1979-2010. Wendler et al. (2014) noted annual precipitation increases and October and November SAT changes at Barrow, 1979-2012, in excess of +6°C during both months coincident with substantial decreases in Beaufort and Chukchi sea ice concentration. This research also finds positive SAT changes of >2°C for December and February during that time span. These low frequency thermal and moisture increases further implicate a longer autumn open ocean period (i.e. extended ice-free conditions) as a factor potentially responsible for local/regional climatic changes coinciding with the timing of the freeze-up and the months thereafter. This sea ice-surface weather relationship
diminishes across space likely in part because the station-observed warming signals wane toward the Alaskan interior. This phenomenon explains the weakening of the spatial pattern in MP-freeze correlation coefficients during the months following the typical freeze-up in October or November.

Composites of extreme late versus early freeze-up years (Section 4.3.3) highlight discrepancies in MP/DP occurrences as MP classified days tend to occur much more often during late freeze years across much of the domain (except the Canadian Arctic Archipelago) compared to when the sea ice freezes anomalously early. The results indicate that during the seven most extreme years the increases in MP frequencies, especially across Alaska, persist through February, suggesting that the autumn freeze-up delays are leading to warmer and more humid surface conditions in the winter months. However, with the exception of the late freeze year of 2007, analysis of the individual most extreme freeze years (Section 4.3.4) does not reveal an especially conclusive and consistent spatial pattern of MP or DP frequency anomalies throughout the NAA. The interannual impacts of sea ice on SSC occurrences are spatially inconsistent across the land surface, perhaps due to year-to-year regional variability in factors such as wind speed and direction (e.g. Stegall and Zhang, 2012; Baule and Shulski, 2014). In contrast, spatial patterns revealed by the MP/DP composites, based on early (late) freeze years that occur before (after) the Arctic amplification signal became apparent from natural variability in the mid-1990s (Francis and Vavrus, 2015), may be more strongly influenced by a long-term regional warming trend versus spatiotemporal variability in large-scale atmospheric circulation patterns (Cassano et al., 2011). This suggests that sea ice influences the likelihood of anomalous
DP/MP occurrences, but other climatic factors may also play a role in the interannual variability of these weather types.

Previous research has investigated the role of atmospheric circulation on monthly SSC frequency variability across some stations lying within the NAA (e.g. Sheridan, 2003). While an assessment of the various atmospheric mechanisms linking sea ice loss to terrestrial climate anomalies is beyond the scope of this paper, research has suggested that atmospheric circulation anomalies resulting in persistent, extreme temperatures (i.e. cold outbreaks and heat events) over North America and Eurasia are intricately tied to boreal high-latitude snow and sea ice losses of the last three plus decades (e.g. Francis et al., 2009; Tang et al., 2013; Mori et al. 2014). Further investigations relating interannual freeze-onset to persistent circulation patterns and associated surface weather conditions across the northern high and middle latitudes is necessary to further elucidate these complex, interconnected sea ice-climate relationships, especially within the context of this current era of massive cryospheric change.

4.5 Conclusions

This study explores monthly SSC MP and DP frequencies, 1979-2013, for October – March across the NAA and their relationships to western Arctic freeze-up. Linear trends indicate that MP frequencies have increased substantially over the period, especially over Alaska and the Yukon Territory during October – February. These monthly MP frequency changes are well-correlated with the date of the western Arctic freeze-up over these areas, while DP occurrences and ice cover are anti-correlated at similar magnitudes across similar portions of the domain, suggesting that local/regional moisture and temperature increases
(yielding MP increases) may be forced by the ongoing delays in the timing of the freeze-up. Composite analyses of extreme freeze-up years further associate MP frequencies with ice cover as MP classified days are much more pervasive, on average, during November and adjacent months for the later freeze-up years versus those where the freeze occurs earlier (in autumn). Overall, the individual most extreme freeze years show inconsistent spatial responses to the ice cover, though 2007 does show some consistent spatial patterns of positive MP anomalies in most of the months.

