Developing Novel Storminess Metrics and Evaluating Seasonal Predictability of Storminess Indicators in the North Pacific and Alaskan Regions

by

Norman Shippee

BSc., Plymouth State University, 2008
MS., Plymouth State University, 2010

A Dissertation Submitted in Partial Fulfillment of the Requirements for the Degree of

DOCTOR OF PHILOSOPHY

in the Department of Geography

© Norman Shippee, 2016
University of Victoria

All rights reserved. This dissertation may not be reproduced in whole or in part, by photocopy or other means, without the permission of the author.
Supervisory Committee

Developing Novel Storminess Metrics and Evaluating Seasonal Predictability of Storminess Indicators in the North Pacific and Alaskan Regions

by

Norman Shippee
BSc., Plymouth State University, 2008
MS, Plymouth State University, 2010

Supervisory Committee

Dr. David Atkinson (Department of Geography)
Supervisor

Dr. Daniel Smith (Department of Geography)
Departmental Member

Dr. Francis Zwiers (Department of Mathematics and Statistics)
Outside Member
Abstract

Supervisory Committee

Dr. David Atkinson (Department of Geography)
Supervisor

Dr. Daniel Smith (Department of Geography)
Departmental Member

Dr. Francis Zwiers (Department of Mathematics and Statistics)
Outside Member

Extratropical cyclones (ETCs) are a common feature of mid- and high-latitudes which, on a large scale, are a primary mechanism by which heat and moisture are transported from equator to pole. ETCs also exert a major impact at smaller scales. Communities along the western coast of Alaska face many types of impacts generated by the winds associated with ETCs, including storm surges, sea water intrusion into fresh water stores, and coastal erosion. Such “strong wind events”, which can occur independent of an ETC, can also generate hazardous sea states and associated impacts on shipping. With no roads, coastal Alaska relies heavily on marine and air transportation. Hazards posed to marine and air travel are often related to two main types of weather: wind and fog. Consultations with stakeholders in the marine transportation community have indicated more precisely specific aspects of poor weather, such as high wind events, that are problematic, including the idea that the periods between strong wind events,
defined as lull periods, represent an important metric when planning travel between points of safe harbour.

Three separate studies of storminess metrics in the North Pacific and Alaskan regions are presented. The first study presents both a comparison of two storm identification and tracking algorithms and an evaluation of the general characteristics of extratropical cyclones for the North Pacific as portrayed in two reanalyses. The second study applies a modified wind event identification algorithm to reanalysis data to evaluate the spatial climatological patterns of wind events in the circum-Arctic. The third study tests the statistical relationships and predictability of two measures of storm activity - cyclone track density (TDEN) and wind event frequency - in the North Pacific using teleconnection indices exhibiting local influence. The first study showed that the general patterns and trends of cyclone characteristics are similar between the two methods, though with increased values of cyclogenesis density, cyclolysis density, and track density when using the relative vorticity based method. A comparison between storm tracks for NCEP1 and the 56-member ensemble of the Twentieth Century Reanalysis v2 (20CR) shows distinct differences between the 20CR and NCEP1 mean climatology for main storminess indicators. The second study evaluated the spatial and temporal characteristics of wind events and introduced a novel indicator that characterizes periods of favorable weather between strong wind events that last 48-hours or longer, termed lull events. Lull periods were found to be an important consideration for northern marine operations – both economic and subsistence. Additionally, combinations of lull and wind event indicators, termed lull/storm winds (LSW), were analyzed and showed that preferred areas of wind events and lull events are not always spatially coherent. The third
study tested the statistical relationships and predictability of two measures of storm activity - cyclone track density (TDEN) and wind event frequency - in the North Pacific using teleconnection indices with local influence for the winter period of 1950 - 2012. Two statistical modeling techniques are applied to evaluate linear and non-linear methods of prediction for the region. For both measures of storm activity, the North Pacific index, Niño 3.4 index, and the AO index were found to be the best predictors. Using a 23-year hindcast period (1980 – 2012), the region of highest wind event anomaly prediction skill is located in the western Bering Sea, with hindcast correlation values as high as +0.5 and root mean squared skill scores (RMSESS) 25% higher than climatology. Highest TDEN predictive skill from the 23-year hindcast is found in the southeast region of the North Pacific, near the California coastline, with correlation and RMSESS as high as +0.7 and 25 - 30%, respectively.
# Table of Contents

Supervisory Committee ........................................................................................................ ii

Abstract .................................................................................................................................. iii

Table of Contents ..................................................................................................................... vi

List of Tables .......................................................................................................................... x

List of Figures ........................................................................................................................ xi

Acknowledgements ................................................................................................................ xiv

Dedication ............................................................................................................................... xvi

Chapter 1 Introduction ........................................................................................................... 1

Chapter 2 Defining a Storm .................................................................................................... 10

  2.1 Theoretical Perspective of Storm Definition ................................................................. 10
  2.2 Pragmatic Perspective of Storm Definition .................................................................. 12
  2.3 Tracking Perspective of Storm Definition ........................................................................ 13
  2.4 Previous Research on Extratropical Cyclones .............................................................. 14
  2.5 Teleconnections and Cyclone Activity Impacts ............................................................ 17

Chapter 3 Methodology ....................................................................................................... 22

  3.1 Objective Identification and Tracking Methods ............................................................. 24
    3.1.1 The Serreze Method .................................................................................................. 26
    3.1.2 The Hodges Method ................................................................................................. 29
    3.1.3 The Atkinson Method ............................................................................................... 31
  3.2 Overall Weaknesses of Objective Identification ........................................................... 31
# Chapter 4. Climatological Patterns of Cyclone Activity in the North Pacific and Alaskan Regions using the Twentieth Century Reanalysis (20CR)

<table>
<thead>
<tr>
<th>Section</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>Article Information</td>
<td>42</td>
</tr>
<tr>
<td>Author Information</td>
<td>42</td>
</tr>
<tr>
<td>Author Contributions</td>
<td>42</td>
</tr>
<tr>
<td>Abstract</td>
<td>42</td>
</tr>
<tr>
<td>4.1 Introduction</td>
<td>44</td>
</tr>
<tr>
<td>4.2 Data and Methodology</td>
<td>47</td>
</tr>
<tr>
<td>4.2.1 Reanalysis Data</td>
<td>47</td>
</tr>
<tr>
<td>4.2.2 The Serreze Identification and Tracking Algorithm</td>
<td>49</td>
</tr>
<tr>
<td>4.2.3 The Hodges Identification and Tracking Algorithm</td>
<td>50</td>
</tr>
<tr>
<td>4.2.4 Study Area and Climatology Specifics</td>
<td>51</td>
</tr>
<tr>
<td>4.3 Results</td>
<td>51</td>
</tr>
<tr>
<td>4.3.1 Serreze and Hodges NCEP1 Comparison</td>
<td>51</td>
</tr>
<tr>
<td>4.3.2 Hodges NCEP1 and 20CR Comparison</td>
<td>53</td>
</tr>
<tr>
<td>4.3.3 Sub-regional Storm Characteristics in the 20CR</td>
<td>55</td>
</tr>
<tr>
<td>4.3.4 Time Series and Trend Analysis</td>
<td>57</td>
</tr>
<tr>
<td>4.4 Discussion</td>
<td>61</td>
</tr>
<tr>
<td>4.4.1 Serreze and Hodges NCEP1 Climatology Comparison</td>
<td>61</td>
</tr>
<tr>
<td>4.4.2 NCEP1/20CR Hodges track comparison</td>
<td>64</td>
</tr>
<tr>
<td>4.5 Summary</td>
<td>68</td>
</tr>
</tbody>
</table>
Chapter 5 Seasonal Climatology and Trends of Strong Wind and Lull Events in the Circum-Arctic During the 1979 – 2010 Period Using a Novel Lull/Storm Wind Indicator

Article Information........................................................................................................ 91

Author Information ........................................................................................................ 91

Author Contributions .................................................................................................... 91

Abstract .......................................................................................................................... 92

5.1 Introduction .................................................................................................................. 93

5.2 Data and Methods ....................................................................................................... 95

  5.2.1 LSW Algorithm .................................................................................................... 95

  5.2.2 Data Sources ....................................................................................................... 97

5.3 Results ........................................................................................................................ 99

  5.3.1 Climatology ......................................................................................................... 99

5.4 Discussion and Conclusions ........................................................................................ 103

Chapter 6. The Potential for Seasonal Forecasting of Winter Storminess Indicators in the North Pacific and Alaskan Regions ........................................................................ 123

Article Information ........................................................................................................ 123

6.1 Introduction ................................................................................................................ 124

6.2 Data ........................................................................................................................... 128

6.3 Results: Correlations ............................................................................................... 130

  6.3.1 Teleconnections and Wind Event Anomaly .......................................................... 130

  6.3.2 Teleconnections and Cyclone Track Density (TDEN) Anomaly ......................... 132

6.4 Results: Statistical Prediction .................................................................................... 133
6.5. Discussion and Conclusion ................................................................. 135
6.6 Future Work ......................................................................................... 139

Chapter 7 Conclusion .............................................................................. 146

7.1 Introduction ......................................................................................... 146
7.2 Main Research Results and Key Points ............................................. 147
7.3 Conclusion ......................................................................................... 151
7.4 Future Work ......................................................................................... 154

References ............................................................................................... 156
List of Tables

Table 3.1. Three methods for extratropical storminess identification. Hodges and Serreze methods identify and track cyclones while the Atkinson method identifies storm events based on wind patterns.................................................................40

Table 3.2. Hodges TRACK algorithm steps for identification and tracking of features, as applied within this study.................................................................41

Table 4.1. Regions and sub-regions used in this study........................................71

Table 4.2. Type-1 Changepoints for the cyclogenesis and cyclolysis annual time series for the listed sub-regions from the Hodges 20CR for the 1871 – 2010 period.................................................................................................72

Table 4.3. Trends and significance of lysis density and genesis density by sub-region for the a) Hodges NCEP1 database and the b) Serreze NCEP1 database. Periods evaluated for trends are 1950 – 2010, 1950 – 1978, 1979 – 2010. Units are % yr⁻¹. Trends that are significant at 95% are indicated by bold and italic numbers. Trends that are significant at 99% are indicated by bold, italic, and underlined numbers. Non-highlighted values are not statistically significant.......................................................................................73

Table 4.4. Trends and significance of lysis density and genesis density by sub-region for the Hodges 20CR database. Periods evaluated for trends are 1950 – 2010, 1950 – 1978, 1978 – 2010, 1920 – 2010, and 1920 – 1949. Units are % yr⁻¹. Trends that are significant at 95% are indicated by bold and italic numbers. Trends that are significant at 99% are indicated by bold, italic, and underlined numbers. Non-highlighted values are not statistically significant.......................................................................................75

Table 5.1. Slopes, expressed as percent change per year, of the seasonal wind event frequency anomaly trendline for selected subregions, 1979 – 2010...............................111

Table 5.2. Slopes, expressed as percent change per year, of the seasonal 48-hour+ lull event frequency anomaly trendline for selected subregions, 1979 – 2010 ....................112

Table 6.1 Teleconnection/climate indices included in both GLM and random forest regression models. Indices were downloaded from NCAR/UCAR Climate Data Guide (http://climatedataguide.ucar.edu) and NOAA Climate Prediction Center (http://cpc.ncep.noaa.gov) in March 2014.........................................................140
List of Figures

Figure 4.1. Map of study area with some key locations highlighted. Sub-regions in the Gulf of Alaska, Bering Sea, Chukchi and Beaufort Seas, and Alaska Interior are also used in this analysis.................................................................77

Figure 4.2. Cyclogenesis density climatology for the a) Winter (JFM), b) Spring (AMJ), c) Summer (JAS), and d) Fall (OND) seasons using the MSLP-based Serreze cyclone identification and tracking algorithm with the NCEP1 reanalysis for the 1950 – 2010 period. Units: density of starting points (10^6 km^2 season)^{-1} ..................................................78

Figure 4.3. As in Fig. 4.2 but for the 850 hPa relative vorticity-based Hodges storm identification and tracking algorithm.............................................................................79

Figure 4.4. Cyclolysis density climatology for the a) Winter (JFM), b) Spring (AMJ), c) Summer (JAS), and d) Fall (OND) seasons using the Serreze cyclone identification and tracking algorithm with the NCEP1 reanalysis for the 1950 – 2010 period. Units: density of ending points (10^6 km^2 season)^{-1} ......................................................................................80

Figure 4.5. As in Fig. 4.4 but for the 850 hPa relative vorticity-based Hodges storm identification and tracking algorithm.............................................................................81

Figure 4.6. Cyclone track density climatology for the a) Winter (JFM), b) Spring (AMJ), c) Summer (JAS), and d) Fall (OND) seasons using the MSLP-based Serreze cyclone identification and tracking algorithm with the NCEP1 reanalysis for the 1950 – 2010 period. Units: storms (10^6 km^2 season)^{-1} ......................................................................................82

Figure 4.7. As in Fig. 4.6 but for the 850 hPa relative vorticity-based Hodges storm identification and tracking algorithm.............................................................................83

Figure 4.8. Cyclogenesis density climatology for the a) Winter (JFM), b) Spring (AMJ), c) Summer (JAS), and d) Fall (OND) seasons using the 850 hPa relative vorticity based Hodges cyclone identification and tracking algorithm for the 20CR reanalysis matched tracks between the 56 ensemble members for the 1950 – 2010 period. Units: density of starting points (10^6 km^2 season)^{-1} ......................................................................................84

Figure 4.9. As in Fig. 4.8 but for 20CR cyclolysis density. Units: density of ending points (10^6 km^2 season)^{-1} ......................................................................................85

Figure 4.10. Difference between climatological cyclogenesis density between the Hodges based NCEP1 and Hodges based 20CR for the a) Winter (JFM), b) Spring (AMJ), c) Summer (JAS), and d) Fall (OND) seasons for the 1950 – 2010 period. Positive (red)
values indicate higher density in the NCEP1 while negative (blue) values indicate higher density in the 20CR climatology. Units: density of starting points \((10^6 \text{ km}^2 \text{ season})^{-1}\).

Figure 4.11. As in figure 4.10, except NCEP1 – 20CR cyclostasis density. Units: density of ending points \((10^6 \text{ km}^2 \text{ season})^{-1}\).

Figure 4.12. a) Annual base anomaly series (reference period 1950 – 2010) with significant Type-1 changepoints of storms undergoing cyclogenesis and b) mean adjusted base series for Bering Sea region in the Hodges 20CR ensemble mean from matched tracks. The same is shown in fig. 13 c and d, except for cyclostasis.

Figure 4.13. Same as Fig. 4.12, except for the Chukchi/Beaufort region.

Figure 4.14. Annual (a) zonal and (b) meridional 850 hPa wind differences between NCEP1 and 20CR for the study area. Units are ms\(^{-1}\).

Figure 5.1. Schematic representation of the LSW algorithm for (a) wind event identification and (b) lull event identification with indications of the various components of the algorithm identified.

Figure 5.2. Wind event climatology from the LSW algorithm for the circum-Arctic region for the (a) winter (JFM), (b) spring (AMJ), (c) summer (JAS), and fall (OND), 1979 - 2010. Units are frequency of wind events per season.

Figure 5.3. Same as figure 5.2 except for 48-hour lull events.

Figure 5.4. Wind event trends from the LSW algorithm, 1979 – 2010. Locations with trends significant at the \(p < 0.1\) level are contoured and shaded. Locations with trends significant at the \(p < 0.05\) level are indicated a black dot. Magnitude of the trend is shown by the color, with increasing trends shown in red and decreasing trends shown in blue. Boxes labeled with letters A. and B have trends highlighted in Table 1.

Figure 5.5. Same as figure 5.4, except for 48-hour lull events. Boxes labeled with letters A. and B have trends highlighted in Table 2.

Figure 5.6. Probability distribution function (PDF) and distribution best fits of all lull events duration, 1979 – 2010. The 48-hour threshold value is indicated on the figure with a vertical red line.

Figure 5.7. Cumulative distribution function (CDF) and distribution best fits of all lull events duration, 1979 – 2010. The 48-hour threshold value is indicated on the figure with a vertical red line.
Figure 5.8. Subregion trends for (a) wind event frequency, (b) 48-hour lull event frequency, and (c) all duration lull frequency for the North Atlantic and Canadian Maritimes. Locations with trends significant at the $p < 0.1$ level are contoured and shaded. Locations with trends significant at the $p < 0.05$ level are indicated a black dot. Magnitude of the trend is shown by the color, with increasing trends shown in red and decreasing trends shown in blue.

Figure 5.9. Climatological mean percentage of time spent in wind event criteria (SP) by season, divided into (a) winter (JFM), (b) spring (AMJ), (c) summer (JAS), and (d) fall (OND). Darker shades indicate higher percentage of time spent in wind event criteria.

Figure 5.10. Same as figure 5.7 except climatological mean percentage of time spent in lull criteria (LP).

Figure 6.1. Study area map with key locations highlighted.

Figure 6.2. Pearson correlation coefficient (R) between winter (JFM) teleconnection index values and wind event anomaly for (a) Nino 3.4, (b) PDO, (c) NP index, (d) PNA, and (e) AO. Grid locations with significant correlation (Alpha = 0.05) are highlighted by white symbols. Time period for correlation analysis is 1950 – 2012 and R is multiplied by 100.

Figure 6.3. Same as Figure 6.2, except for cyclone track density (TDEN) anomaly.

Figure 6.4. Hindcast skill obtained from the use of GLM and Random Forest ensemble regression with wind event anomaly as the predictand. Pearson correlation coefficient with hindcast ($r_{hind}$) with significant positive correlations ($\alpha = 0.05$, $r_{hind} > +0.337$) for the a) 23-year GLM hindcast and c) 23-year Random Forest hindcast. As shown are the root-mean squared skill score (RMSESS) values above zero with reference to climatology for the a) 23-year GLM hindcast and the d) 23-year Random Forest hindcast.