This manuscript explores links between western Arctic sea ice cover freeze-up and DP and MP weather type variability across high latitude, terrestrial North America. Arctic amplification, driven in part by the losses of sea ice cover in summer and autumn, appears to regionally manifest by forcing MP increases across much of the NAA, especially across Alaska and the Yukon Territory. It is acknowledged that the freeze-up patterns of the Beaufort and Chukchi Seas are not uniform and therefore certain portions of each sea may be open longer than the annual mean freeze date derived in this paper. Differences in freeze-up amongst the Arctic marginal seas may be influencing some of the variability in the weather type frequencies, especially for the stations closer to maritime environments. Future work using similar analytical techniques may look to expand/alter the freeze dates dataset by comparing other Arctic marginal seas against the SSC types or examine associations between sea ice freeze-up patterns at different latitudes and the frequencies of the SSC types to better understand spatial patterns in regional ocean-atmosphere interactions. Beyond the regional scale (i.e. NAA SSC domain), future work incorporating the frequency changes of more SSC stations across North America is necessary to
understand the geographic boundaries and potential equatorward extent of sea ice-
weather type relationships.

Acknowledgments

The authors would like to thank Julien Nicolas (Polar Meteorology Group, Byrd
Polar and Climate Research Center, The Ohio State University) for assistance creating
Figure 2 of this manuscript.

The SSC data is obtained from the Spatial Synoptic Classification homepage
(http://sheridan.geog.kent.edu/ssc.html), while the freeze-up data is acquired from the
dataset for 2012-2013 are supplied by Jeffrey Miller (NASA Cryospheric Sciences
Laboratory/Wyle, Inc.).
References