Figure 6.5. As in Figure 6.4, except for TDEN.
Acknowledgements

My PhD would not have been possible without the multiple friends and colleagues whom have aided me along the way. It began with a cross-country, international adventure with my wife and a dog. Along the way, there were challenges that I never could have expected to face, but I am indebted to many people for their help along the way. My most heartfelt thanks go to my supervisor, David Atkinson. David, you and your family have been my extended family here. Thank you for the countless rides to the airport, entertaining my family and friends on their visits, dog sitting for us when we have been traveling, the never-ending pumpkin scones, and everything else I am forgetting to mention. I could not have had a more supportive supervisor, and I am forever grateful for that. A most sincere thank you to my committee members Dan Smith and Francis Zwiers for your support and guidance. It has been a wonderful experience learning from such great scholars and kind people. Additionally, thank you to Jim Overland for the thoughtful defence, invaluable comments, and the inspirations for future work.

This dissertation was funded in part by the National Oceanic and Atmospheric Administration (NOAA) under the project “Social Vulnerability to Climate Change in the Alaskan Coastal Zone” (grant NA08OAR4600856) and the Marine Environmental Observation Prediction and Response (MEOPAR) network project 2.3 “User-driven Monitoring of Adverse Marine and Weather States in the Eastern Beaufort Sea.” I was lucky enough to spend time in Fairbanks, Alaska at the International Arctic Research Center (IARC) during my studies, and truly appreciate the kindness and guidance of many of the faculty and staff there, including Javier Fochesatto, Larry Hinzman, and John Walsh. Thank you for your support and great discussions during my studies.

To my climate lab comrades - Adam, Weixun, Connie, Vida, Mohammed, Chris, Laura, Eric, and Ben: thank you all for being such great colleagues and friends. Each of you owns a piece of this work through your friendship and support throughout the years. Thank you for making the climate lab such a great place to “work” in the Geography
department, and I look forward to more invigorating hockey talk in the coming years. There are others that I am forgetting to mention, but your friendship is not forgotten.

To my family – Mom and Dad, Ian, Kai, Tilly, Jane, Gus, Dan, Maureen, Steve, James, and Lindsay – thank you for all of your support and love throughout the past five years. It may seem like it has been a long time, but it never felt that long thanks to all of your messages, phone calls, and visits. I look forward to having more time for visits now!

To my east coast friends – Matt, Jared, Justin, Brian, Peter, and Eddie – thank you for all of your help, support, and visits (both virtual and physical) throughout the years. From laughter and programming to football and data requests, I truly appreciate all of the time and messages during this process. We may live far apart, but our friendships remain tied to the East Coast. Brian, we can go skiing or hiking next time you are out here. I’m sure we can find a good set of maps at Costco.

Finally, to Katherine and Denali – Thank you for being my biggest fans and best editors. I will never be able to repay how much you have done for me. From the bottom of my heart, thank you.
Dedication

For Katherine, Beebee, and Denali – who each are part of this work.
Chapter 1 Introduction

Climate change presents one of the most pivotal and controversial challenges human society has ever faced. Often at the forefront of climate change discussions are fluctuations in storms and storminess. Extratropical cyclones (ETCs) are the predominant transport mechanism of heat and moisture from equator to pole (IPCC, 2012), however the discussion is much more relevant to the human scale when the focus shifts to changes in the frequency, timing, and intensity of extreme events. Thus, future changes in ETC activity are important to understand due to the associated impacts on the environment and society. The favoured storm tracks in the mid- and high-latitudes of the Atlantic and Pacific Ocean basins have a great impact on regional climate, exposing coastal regions to potential extreme events, such as storm surge, coastal erosion, high winds, and flooding. These events impact low lying towns and coastally located cities and represent a primary coastal geomorphological agent. Within the North Pacific and Alaskan regions, ETCs are known for generating these types of high impact weather events, as well as phenomena such as enhanced sea states (Mason et al., 1993; Blier et al., 1997; Hufford and Partain, 2004; Robinson, 2004; Cross, 2010; Mesquita et al., 2010; Pingree-Shippee et al., 2016), and are known to have a major environmental and human impacts. For example, within Alaska, many small coastal villages face relocation due to the impacts of successive storms in a climate of decreasing sea ice, which previously provided a protective buffer against storm surges and associated coastal erosion (Robinson, 2004; Mittal, 2009). As of 2004, the United States Government Accountability Office (GAO) noted that 86% of Alaska Native villages faced impacts from flooding and erosion (Robinson, 2004). In 2009, the date of the latest government report on Alaska Native Village vulnerability, the
GAO noted that little progress had been made to address vulnerability from flooding and erosion impacts (Mittal, 2009). Projections of climate change show that main storm tracks in both the Atlantic and Pacific are expected to shift poleward, which would bring increased storm activity to the Arctic regions (Bengtsson et al., 2006; Ulbrich et al., 2008; Ulbrich et al., 2009; Seiler and Zwiers, 2015). As such, future changes to ETCs, storm tracks, and storminess can be expected to have major impacts on both human and environmental systems at higher latitudes, including the Alaska region.

In the Alaska region, ETC impacts are a particularly important factor for stakeholders – both economically and subsistence based. Strong winds often generate hazardous sea states and, when combined with slow moving cyclones, can lead to both open sea and coastal impacts (Pingree-Shippee et al., 2016). Some of the most productive and dangerous fisheries in the world are located in the Bering Sea. The risks associated with these fisheries are complicated by their fall and winter timing, coincides with the climatological peak of storminess within the region (Mesquita et al., 2008; Mesquita et al., 2010).

Transport of people and commodities throughout much of Alaska is dependent on marine or air transportation due to the lack of roadways connecting coastal areas to the interior or major cities. Hazards posed to marine and air travel are often related to two main types of weather: wind and fog. The occurrence of strong wind events can delay or cancel transportation in both the marine and air sectors, posing economic and social constraints on the North. In the marine sector, many goods are shipped via tug-and-barge style transportation due to the lack of deep water ports along much of the western
coastline of Alaska. This style of transportation is highly impacted by marine state, primarily due to the flat-bottom design of barges.

Coastal communities within Alaska face different types of impacts, including storm surges, sea water intrusion into fresh water stores, and coastal erosion (Robinson, 2004; Mittal, 2009; Francis and Atkinson, 2012). Throughout the years, villages and towns along the Alaskan west coast have been impacted, sometimes severely, by powerful extratropical cyclones. For example, two of the most powerful cyclones to impact occurred in the fall seasons of 2004 and 2011. Both storms generated a 3-metre storm surge in the coastal hub community of Nome. Both surges inundated parts of the town, causing millions of dollars in damages. Following the 2011 storm, additional economic impacts were felt throughout the winter, as the timing of the storm caused the rescheduling (and eventual cancellation) of a fuel delivery to Nome by tug-and-barge. This forced a high-risk wintertime fuel delivery, requiring icebreaker support to reach the port, and resulted in a large expense for the community. As such, it is important to gain understanding of preferred cyclone trajectories in the North Pacific and Alaska region and the characteristics of impactful weather associated with such events.

The question of how to objectively identify cyclones has been approached from several perspectives. Some methods are based on an algorithm that attempts to replicate how a human observer might identify and track a cyclone. There are many objective methods for cyclone identification and tracking, which can be classified in different ways. Methods are often divided into categories according to the perspective and theoretical base. Perspective refers to the way that storminess is viewed; theoretically, pragmatically, or from a tracking perspective. These three categories are defined in
further detail in Chapter 2. The theoretical base of a storm identification and tracking algorithm also affects the way that storm activity is evaluated, allowing for identification and tracking (Lagrangian) or point-based assessments of storm activity (Eulerian). Lagrangian methods typically identify individual systems (e.g. a cyclone) one at a time and track them as they move. Climatological storm tracks can be created using many years worth of track data. These climatological storm tracks are useful for determining regions of cyclogenesis and cyclolysis and the storm track density. Alternatively, Eulerian methods focus on determining the synoptic patterns associated with storminess by analyzing the variance at multiple atmospheric levels at a synoptic temporal scale or by evaluating storminess at a given grid location, as defined by a parameter such as wind speed.

Lagrangian-based methods have been used extensively to create cyclone track climatologies and to conduct studies of trends and potential future changes to storm tracks and characteristics (e.g., Serreze et al., 1993; Murray and Simmonds, 1991a; Hoskins and Hodges, 2002; Hodges et al., 2003; Raible et al., 2008; Seiler and Zwiers, 2015). Eulerian methods (e.g. Blackmon, 1976; Blackmon et al., 1977; Hoskins and Hodges, 2002; Atkinson, 2005) are utilized less frequently in current literature primarily due to the inability of the method to directly identify and track individual cyclones. The Eulerian methods do, however, provide an ability to establish a distribution of events at a single observation station or grid point, or gather statistics about the atmosphere from a fixed frame of reference. Additionally, many Eulerian methods allow for the development of descriptive statistics about the variance of storm tracks (Blackmon, 1976; Blackmon et al., 1977). Nevertheless, some objective Lagrangian/Semi-Lagrangian
methods do duplicate this type of statistical effort by gathering statistical information about storms within the developed algorithm (e.g. Murray and Simmonds, 1991a; Hoskins and Hodges, 2002).

Extratropical cyclone activity is often delineated by different measures of storminess, mostly related to traditional cyclone or storm identification methods. For example, when using Lagrangian tracking methods, measures like cyclogenesis density, cyclolysis density, and storm track density, among others, are used to define spatial and temporal characteristics related to storm formation, decay, and track locations, respectively. Beyond storm activity definitions that focus on storms, the idea of breaks in storm activity, termed lulls, are also important for human activities – both commercial and subsistence – along the coastal regions of Alaska and the Arctic. Anyone who makes their living on the water in these areas is highly interested in how frequent and persistent breaks in storminess are within the context of their work. For instance, interviews conducted with captains from multiple marine transport corporations highlight the need for breaks between events (i.e. lulls) that allow for approximately 48 hours or more of safe operations. Without these breaks, operations risk being shut down and ships in transit may be forced into ports of refuge for long periods of time. In particular, along the sparsely settled Alaskan coastline where wind effects on sea state highly impacts tug-and-barge operations, lulls lasting at least 48-hours are often needed to permit navigation between points of refuge.

In addition to methods that summarize general characteristics of the cyclone climatology, other approaches assess cyclone activity in a region by using an integrative index. For example, Zhang et al. (2004) developed the Cyclone Activity Index (CAI) and
assessed the inter-annual variability of cyclones in the Arctic. This method included an identification and tracking component, using the Serreze (1995) method, to create a cyclone climatology before summarizing statistical information into the CAI. The index uses information about cyclone intensity, frequency, and duration to determine the CAI value. Cyclone intensity is determined by calculating the absolute value of the difference in central MSLP of cyclones to the climatological mean MSLP at a given grid point. Cyclone frequency is determined by counting the number of cyclone trajectories over a region in a given time period. Cyclone duration is defined by determining the average time that a cyclone exists somewhere in the grid field and is then averaged over the region of interest. Zhang et al. summed these intensity, frequency, and duration values to determine the value of CAI. Since its inception, CAI is often used in studies to analyze extratropical cyclone activity. For example, Bartholy et al. (2006) used CAI to research cyclones that impact the European region. More recently, Wang et al. (2012) applied CAI, using a modified version of the Serreze algorithm, to evaluate extratropical cyclones at the global scale as identified using mean sea level pressure in the ensemble members of the Twentieth Century Reanalysis (20CR).

A number of studies have identified relationships between teleconnections and various measures of storminess, linking large scale climate variability and various measures of extratropical cyclone activity. For example, correlations between storm activity and the El Niño Southern Oscillation (ENSO) have been established using various indices. A reduction in cyclogenesis in the southern Bering Sea and Aleutian Islands has been shown during El Niño years with respect to La Niña years (Key and Chan, 1999; Graham and Diaz, 2001). Mesquita et al. (2008) found correlation between
the summertime Northern Annular Mode (NAM) and the positioning of major baroclinic zones in the Arctic. Furthermore, the Pacific North American pattern (PNA) has been found to be correlated with mean sea level pressure (MSLP) based measures of storminess in the North Pacific (Mailier et al., 2006; Seierstad et al., 2007). In general, correlations between other indices of climatic variability and storm activity measures are weaker for the Pacific region, though some studies have shown correlations between the Pacific Decadal Oscillation (PDO) in the region (Chang and Fu, 2002). Further information about relationships between individual teleconnections and correlation with measures of extratropical storminess is provided in Chapter 2.

Despite the considerable research conducted thus far, there remain gaps in the knowledge of storm activity in the North Pacific and Alaskan regions. The effect that choosing a specific cyclone tracking method and its associated attributes has on the overall cyclone statistics, particularly in relation to the definition of a storm and the gridded field chosen, is not well understood. Also, new, longer term reanalyses ostensibly allow the ETC climatology within the study region to be extended back to the early 20\textsuperscript{th} century. The reliability of being able to extend the data back to this date, however, needs to be studied. Additionally, an assessment of storminess from a more pragmatic perspective within the study area is needed, particularly as it relates to stakeholders. These needs include assessments of on-the-ground manifestation of storminess and associated storm impacts. Assessments of the characteristics of periods between storm events are also of interest to these user groups. Furthermore, the ability to predict seasonal storminess indicators using teleconnections as predictors needs to be evaluated and understood for this region and its unique stakeholders.
Mesquita et al. (2008) argued that a contest between methods in the process of selecting the ‘best’ overall at representing storminess is not necessarily a profitable exercise, as this process involves an amount of subjectivity. This research takes an alternate approach, using the assumption that the means of defining a storm affects the detection and representation of storminess, particularly when approached from a stakeholder perspective. As such, this research is guided by the following hypothesis: the definition of a storm impacts the ability to define the statistical relationships between measures of climatic variability (e.g. teleconnections) and the storminess indicator. To test this hypothesis, the following research objectives are established:

- Determine how existing representations of storminess can be improved to better reflect the needs of end-users; in particular, Northern marine transportation interests. (**Objective 1**)

- Develop and evaluate climatologies based on multiple definitions of storminess, including Lagrangian and Eulerian methods, to discover the influence of both different reanalyses and tracking methods. (**Objective 2**)

- Determine statistical relationships between storminess indicators and teleconnection indices and establish the predictability of the storminess indicators during the winter season for the North Pacific and Alaskan regions. (**Objective 3**)

The remainder of this dissertation is divided as follows: Chapter 2 provides background into the concept of how a storm is defined and the typical influences of different teleconnections on storm tracks. Chapter 3 explores the methodology used in this dissertation, including cyclone tracking algorithms, storm event algorithms, and the basics of the statistical methods used for seasonal prediction in this study. Chapter 4
investigates the climatological differences between two popular objective storm identification methods and the assesses the value-added obtained by using a longer term ensemble-based reanalysis for creating extended climatologies in the North Pacific and Alaskan regions. Chapter 5 evaluates storminess from a pragmatic, end-user based perspective, presenting high spatial resolution climatologies of wind events and non-windy periods (lulls) and providing a novel assessment of circum-Arctic storminess. Chapter 6 explores the statistical relationships between storminess indicators and teleconnection indices and the predictability of extratropical storminess indicators during the winter season in the North Pacific and Alaskan regions. Chapter 7 will present a conclusion.
Chapter 2 Defining a Storm

As discussed in Chapter 1, ETCs generate high impact weather across much of the North Pacific region. When conducting investigations into the differing aspects of a storm, the means by which an ETC is identified and, potentially, tracked is dependent on the underlying assumptions made to define a storm system. The way that the occurrence of a storm is defined is one of the driving questions behind the differences that exist both inter- and intra- method. Mesquita (2009) provided a framework for the perspectives by which most storm identification is categorized. Three possible categories of these perspectives exist; a theoretical perspective, a pragmatic perspective, and a tracking perspective. Ideally, it would be prudent to define a storm by a single measure. However, the reality is that the way a storm is defined is highly dependent on the application and needs of the researcher or stakeholders. Each perspective is further expanded upon in the following subsections.

2.1 Theoretical Perspective of Storm Definition

Storm identification from a theoretical perspective is rooted in the basic meteorological theory of formation and evolution of ETCs. As such, this perspective is highly dependent on “traditional” definitions of a storm from the meteorological community, including variables such as mean sea level pressure (MSLP) and relative vorticity (ζ). This perspective is grounded in synoptic and dynamic meteorological theory, with a dependence on baroclinic instability to generate and maintain storms. This perspective is highly useful for research based in concepts such as storm formation, storm intensity, and others. The concepts of vorticity (ζ) and baroclinic instability will be further discussed in section 2.4.
An explanation of how a cyclone works is needed to understand how MSLP and relative vorticity are applied in these objective methods. The atmosphere, when classified by its density, is commonly referred to in two primary ways: barotropic or baroclinic (Holton, 2004). A barotropic atmosphere is one in which the density is determined solely by pressure. In general, the atmosphere is considered to be barotropic in tropical regions of the Earth. In contrast, a baroclinic atmosphere is one in which the density depends on both temperature and pressure. Baroclinic atmospheric conditions are found in the mid-to high-latitudes of the Earth. For the purposes of ETCs, a baroclinic atmosphere is needed to generate vertical wind shear that drives the necessary large scale vertical motions. Baroclinic instability, which can be defined by how perturbations (such as storms) draw energy from the overall mean flow, is generally driven by the existence of a strong jet core aloft above a strong meridional (North-South) temperature gradient, generating shear. Warm, western ocean currents supply areas of sharp land-ocean temperature difference near the surface of the earth, enhancing local baroclinicity (Hoskins and Valdez, 1990). With this baroclinic enhancement and the existence of the Pacific and Atlantic jets, ETC formation and dissipation regions are organized into common areas, represented by the entrance and exit regions of the climatological storms tracks. In general, the storm tracks are most active in the fall and winter seasons due to both local and large scale temperature differences near the surface (land-ocean) and aloft (between very cold polar regions and relatively warm mid-latitudes) (Holton, 2004; Mesquita et al., 2010).