Figure 4.1. Time series of western Arctic freeze-up date anomalies, 1979-2013. The 7 most extreme years, with respect to ±1 standard deviation (sigma) from the 1981-2010 freeze-up mean (Day of Year = 276), are utilized in composite and individual freeze year analyses described in Sections 4.3.3 and 4.3.4.
Figure 4.2. Map of 27 first-order weather stations poleward of 60°N with Spatial Synoptic Classification (SSC) weather types in North America (including Greenland). Table 4.1 provides details on each location. The green polygon outlines the approximate western Arctic sea ice freeze-up domain used in the analysis.
Figure 4.3. Monthly mean frequencies of MP classified days for October-March following data records listed in Table 4.1.
Figure 4.4. Monthly mean frequencies of DP classified days for October-March following data records listed in Table 4.1.
Figure 4.5. Linear trend (slope) of MP classified days year\(^{-1}\) for October-March following data records listed in Table 4.1.
Figure 4.6. Linear trend (slope) of DP classified days year\(^{-1}\) for October-March following data records listed in Table 4.1.
Figure 4.7. Pearson bivariate correlations between November MP weather type occurrences and the Beaufort Sea, Chukchi Sea, and western Arctic freeze-up dates.
Figure 4.8. Pearson bivariate correlations between October – March MP weather type occurrences and the freeze date of western Arctic sea ice. Correlation coefficients of $-0.34 > r > +0.34$ are statistically significant.
Figure 4.9. Pearson bivariate correlations between October – March DP weather type occurrences and the freeze date of western Arctic sea ice. Correlation coefficients of $-0.34 > r > +0.34$ are statistically significant.
Figure 4.10. Monthly differences in MP classified days observed for the 7 latest minus the 7 earliest freeze-up years. The difference between the average of these two 7-year periods is portrayed in each plot.
Figure 4.11. Monthly differences in DP classified days observed for the 7 latest minus the 7 earliest freeze-up years. The difference between the average of these two 7-year periods is portrayed in each plot.
Figure 4.12. MP October frequency anomalies versus the 1981-2010 MP October mean during the extreme early (E) freeze years (1983, 1985, and 1994) and late (L) freeze years (2007, 2011, and 2012).
Figure 4.13. DP October frequency anomalies versus the 1981-2010 DP October mean during the extreme early (E) freeze years (1983, 1985, and 1994) and late (L) freeze years (2007, 2011, and 2012).
Figure 4.14. MP November frequency anomalies versus the 1981-2010 MP November mean during the extreme early (E) freeze years (1983, 1985, and 1994) and late (L) freeze years (2007, 2011, and 2012).
Figure 4.15. DP November frequency anomalies versus the 1981-2010 DP November mean during the extreme early (E) freeze years (1983, 1985, and 1994) and late (L) freeze years (2007, 2011, and 2012).
Figure 4.16. MP December frequency anomalies versus the 1981-2010 MP December mean during the extreme early (E) freeze years (1983, 1985, and 1994) and late (L) freeze years (2007, 2011, and 2012).
Figure 4.17. DP December frequency anomalies versus the 1981-2010 DP December mean during the extreme early (E) freeze years (1983, 1985, and 1994) and late (L) freeze years (2007, 2011, and 2012).
Figure 4.18. MP January frequency anomalies versus the 1981-2010 MP January mean during Januaries immediately following the extreme early (E) freeze years (i.e. 1984, 1986, and 1995) and late (L) freeze years (i.e. 2008, 2012, and 2013).
Figure 4.19. DP January frequency anomalies versus the 1981-2010 DP January mean during Januaries immediately following the extreme early (E) freeze years (i.e. 1984, 1986, and 1995) and late (L) freeze years (i.e. 2008, 2012, and 2013).
Figure 4.20. MP February frequency anomalies versus the 1981-2010 MP February mean during Februaries immediately following the extreme early (E) freeze years (i.e. 1984, 1986, and 1995) and late (L) freeze years (i.e. 2008, 2012, and 2013).
Figure 4.21. DP February frequency anomalies versus the 1981-2010 DP February mean during Februaries immediately following the extreme early (E) freeze years (i.e. 1984, 1986, and 1995) and late (L) freeze years (i.e. 2008, 2012, and 2013).
Figure 4.22. MP March frequency anomalies versus the 1981-2010 MP March mean during March months immediately following the extreme early (E) freeze years (i.e. 1984, 1986, and 1995) and late (L) freeze years (i.e. 2008, 2012, and 2013).
Figure 4.23. DP March frequency anomalies versus the 1981-2010 DP March mean during March months immediately following the extreme early (E) freeze years (i.e. 1984, 1986, and 1995) and late (L) freeze years (i.e. 2008, 2012, and 2013).
Table 4.1. North American Arctic weather stations with SSC weather types corresponding to Figure 4.2 for Alaska (AK), USA, Yukon Territory (YT), Northwest Territories (NWT), and Nunavut (NU), Canada, and Greenland (GL). Stations 3 (YRB), 17 (YFB), 20 (YZF), 24 (BET), and 25 (YXY) are not included in the composite analyses because their records terminate prior to 2013.