Along with baroclinic instability, some amount of rotation or spin in the atmosphere is needed for cyclogenesis to occur and the cyclone to evolve. This spin,
called vorticity, is defined mathematically as the curl of the velocity vector (Holton, 2004):

\[ \omega = \nabla \times U \]

In the atmosphere, vorticity can be defined as either absolute (\( \eta \)) or relative (\( \zeta \)), where absolute vorticity is the sum of the planetary vorticity (\( f \)) and relative vorticity:

\[ \eta = \zeta + f \]

where

\[ f = 2\Omega \sin \phi \]

Relative vorticity is the relative portion of the vorticity, with \( f \) removed (i.e. the amount of spin in the atmosphere with the effect of the Coriolis force removed):

\[ \eta = \frac{\partial v}{\partial x} - \frac{\partial u}{\partial y} + f, \quad \zeta = \frac{\partial v}{\partial x} - \frac{\partial u}{\partial y} \]

The use of relative vorticity to identify and track cyclones is possible due to the necessity of the presence of a local relative vorticity maximum for the genesis and evolution of an ETC. Additionally, the local maxima of relative vorticity are generally found at the centers of cyclones, further allowing for it to be used as a measure for identification and tracking of cyclone activity.

2.2 Pragmatic Perspective of Storm Definition

The pragmatic perspective is guided primarily by “storm” impacts, relying on the expression of a storm in terms of how it may be observed. In this case, “storms” are viewed by their manifestation at the surface, such as high wind events. Thus, this perspective relies on methods that capture the expressions of storminess as experienced by a person on the ground.
Methods that definition of a storm from this perspective are reliant on a more Eulerian (or semi-Eulerian) approach to describing storm activity. This perspective can be, in general, the easiest for many stakeholders to describe as it connects the tangible elements of the manifestation of a storm (e.g. storm surge, wind speed, coastal erosion) to the occurrence of an event. For observers, this perspective can often prove more useful than others, as storm impacts are not always directly associated with a low pressure system and can be displaced from the center of an ETC. Atkinson (2005) presents a method which is an example of evaluating storm activity within a pragmatic context. This method uses the local wind event frequency and duration to serve as proxies for storminess at multiple locations in a very practical and threshold based approach. Other Eulerian based methods have been applied in previous research to determine overall storm tracks and storm track characteristics. For example, several studies have used bandpass filtering of atmospheric fields at synoptic timescales, often between 2 – 10 days, allowing for the understanding of large scale atmospheric dynamics from a pragmatic perspective (Blackmon, 1976; Blackmon et al., 1977; Barry and Carleton, 2001; Anderson et al., 2003). Further examples of Eulerian methods, including Atkinson (2005) focusing on the North Pacific, will be provided in Chapters 3 and 5.

2.3 Tracking Perspective of Storm Definition

A third perspective for identification and tracking of ETCs is termed a “tracking perspective” (Mesquita, 2009). This is an expansion of the theoretical perspective, by adding a tracking component to storm identification. This focuses less on storm lifecycle mechanics (i.e. how the storm forms or dissipates) and more on storm pathway distributions. Many unique storm identification and tracking methods exist. Common to
all objective methods, though, is the use of thresholds for distinguishing and resolving individual storm systems. Most objective methods are designed to emulate how an objective human observer (e.g. a meteorologist) might manually identify and track a storm.

The general approach of the objective method to identify and track ETCs is as follows. First, an atmospheric field of interest (e.g. MSLP) is selected. The field that is selected will influence the application (and results) of an objective method. Objective methods have been applied to a wide range of atmospheric fields, ranging from surface based fields (e.g. MSLP) to upper level fields (e.g. 250-hPa omega). Second, local maxima/minima are identified within the field, which are presumed to indicate discrete “storms.” Third, the method searches subsequent time-steps for the storm within the atmospheric field. This process is also complex, as linking events between time steps requires additional thresholds for many cyclone-specific parameters, such as the maximum distance that a storm can travel or the strengthening or weakening rate of the cyclone.

2.4 Previous Research on Extratropical Cyclones

ETCs are well-studied phenomena that occur throughout the middle to high latitudes of both the Northern and Southern hemispheres. ETCs form and move in preferred areas of the globe, often related to areas of maximum local baroclinic instability. The determination of storm tracks has been previously well defined in the literature. In general, the positioning of major storm tracks is controlled, in part, by the processes described in section 2.3 (Hinman, 1888; Blackmon, 1976; Blackmon et al., 1977; Hoskins and Hodges, 2002; Hoskins and Hodges, 2005; Bengtsson et al., 2006).
Within the North Pacific region, ETCs tend to follow a track stretching from the east coast of Asia across the North Pacific Ocean into the Gulf of Alaska (Mesquita et al., 2010). Some of the most intense cyclones in the North Pacific Ocean basin originate as tropical cyclones, with initial formation found in the warm waters of the tropical Eastern North Pacific Ocean. These storms track west across the Central Pacific Ocean, and eventually recurve to the north and east, transitioning to ETCs before traversing the North Pacific Ocean basin (Graham and Diaz, 2001). This track can diverge into other, secondary regions of track clustering within the North Pacific region, including the northern Bering Sea and Bering Strait (Mesquita et al., 2010). Once reaching the Gulf of Alaska, ETCs tend to stall, and thus this region represents the primary area of cyclolysis, or ETC “death”, in the North Pacific Ocean (Mesquita et al., 2010). Additional secondary cyclolysis centers occur over the coast of British Columbia and south to the Olympic Peninsula (Martin et al., 2001; Bengtsson et al., 2006; Mesquita et al., 2010). Most storms that follow the main Pacific storm track move towards North America, eventually dissipating in areas off the western coastlines of the continent (Martin et al., 2001). Though the surface low may dissipate, the potential vorticity signature of cyclones carries over the mountains, helping to trigger new storm development in the North American plains (e.g. Alberta Clippers, Colorado Low).

When juxtaposed with the North Pacific, polar regions are not a preferred area of cyclogenesis due to the entrenched areas of cold air and lack of local baroclinic sources (Sepp and Jaagus, 2010). As such, many high latitude areas are, therefore, viewed as net importers of ETCs rather than formation regions, with an increase in the frequency of “deep” cyclones being noted in the last 50 years (Sepp and Jaagus, 2010). Some recent
research (e.g. Wang et al., 2012) provides evidence that the reliability of trends found in the region is questionable, due to inhomogeneity resulting from changes in observational networks and data assimilation during the last 50 years. Other types of storm activity, such as polar lows and mesoscale cyclones, tend to be very small and short-lived and are not well understood in terms of their mechanics and impacts in the North Pacific. Studies analyzing the impacts of these smaller scale storms have been conducted in the North Atlantic (Condron et al., 2006; Condron and Renfrew, 2012), but are lacking in the North Pacific and Alaskan regions.

Previous studies have found changes in North Pacific cyclone activity since the mid-20th century. Graham and Diaz (2001) showed an increase of nearly 50% in the frequency of “strong/deep” cyclones after 1950 using the NCEP1 reanalysis. Similar to the Sepp and Jaagus findings, the reliability of this increase is questionable due to the homogeneity of the reanalysis, particularly due to a jump at the beginning of the satellite era of observations in 1979. However, some more recent work showed an increase in the frequency of explosive ETCs in the Northwest Pacific and a decrease in the number of overall cyclones in the 1979 – 2011 period using the Japanese 25-yr Reanalysis (JRA-25) (Iwao et al., 2012). At the same time, the lowest MSLP recorded in these storms decreased by about 5 hPa, indicating an intensification of the strongest storms within the region. Seasonal variability of storm tracks is greatest between the winter and summer and least between the spring and autumn (Mesquita et al., 2010). In general, Pacific sea surface temperatures (SST) are not a major control of cyclogenesis in the Gulf of Alaska and Bering Sea (Mesquita et al., 2010), implying dynamical processes such as local baroclinicity likely are a major factor in ETC formation over the North Pacific.
2.5 Teleconnections and Cyclone Activity Impacts

Physical drivers of low frequency climate variability can be considered in two ways. First, variability can be interpreted as a linear combination of a few dominant physical modes, such as the El Niño/Southern Oscillation (ENSO), the North Atlantic Oscillation (NAO), and the Pacific North American pattern (PNA) (Franzke and Feldstein, 2005). An alternate view of climate relates variability as a series of low frequency patterns, each described by their own specific temporal and spatial structure. These patterns are reflected by the top principal components of low frequency variability, as calculated by rotated empirical orthogonal function (EOF) analysis (Franzke and Feldstein, 2005; NOAA, 2014).

Many large-scale climate patterns are known to influence weather in the study region. Previous research has shown links between storm tracks and teleconnections such as ENSO, the Pacific Decadal Oscillation (PDO), PNA, North Pacific Index (NP), and Arctic Oscillation/ Northern Annular Mode (AO/NAM) (Overland and Pease, 1982; Zhu et al., 2007; Rodionov et al., 2007; Mesquita et al., 2010). Changes to spatial patterns of cyclone frequency and storm track density are often viewed as modifications in the large scale flow (i.e. global and synoptic atmospheric wave patterns). A brief description of teleconnections with influence on the North Pacific and their linkages to storm track variability follows.

The El Niño/Southern Oscillation is a coupled ocean-atmosphere interaction in the equatorial Pacific that has been shown to have global climate patterns. ENSO has three primary phases: a warm phase (El Niño), a neutral phase, and a cool phase (La Niña). El Niño events involve the warming of the tropical eastern Pacific, particularly in
the Peru region, which leads to a weakening of the typical sea surface temperature gradient across the equatorial Pacific. The Southern Oscillation, an east-west oscillation in surface atmospheric pressure across the equatorial Pacific, responds with changes in trade winds and precipitation patterns (Philander, 1983; IPCC, 2012). El Niño events typically peak in the boreal winter and occur with a periodicity of 3 to 7 years, alternating with the neutral and cool phases of the oscillation.

The locations of wintertime mid-latitude jet streams have been correlated with the different phases of ENSO and shifts in the location of the primary storm track in the Pacific. In El Niño years, an eastward and equatorward shift of the Pacific storm track is typically observed (Eichler and Higgins, 2006). Additionally, the Pacific jet tends to split, resulting in two storm tracks during these years: The first, stronger track is directed toward California while the second, weaker track is directed poleward towards Alaska and the Arctic (Eichler and Higgins, 2006). In La Niña years, the storm track tends to be directed poleward (Eichler and Higgins, 2006).

The Pacific Decadal Oscillation is the leading EOF of monthly sea surface temperature anomalies over the North Pacific (Mantua and Hare, 2002). The positive phase of the PDO shows warm anomalies in the eastern Pacific and cold anomalies in the western Pacific. The negative phase provides the reverse of this scenario. Within the Alaskan region, an equatorward shift of the location of the Aleutian low is evident in the positive phase, which correlates well with above average temperatures and precipitation for much of the Pacific Northwest (Mantua et al., 1997; Mantua and Hare, 2002; Zhu et al., 2007). The reverse is true in the negative phase of the PDO. PDO and ENSO reflect similar spatial patterns and influence storm tracks in similar ways (Mantua and Hare,
2002), with the temporal scale of PDO phases lasting between 20 – 30 years versus the 6 – 18-month phase duration of ENSO events (Mantua and Hare, 2002).

The Pacific/North American pattern is a Northern Hemisphere extratropical oscillation focused between four centers of action at the 500 hPa level (Wallace and Gutzler, 1981). These centres are located over Hawaii, the North Pacific, Alberta, Canada, and the Gulf of Mexico. PNA can be characterized by a standing wave pattern between these four centres of action. In the positive phase, PNA features an enhanced wave amplitude, leading to a strong ridge over western Canada, a deeper than normal Aleutian Low, and blocking episodes along the North American west coast. The negative phase of the PNA is a standing wave of lower amplitude, leading to more zonal flow at the 500 hPa level and a weaker Aleutian Low. PNA has been tied to North Pacific storm activity with some success; Gulev et al. (2001) reported correlations as high as 0.74 between the PNA and storm activity within the North Pacific.

The North Pacific index is the area-weighted mean sea level pressure over the region between 30 - 65°N latitude and 160 - 220°E (Trendberth and Hurrell, 1994). The primary use of the NP index is to depict the changes in intensity of the Aleutian low, with low values of the index indicating a stronger Aleutian low and higher values of the index indicating a weaker Aleutian low. Strong ties between the SSTs in the North Pacific and tropical Pacific are generally found, with a three-month lead of tropical Pacific SSTs on those in the North Pacific.

The Arctic Oscillation, also known as the Northern Annular Mode (NAM), is defined as the first EOF of Northern Hemisphere winter mean sea level pressure data in the region between 20N – 90N (Thompson and Wallace, 2000). The AO can characterize
changes in the position and strength of the mid-latitude jet stream. Positive phases of the AO typically correlate well with a poleward shift of the Pacific storm track. Negative phases of the AO typically correlate with an equatorward shift of storm tracks and often a high latitude blocking pattern over the Alaska and North Pacific regions. AO is typically linked with a regional manifestation of the AO in the North Atlantic known as the North Atlantic Oscillation (NAO) such that the AO and NAO often share the same phase (Hurrell and Deser., 2009).

Some previous studies have used combinations of the phases of teleconnection indices to determine their correlation with repeated tracking of severe ETCs over the same region, also known as seriality (Mailier and Stephenson, 2006), in an attempt to determine how one possible definition of “storminess” is associated with teleconnections (Seierstad et al., 2007). Within this study, the PNA was found to be the teleconnection with the highest correlation with extratropical storminess, as measured by MSLP, in the North Pacific. In addition to the PNA and PDO, the position of the primary Pacific storm track is also correlated with other teleconnections such as the West Pacific Oscillation (WPO) and the East Pacific Oscillation (EPO).

Other work has suggested that multiple teleconnections have some correlation with temperature and precipitation patterns in the North Pacific and Alaska (Bienek et al., 2012). Significant correlations at the 5% or stronger level between many teleconnection indices and station temperatures during fall and winter months were found in a study of the climate divisions of Alaska (Bienek et al., 2012). In particular, the Gulf of Alaska, showed significant correlation between temperatures and each teleconnection indices of the AO, NP, PDO, Niño-3.4, the Southern Oscillation (SO), and PNA in the winter
months. In the fall months, the East Pacific/North Pacific Oscillation (EP/NP), NP, PDO, and PNA each showed some influence on the subdivisions of the Gulf of Alaska. On the west coast (i.e. along the Bering Strait), PNA, EP/NP, and PDO showed the largest association with temperatures in the fall months, but not in the winter months. In the Aleutians, AO and PNA, along with PDO each have association with the temperatures in the climate zone. Given that temperature changes in the fall and winter months in these regions are primarily driven by cyclones (given that heat transport into the region is large scale and, especially in winter months, generated mostly by advection and not diurnal heating cycles), it is reasonable to expect these teleconnections to correlate well with storm activity for these same climate zones. More on this will be presented in Chapter 6.
Chapter 3 Methodology

This chapter provides an overview of objective ETC tracking methods – Lagrangian and Eulerian – and the main statistical techniques used for seasonal prediction in this dissertation. Various analytical methods are available that can objectively identify cyclone locations. A variety of meteorological variables are used, combined with various decision rule sets, as the basis for determining a “storm” (e.g. mean sea-level pressure, relative vorticity, geopotential height). Many of these methods were developed to simulate or reproduce manual methods of cyclone identification and tracking. The meteorological variables used in objective approaches are arrayed in regular spatial grids, which are drawn from numerical models of the atmosphere. When operated over a span of time, an objective approach provides a database of cyclone occurrence.

The performance of an objective cyclone identification and tracking method is a function of the underlying conditions specified within the method. These conditions are directly related to the gridded field and data projection. Scheme specific parameterizations, such as intensity cutoff or temporal filters, also affect the results from a method. The majority of objective approaches utilize a Lagrangian (or Semi-Lagrangian) perspective; that is, identifying and tracking cyclones as they move within the overall atmospheric flow. This approach encompasses both the theoretical and tracking perspectives described in Chapter 2. The alternative pragmatic approach encompasses most Eulerian methods, which analyze gridded fields from a site specific, hemispheric, or global perspective at time scales specific to synoptic activity to gain information about pattern variance. For example, wind speed could be analyzed on a grid point-by-point basis to determine when it is high or low, with no consideration of its
geographical context - there is no storm that is tracked with these approaches. Eulerian methods are less popular in current literature outside of a few studies related to patterns of storminess in the circum-Arctic (Atkinson, 2005), but were widely used in the mid 20th century to distinguish storm tracks and storm track statistics from a hemispheric perspective (e.g. Blackmon, 1976; Blackmon et al., 1977). Historically, a methodological shift towards objective Lagrangian methods began in the late 1980s with the advent of large numerical model and reanalysis datasets. For example, the work of Lambert (1988) generated a global cyclone climatology using a general circulation model from the Canadian Climate Centre through cyclone counting, an Eulerian-based approach. Soon after this, other objective identification and tracking methods, such as those of Murray and Simmonds (1991) and Serreze et al. (1993), were developed to research the patterns and trends of cyclone activity from a hemispheric or regional perspective. Each of these studies developed their own independent cyclone identification algorithm.