<table>
<thead>
<tr>
<th>Number</th>
<th>Station Name</th>
<th>Abbreviation</th>
<th>Lat (°N)</th>
<th>Lon (°W)</th>
<th>Data Record</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Eureka, NU</td>
<td>WEU</td>
<td>79.98</td>
<td>85.13</td>
<td>1979-2013</td>
</tr>
<tr>
<td>2</td>
<td>Thule, GL</td>
<td>THU</td>
<td>76.53</td>
<td>68.70</td>
<td>1979-2013</td>
</tr>
<tr>
<td>3</td>
<td>Resolute, NU</td>
<td>YRB</td>
<td>74.72</td>
<td>94.97</td>
<td>1979-2010</td>
</tr>
<tr>
<td>4</td>
<td>Barrow, AK</td>
<td>BRW</td>
<td>71.29</td>
<td>156.77</td>
<td>1979-2013</td>
</tr>
<tr>
<td>5</td>
<td>Clyde River, NU</td>
<td>YCY</td>
<td>70.48</td>
<td>68.52</td>
<td>1979-2013</td>
</tr>
<tr>
<td>6</td>
<td>Cambridge Bay, NU</td>
<td>YCB</td>
<td>69.10</td>
<td>105.13</td>
<td>1979-2013</td>
</tr>
<tr>
<td>7</td>
<td>Hall Beach, NU</td>
<td>YUX</td>
<td>68.78</td>
<td>81.25</td>
<td>1979-2013</td>
</tr>
<tr>
<td>8</td>
<td>Inuvik, NWT</td>
<td>YEV</td>
<td>68.31</td>
<td>133.50</td>
<td>1979-2013</td>
</tr>
<tr>
<td>9</td>
<td>Kangerlussuaq, GL</td>
<td>SFJ</td>
<td>67.01</td>
<td>50.72</td>
<td>1979-2013</td>
</tr>
<tr>
<td>10</td>
<td>Bettles, AK</td>
<td>BTT</td>
<td>66.92</td>
<td>151.53</td>
<td>1979-2013</td>
</tr>
<tr>
<td>11</td>
<td>Kotzebue, AK</td>
<td>OTZ</td>
<td>66.89</td>
<td>162.60</td>
<td>1979-2013</td>
</tr>
<tr>
<td>12</td>
<td>Norman Wells, NWT</td>
<td>YVQ</td>
<td>65.28</td>
<td>126.79</td>
<td>1979-2013</td>
</tr>
<tr>
<td>13</td>
<td>Fairbanks, AK</td>
<td>FAI</td>
<td>64.82</td>
<td>147.87</td>
<td>1979-2013</td>
</tr>
<tr>
<td>14</td>
<td>Nome, AK</td>
<td>OME</td>
<td>64.51</td>
<td>165.43</td>
<td>1979-2013</td>
</tr>
<tr>
<td>15</td>
<td>Baker Lake, NU</td>
<td>YBK</td>
<td>64.30</td>
<td>96.08</td>
<td>1979-2013</td>
</tr>
<tr>
<td>16</td>
<td>Coral Harbor, NU</td>
<td>YZS</td>
<td>64.19</td>
<td>83.35</td>
<td>1979-2013</td>
</tr>
<tr>
<td>17</td>
<td>Iqaluit, NU</td>
<td>YFB</td>
<td>63.75</td>
<td>68.54</td>
<td>1979-2010</td>
</tr>
<tr>
<td>18</td>
<td>Mayo, YT</td>
<td>YMA</td>
<td>63.62</td>
<td>135.87</td>
<td>1979-2013</td>
</tr>
<tr>
<td>19</td>
<td>McGrath, AK</td>
<td>MCG</td>
<td>62.95</td>
<td>155.60</td>
<td>1979-2013</td>
</tr>
<tr>
<td>20</td>
<td>Yellowknife, NWT</td>
<td>YZF</td>
<td>62.47</td>
<td>114.44</td>
<td>1979-2010</td>
</tr>
<tr>
<td>21</td>
<td>Talkeetna, AK</td>
<td>TKA</td>
<td>62.32</td>
<td>150.09</td>
<td>1979-2013</td>
</tr>
<tr>
<td>22</td>
<td>Fort Simpson, NWT</td>
<td>YFS</td>
<td>61.75</td>
<td>121.23</td>
<td>1979-2013</td>
</tr>
<tr>
<td>23</td>
<td>Anchorage, AK</td>
<td>ANC</td>
<td>61.17</td>
<td>150.00</td>
<td>1979-2013</td>
</tr>
<tr>
<td>24</td>
<td>Bethel, AK</td>
<td>BET</td>
<td>60.78</td>
<td>161.83</td>
<td>1979-2012</td>
</tr>
<tr>
<td>25</td>
<td>Whitehorse, YT</td>
<td>YXY</td>
<td>60.72</td>
<td>135.07</td>
<td>1979-2010</td>
</tr>
<tr>
<td>26</td>
<td>Watson Lake, YT</td>
<td>YQH</td>
<td>60.12</td>
<td>128.83</td>
<td>1979-2013</td>
</tr>
<tr>
<td>27</td>
<td>Fort Smith, YT</td>
<td>YSM</td>
<td>60.02</td>
<td>111.95</td>
<td>1979-2013</td>
</tr>
</tbody>
</table>
CHAPTER 5