Over time, the Lagrangian algorithms for cyclone detection and tracking have been updated and enhanced. For example, the Murray and Simmonds (1991a, 1991b) algorithm has been changed slightly to incorporate additional techniques for tracking cyclones. These changes include improvements to resolution and filtering of systems, in addition to incorporating an improved “probability of association” algorithm for cyclone tracking from the first time step to the next. The cyclone identification and tracking algorithm developed in Serreze et al. (1993) was updated in both the work of Serreze (1995), Serreze et al. (1997), and Wang et al. (2006) to account for some issues with cyclone exclusion and edge effects in the dataset.
For the purposes of this chapter, a short review of two objective methods of cyclone identification is presented and one method of storm identification from the pragmatic perspective. In the past 10 years a number of method comparison efforts have been undertaken. More recently, method comparison studies were conducted by the IMILAST group (Neu et al., 2013; Simmonds and Rudeva, 2014); in this case, a cross-method comparison using the same input dataset for all of the tracking methods was conducted. However, this study did not include the Hodges method due to concerns from Kevin Hodges about the experimental design (K. Hodges, personal communication). The methods of Serreze and Hodges will be discussed in greater detail in the remainder of the chapter.

The Serreze method is highlighted within the work for this study due to its popularity and use in the Alaska region. In the United States, the Climate Prediction Center operates the Serreze method operationally to identify and track cyclones in the MSLP field, making analysis of the dataset of interest for this study. Additionally, the Hodges method is selected due to its versatility and ability to analyze cyclones in different atmospheric fields. Mesquita et al. (2010) used the Hodges method to look at the climatology of cyclones in the North Pacific and Alaskan regions. For the sake of comparison, these two methods provide two possible ways of identification of a storm. The remainder of this chapter will further explore the two methods.

3.1 Objective Identification and Tracking Methods

The question of how to objectively identify cyclones is approached in very different ways depending on the algorithm. Often, the method starts with a precise definition of a storm event. Each algorithm discussed below uniquely identifies a
“storm”. Raible et al. (2008) show differences in cyclone climatologies produced when different cyclone detection algorithms are applied to the same reanalysis datasets. More recently, the Hodges method was used to analyze several recently released reanalysis datasets, including the 25-year Japanese Reanalysis (JRA-25), the NASA Modern Era Retrospective-Analysis for Research and Applications (NASA MERRA), the ECMWF Interim Re-Analysis (ERA-I), and the NCEP Climate Forecast System Reanalysis (NCEP CFSR) (Hodges et al., 2011). When the Hodges method was applied to each of these reanalyses and analyzed on a T-42 grid, the higher resolution reanalyses showed agreement between counts and spatial distribution of cyclones. They did not tend to find as much agreement with older, lower resolution reanalyses, such as NCEP1. Reanalyses with finer spatial resolution, such as the CFSR and ERA-I, exhibited more frequent occurrence of intense cyclones than the lower resolution datasets like the JRA-25.

Cyclone detection is dependent on the criteria specified within a scheme, the dataset being analyzed, and the parameter being used as the cyclone proxy (Greeves et al., 2008). Small changes in any of these specifications can have large impacts on the results. Walsh (2004) notes that, although there are multiple objective means of cyclone identification, there is a lack of objective means of specifying parameters needed in the methods. This issue is the main reason for the spread that exists between the findings of different methods, as this causes varying objective definitions of how a “storm” might be identified and tracked. In this dissertation, three particular methods - Serreze, Hodges, and Atkinson - are used (Table 3.1).

Most cyclone identification and tracking methods perform similar steps to analyze gridded fields. In general, objective methods begin by identifying feature points for each
time step within the dataset. The actual representation of the feature is dependent on the method. For example, the Serreze method uses MSLP minima, while the Hodges method often uses relative vorticity maxima. The feature detection can be best described as a binary method (yes/no) for if a feature does or does not exist at a grid point. Linkages between feature points at successive time steps are made through methods such as nearest neighbor checks (Serreze) or energetic cost functions (Hodges), which allows for the tracking of features that were identified. Finally, cyclone and storm track statistics are calculated by most methods. More specifics about the Serreze and Hodges methods are given in the following sections.

3.1.1 The Serreze Method

The cyclone identification and tracking method developed in Serreze et al. (1993, 1997) and Serreze (1995) analyzes MSLP fields. When applied to the NCEP1 reanalysis, the initial check of the algorithm performs a search for MSLP minima that are at least 2-hPa lower than the surrounding eight grid points. Modern, higher resolution reanalyses would require far more grid points to cover the same spatial area. This check is performed to determine if an MSLP minimum value is enclosed by at least one isobar, meaning a low needs to be “closed” in order to be counted. The 2.0 hPa threshold has been changed in multiple studies using the Serreze method (Wang et al., 2006). If the initial check is failed, the algorithm expands its search outwards to 3 grid points from the cyclone center. In the case of the Serreze (1995) study, grid points from 800 km out to 2400 km away are searched when establishing a closed cyclone. Previous versions of this algorithm had required a minimum value of central pressure for the identification of a cyclone (Serreze et al., 1993). However, this requirement was removed in S95 due to the
importance of the background pressure field in specifying the strength and location of a low pressure system. This acknowledges the importance of pressure gradient in relation to the strength of a ETC rather than only identifying strong storms by applying a specific threshold as a cutoff value, as is seen in studies like Graham and Diaz (2001) and Sepp and Jaagus (2010), among others.

In the Serreze algorithm, cyclone tracking through time is performed using a nearest neighbor approach. This assigns a unique identifying number to each cyclone identified on a given time step \((t)\). At the subsequent time step \((t + 1)\), each previously identified cyclone location is matched with a \(3 \times 3\) array of grid points outward from the center location at the previous time step \((t - 1)\). In the Serreze (1995) study, another parameterization specified is the cyclone search distance between time steps; this is set to a maximum of 1400 km, or no more than 10 degrees of latitude to the south in the Northern Hemisphere, and must exhibit an absolute sea level pressure change of no more than 40 hPa in 12 hours. These values are specific to the Serreze (1995) study and are often adjusted in studies which utilize the Serreze method for cyclone identification and tracking (Wang et al., 2006). If these criteria are not met, the identified cyclone is determined to be a new cyclone. If a cyclone was not matched during this process, it is considered dissipated and no longer tracked.

Many criteria specified in the Serreze (1995) method were shown to have direct impacts on the sensitivity of results. For example, changes made to the detection threshold, set to 2 hPa in Serreze (1995), result in higher or lower numbers of cyclones detected. In the Serreze (1995) method, a smaller detection threshold leads to the detection of a larger number of weaker cyclones in the same dataset. The reverse is also
true. For example, in Wang et al. (2006), a detection threshold of 1.0 hPa produced far fewer cyclones than a threshold of 0.2 hPa utilized on the same dataset. Changing this parameter targets weak cyclones in particular, which are often lost with a higher detection threshold value. The Serreze method results are highly sensitive to the reanalysis being used in the study, as grid resolution plays directly into the 3 x 3 grid point matrix used for matching in the future. For example, Wang et al. (2006) modified the Serreze method to handle higher resolution reanalyses by expanding the number points of initial grid search from 3 x 3 to 7 x 7.

Other examples of potential weaknesses in the Serreze method were shown by the Mesquita et al. (2009) study. This study performed a comparison of the performance of two tracking algorithms – Serreze (1995) and that of Murray and Simmonds (1991a, 1991b) - examining their performance alongside a manual tracking result for a particular severe Bering Sea cyclone in October 1992. The Serreze method broke the storm into two separate events when the low weakened while the Murray and Simmonds method described one, continuous track for the storm. The discrepancy occurred primarily because the Serreze method “counts” a storm only as long as it exhibits a closed center. If a storm re-strengthens and closes again, it is then counted as a new storm. Comparatively, the Murray and Simmonds method allows low-pressure systems to become temporarily open as they are being tracked. This reduces the chance of counting a single cyclone more than once as storms at high latitudes can undergo rapid cyclogenesis, cyclolysis, and can exhibit re-strengthening. Such issues have implications for storm frequency assessments in terms of trends they might be exhibiting due to climate change.
3.1.2 The Hodges Method

The Hodges (1994, 1995, 1999) method of tracking cyclones differs from the Serreze method in that it is capable of tracking features using any continuous variable in a gridded field. Most commonly, this method has been used to track cyclones in the 850 hPa relative vorticity field. The distinct steps for the Hodges method are shown in Table 3.2. The Hodges method determines maxima or minima within a gridded field that are clustered closely to determine the center of mass of a feature, which is then identified as a storm. This method allows for the analysis of both areas of maximum and minimum relative vorticity values. Features are tracked by using a cost function to determine the path of “least resistance” (i.e. lowest energetic cost) in the next time step. This cost function, therefore, determines the most likely location of the feature at the next time step. If a feature is found where the cost function determines the most likely location will be at the next time step, the original feature attributes are carried over to the next time step and assigned to the new feature.