DISCUSSION AND CONCLUSIONS

5.1 Synthesis of Research Findings

This dissertation research utilizes synoptic climatological techniques to analyze interactions between regional atmospheric patterns and sea ice variability in the western Arctic Ocean and assess the extent to which that ice variability affects the frequency of polar surface weather types throughout high latitude North America. The results herein clearly link changes in western Arctic MSLP (Chapter 2) coupled with 1000-500 hPa thickness patterns (Chapter 3) to the region’s September sea ice extent and freeze-up variability respectively. These relationships are especially strong during recent years of noteworthy sea ice change. The long-term change in freeze-up onset dates is also congruent with the increase (decrease) in the October – March frequency of MP (DP) weather types over time across much of Alaska and western Canada (Chapter 4).

More specifically, Chapter 2 (e.g. Ballinger and Sheridan, 2014) identifies statistically significant multidecadal linkages between collective, warm season MSLP patterns and western Arctic sea ice extent variability from 1979-2011. The occurrence of the Beaufort Sea High (BSH; i.e. CP 11) during the summer months plays a key role in impacting some of the recent regional losses likely through a combination wind-related
forcing and solar heating of the open ocean. In 2007, the pattern occurred ~7-15 days more than average (1979-2006) during each summer month (June – August) preceding the record-setting ice minima observed that September (later eclipsed in 2012). During the four summers that followed (2008-2011) the BSH continued to occur more often than the climatology during each summer month as the ice continued a precipitous decline. Further, CP 11 June and July time series are significantly anti-correlated with the AO and AD teleconnections during those months, indicating that the persistence of negative AO and AD phases, indicative of high amplitude upper-level flow, may be responsible for concurrent spikes in BSH occurrences.

Chapter 3 builds on the dynamical implications of Chapter 2 by assessing links between warm season MSLP and 1000-500 hPa thickness pattern variability and the western Arctic freeze dates observed from 1979-2013. Summer-dominant MSLP P11 (i.e. BSH) increased ~30 total days (>4 days year⁻¹) during both June and July from the 2000-2006 to 2007-2013 periods. One relatively warm 1000-500 hPa pattern, T13, increased at least 17 total days (≥2.43 days year⁻¹) during June-August, while another, T15, increased 19 days in September and 8 days during October over the same periods (as the BSH) coincident with increasingly persistent freeze-up. Correlation and regression analyses further associate these increasing disparate atmospheric pattern frequencies with the changing freeze date over time and support the notion that early/middle warm season (June/July) BSH occurrence increases and thickness patterns occurring unseasonably late, in particular, are driving the delay of the western Arctic sea ice freeze-up.

Chapter 4 bridges the causal analyses presented in Chapters 2 and 3 with a foray into the impacts of the freeze-up date variability on high latitude North American climate,
1979-2013. The two autumn and winter dominant SSC surface weather types, Moist Polar (MP) and Dry Polar (DP), are found to be changing with MP increases largely offset by DP decreases over much of the domain. Specifically, the rise of MP (DP) occurrence is positively (negatively) correlated with the freeze-up dates across much of the western portion of the SSC station domain, including Alaska and the Yukon Territory. Composites of MP and DP frequencies during anomalous late versus early freeze-up years further exaggerate spatial patterns of lower atmospheric warming and moistening manifested through larger numbers of MP days, particularly at stations near the western Arctic Ocean. Assessments of individual, extreme late and early freeze years revealed some spatial patterns in MP and DP classified days, however the spatial responses are stronger and more conclusive in the composite results encompassing multiple extreme sea ice years, suggesting other factors may play a critical role in surface weather type variability as well.

5.2 Discussion of Results

The primary results of this dissertation have interrelated themes that collectively imply that the greater western Arctic Ocean and surrounding high latitude lands are in the midst of notable climatic and environmental change. Remarkable changes in sea ice and climatic conditions seem to be common in western Arctic over the last decade and especially during summers since 2007, perhaps signifying the beginning of a “new normal” for the region as suggested by Wood et al. (2013). The observed sea ice extent retreat and freeze date delays are statistically linked to increases in summertime high pressure patterns over the western Arctic Ocean and increases in thermal patterns during the mid-to-late summer which may be the result of changing ocean-atmosphere heat exchanges
fueled by ice losses. Western Arctic sea ice freeze-up persistence also appears to feed back on the climate system by leading to more humid and warm surface weather conditions over time across high latitude North America.

Recent research has documented various aspects of the climatic and environmental change observed in the western Arctic (e.g. Overland, 2009; Wood et al., 2013; Ballinger and Rogers, 2013, 2014; Stroeve et al., 2014; Xia et al., 2014). The objectives explored in this dissertation will contribute to and expand upon the rather limited, existing body of literature addressing the regional atmospheric causes and impacts western Arctic sea ice changes. While many of studies have utilized pan-Arctic-to-hemispheric scale atmospheric fields aggregated over seasons to explore questions surrounding sea ice-climate interactions, this research constrains the atmospheric data largely to the ice domains studied, thereby resolving the typical, daily synoptic patterns across these areas. The use of concise domains combined with synoptic classification of daily atmospheric data yields MSLP and 1000-500 hPa thickness patterns specific to the western Arctic environment and these patterns are monitored as it relates to the region’s ice cover. Mean low-level summer circulation in the western Arctic has become increasingly anticyclonic in recent years (e.g. Overland et al., 2012; Ogi and Rigor, 2013), but the fact that the predominant flow has shifted from cyclonic to anticyclonic conditions reveals little about how often any particular iteration of a BSH MSLP pattern occurs over a given period of time. Quantifying the temporal variability of atmospheric patterns at subseasonal time steps is one unique outcome of this analysis and allows for a more complete understanding of how persistent atmospheric conditions, or a lack thereof, explain some of the ice cover variability within a region of rapid change.
This research also advances western Arctic ice-climate understanding by establishing a clear link between the multidecadal and interannual frequency variability of June BSH patterns and both the end-of-summer ice minimum and the autumn freeze-up dates. Increased June BSH frequencies since 2007 parallel large sea ice melt summers and later freeze onset during autumn, suggesting that anomalous occurrences of early summer surface anticyclones in the future may significantly impact aspects of the sea ice cover in the months that follow. The structure of upper level atmospheric circulation across the high and middle latitudes has become more meridional in recent years, yielding more blocking patterns over the North American Arctic (e.g. Overland et al., 2012; Francis and Vavrus, 2015). Results described at the end of Chapter 2 and inferred in Chapter 3 (as well as corroborated by the tangential work of Ballinger et al. [2014]) frame recent BSH pattern persistence as a regional climatic impact of increasingly amplified, wavy atmospheric circulation.

This dissertation research also shapes understanding of regional sea ice impacts on high latitude climate by providing a case study focused on the North American Arctic (Chapter 4). Sea ice loss plays a central role in lower tropospheric warming and increasing humidity in the Arctic (e.g. Serreze et al., 2009, 2012). In this chapter, the low frequency variability of the sea ice freeze-up time series appears to play a role in creating a warmer and more humid environment for much of the study area, through the transition of SSC DP to MP weather types, during the fall and into the winter months. Francis et al. (2009) noted that anomalous autumn and winter SAT across North America lead areal summer sea ice extremes, but the authors make general regional comments about these relationships across space. By limiting the study domain to stations poleward of 60°N and examining
individual months of SSC data, this research identifies spatial patterns where the freeze-up co-varies with relatively warm/humid or cool/dry daily surface weather conditions across high latitude North America. The results also provide a baseline for future research seeking to understand the potential spatiotemporal boundaries of a sea ice-related climate signal.

Aside from acknowledging some of the causes and impacts of regional sea ice variability, this dissertation has showcased different synoptic climatological methods and datasets as useful tools for evaluating Arctic sea ice-climate interactions, particularly in the western Arctic study area. The application of the atmospheric/circulation pattern classification described in Chapters 2 and 3 is shown to be useful in not only portraying the typical spatial patterns of atmospheric circulation, but also creating patterns whose temporal characteristics (i.e. frequencies) robustly describe the sea ice minimum and freeze-up. Future synoptic climatological studies relating gridded, model-derived atmospheric circulation fields to Arctic marginal sea ice cover may consider a similar atmospheric/circulation pattern classification procedure and accompanying analytical techniques.

The SSC dataset was also studied and applied in unique ways in this dissertation analysis. Relatively little research has used the SSC data in an attempt to understand cryospheric change (e.g. Leathers et al., 2002; Dyer and Mote, 2007) let alone Arctic climate variability and change. In light of anticipated future climatic changes, the frequency of weather patterns (and extreme conditions) is also expected to change (IPCC, 2012). The results of Chapter 4 present North American DP and MP weather type frequency changes within the autumn and winter months as potentially affected by the evolving sea ice cover bordering the continent. Forthcoming projects may expand or explore separate analysis
domains to better understand relationships between sea ice variables and anomalous or changing frequencies of specified weather regimes.

5.3 Potential Improvements and Research Directions

While the dissertation research herein highlights robust regional sea ice-climate relationships, some alterations to the methodological framework may improve future results. Chapter 2 describes a circulation pattern scheme developed using daily MSLP data, 1979-2011. The high degree of statistical association between the MSLP patterns and the ice extent shown in that chapter motivated its use, with updated pressure data through 2013, in Chapter 3. However, the simple integration of the “new” data proved problematic when the circulation pattern classification procedure was re-run and revealed spatial and temporal inconsistencies relative to the 1979-2011 classification. As a result, discriminant function analysis was conducted to integrate the 2012-2013 data, and MSLP patterns of similar spatial appearance (i.e. pressure field magnitudes comparable to those produced in the original classification), but slightly different temporal characteristics were generated (i.e. monthly mean frequencies slightly changed). Future work may look to improve the methodology surrounding data integration into existing synoptic categorizations. Subsequent research may also use more innately stable classification schemes, such as self-organizing maps (SOMs), which show a continuum of patterns, versus the discretized, unordered number of circulation patterns across a region typical of traditional cluster analysis (e.g. Sheridan and Lee, 2011). Use of SOMs may ease both the interpretation of temporal transitions between patterns as well as differentiate between multiple spatial
orientations of semi-permanent climatic features, such as the BSH, which typically manifest as only one or two patterns as shown in the PCA/cluster-based synoptic classifications.

Studies of regional sea ice-climate interactions will undoubtedly progress in various directions in the future, many of which have been suggested or implied throughout this dissertation document. Further spatiotemporal evaluation of a potential sea ice-related climate signal across North America, which could impact large numbers of people and ecosystems, is warranted. This dissertation identifies increases in MP weather types to 60°N as a potential manifestation of sea ice change. However, future work expanding the weather type domain to include all of Canada and the contiguous United States, where the majority of stations producing SSC data are located, may be useful in understanding both the geographic extent of these holistic weather type-sea ice relationships across the middle latitudes and how they vary over time. Development of continental-to-hemispheric scale upper-level circulation pattern catalogues may also be useful to evaluate potential links between recent Arctic sea ice changes and cold outbreaks across North America (e.g. Overland et al., 2011; Francis and Vavrus, 2012; Ballinger et al., 2014; Francis and Vavrus, 2015). Research involving connections between the Arctic and boreal middle latitude climates is critical moving forward given the likelihood that these regions will continue to experience climatic changes during at least the 21st century (e.g. Overland et al., 2014; Screen et al., 2015).
References