The Hodges method was initially developed to use a simple cylindrical projection to determine locations and tracks (Hodges, 1994). In Hodges (1995), this method was transformed to handle feature tracking using spherical coordinates projected to the unit sphere. Typically, the Hodges method analyzes relative vorticity maxima and minima, which allows it to track tropical and extratropical cyclones in the Northern and Southern Hemispheres. Data fields are spatially filtered to the T42 spatial resolution (~300km) to reduce the impact noise in the background field. More information on this filtering is provided in the following sections.
The Hodges method filters storm tracks lasting less than two days or that travel less than 5 degrees in a great circle arc, which is approximately 500 km. Tracking and statistics for the Hodges method are computed on a spherical grid, which reduces the influence of latitude on the results. A wide array of storm-related statistics are generated by the Hodges algorithm, including storm track density and cyclogenesis and cyclolysis density, among others.

The Hodges method is often preferred in many studies due to its versatility. Examples of such studies include Hoskins and Hodges (2002, 2005), which applied the Hodges method of identification and tracking to global reanalysis datasets for detailed analysis of Northern and Southern Hemisphere storm tracks and trends. Bengtsson et al. (2006) use the Hodges method to investigate the effects of the IPCC A1B climate change scenario on future storm tracks when compared to the storm tracks determined from the European Centre for Medium-Range Weather Forecasting (ECMWF) ERA-40 dataset. Both of these studies used the Hodges method to identify features in the low-level relative vorticity field at 850 hPa. It is noted in Bengtsson et al. (2006) that the ability to use other fields for the analysis of cyclone tracks and trends exists, but it was determined that the relative vorticity field is better for identification of cyclones due to the lack of vertical extrapolation effects on the field. However, the noise level of the data when looking at 850 hPa relative vorticity is considerably higher than is seen in MSLP or z1000 studies, primarily due to the dataset being on the higher frequency end of the synoptic scale and thus representing features of synoptic scale and those of high frequency features outside of the interest of the tracking algorithm. (Bengtsson et al., 2006). Therefore, the gridded field often requires the removal of background noise.
through spatial filtering from initial reanalysis resolution to a lower spatial resolution, typically T-42 (~300 km). This concept will be further expanded upon in the following sections.

3.1.3 The Atkinson Method

The Atkinson (2005) algorithm was designed to identify strong wind events that produced waves and/or surges that were of sufficient magnitude to impact a coastal region, defined by setting a wind speed threshold. In Atkinson (2005), ‘impacts’ means conditions that damage habitats or infrastructure or could potentially perform geomorphological work. Storms with winds above 10 ms\(^{-1}\) that last for approximately 6 hours were the criteria established in this study, consistent with thresholds used in previous studies of coastal impact (e.g. MacClenehan et al., 2001, Solomon et al., 1994). In addition to the physical basis for the 10ms\(^{-1}\) threshold, the standard is maintained to allow for comparison between previous studies and the results from this iteration of the algorithm. MacClenehan et al. (2001) set their speed threshold with reference to one of the more powerful storms to reach the Irish coast, although the selection of a final threshold was also influenced by the need to have a reasonably populated event database. Event duration threshold was set at six-hours, standard at many weather forecasting offices and within the aforementioned studies. Further details on the Atkinson method are discussed in Chapter 5.

3.2 Overall Weaknesses of Objective Identification

3.2.1 The Correspondence Problem

One of the factors that affects storm identification and tracking schemes is the ability to correctly identify and track a cyclone from time step to time step. Hodges
(1994) terms this problem as the “correspondence problem.” Objective tracking schemes are highly dependent on both the spatial and temporal resolution of the dataset as tracked features may appear or disappear between time steps. If the temporal resolution of the analyzed data field is too coarse, feature tracking between time steps becomes more difficult as the spread of possible “next” locations for the feature increases with the passage of time. Ideally, the position of the tracked feature at the next time step could be checked at all possible positions throughout the dataset. This solution is often not possible due to computational limitations, as it requires an extension of the search to every possible grid point for every identified feature. Similar to the difference in definition of a “storm” between methods, each algorithm provides a unique solution to the problem. For example, one technique that has been used is to associate climatological velocity statistics and recorded previous movement with cyclone positions to probabilistically forecast the new position of the cyclone (Murray and Simmonds, 1991).

Serreze et al. (1993, 1997) and Serreze (1995) address the correspondence problem by associating each cyclone with a three by three grid box and performing a nearest neighbour search within that box at the next time step. This approach is simple and imposes a constraint of a maximum distance traveled between time steps that can be adjusted depending on the temporal resolution of the data set. This constraint assumes that a storm can only move so far in a single time step of the data. One potential problem with this approach is the case where an extreme event or fast moving cyclone moves faster than the method allows and is either missed in the next time step or identified as a new storm. However, there are potential solutions to this double counting problem. One is to set the maximum distance constraint abnormally high to not miss any cyclones in the
next time step. Another is to conduct an analysis of the dataset at a finer temporal resolution and specify a smaller maximum distance to achieve greater correspondence between time steps in the dataset.

Hodges (1994, 1995, 1999) approaches this issue by assuming attributes about the smoothness of motion of a feature. By doing this, a cost function determines the smoothest possible track for a feature to move. This function prevents sudden changes to feature tracks by placing constraints on the value of the cost function keeping identified features from making unrealistic spatial movements across the grid between time steps and allows for higher temporal correspondence. As with other methods, finer temporal resolution reduces the likelihood a storm track abruptly ends or disappears from the search grid.

One of the main weaknesses facing all objective methods is the inability to identify secondary cyclogenesis and associate the correct storm attributes with the new storm. While secondary cyclogenesis process is occurring (e.g., a “triple-point low”), the original cyclone is often weakening but still identifiable in a MSLP field. At the same time, the new secondary cyclone is developing in a geographically close location to the original cyclone. On an analysis, this may have the appearance of a “dual-lobed” cyclone, with two cyclone centers (Serreze, 1995). When manually analyzed, an observer can be trained to identify the secondary cyclone as a new storm and count it correctly. Objective methods, however, can have problems with this process due to temporal resolution, grid resolution, or detection parameters specified within the scheme. Attributes of the original cyclone can be incorrectly associated to the triple point low, resulting in an unrealistic track jog to the south/east and/or extending the track of the original cyclone past the true
area of cyclolysis (Wernli and Schwierz, 2006). This can be caused by the time step being large enough that the secondary cyclone develops before the next iteration of the scheme. It can also be caused by the tracking scheme limitations related to the parameters specified in the scheme. If a cyclone is expected to move somewhere within a fixed distance from its previous location and secondary cyclone develops in that area, the secondary cyclone could assume the attributes of the original low. Due to this, undercounting cyclones becomes a potential issue.

The issue of secondary cyclogenesis is directly addressed in many of the above studies. Serreze (1995) addresses this issue by noting the potential for the identification of a new low manifesting as the original low due to similar spatial location. Additionally, a second issue can occur in the Serreze method where two systems are located and share the same closed isobar. This situation leads to the merging of the two systems on the next chart and the cyclone closest to the new location being selected as the new cyclone. There is no specific way of eliminating the problem in the Serreze method when this situation occurs, however infrequent.

Dependent on the grid field, there can be a number of problems that need to be addressed. Schemes that employ MSLP fields can experience issues associated with topography and temperature. The act of extrapolating surface pressure to sea level creates issues over complex topography, specifically in areas of higher altitude terrain. Due to these topographical influences on the sea level pressure field, spurious cyclones are often observed (Simmonds et al., 1999). These “cyclones” can develop in areas where ground surface temperatures are warmer than that of the normal ambient air at the same altitude. The opposite can happen when a ground surface is colder than the ambient air, as
spurious anti-cyclones can be found in this situation (e.g. Greenland). Hoskins and Hodges (2002) experienced these issues with both MSLP and 850 hPa relative vorticity fields over the Tibetan plateau and Himalayan Mountains. These topographic features caused issues in analysis by generating spurious cyclones, such that the region was ultimately rejected for inclusion in the final analysis. Many schemes that use MSLP or z1000 fields, such as Murray and Simmonds (1991a, 1991b) and Blender et al. (1997), address this issue by ignoring terrain with a height of over 1 km. While this criterion eliminates the existence of spurious cyclones generated in final analysis, it could cause issues where an analysis is particularly interested in lee cyclogenesis or features over higher terrain. Simmonds et al. (1999) also noted that the same problems occur in areas of steep or complex topographic slope, regardless of height. The elimination of high altitudes leaves large holes in hemispheric analysis, particularly in mountainous regions like the Himalayas, Antarctica, and some regions of Alaska.

Secondary to the problems with vertical extrapolation, the temporal and spatial scaling of the feature is an important consideration. Depending on the purpose of the study, the frequency of tracked features in the synoptic range can be important. Methods that use MSLP as the identification field generally find fewer systems than those using relative vorticity (Greeves et al., 2006). This occurs because MSLP features are considered to be on the low end of the power spectra when compared to features identified in relative vorticity fields (Hoskins and Hodges, 2002). In directly comparing methods using MSLP and relative vorticity fields, Greeves et al. (2006) found that the number of features identified was between 20-50% higher using 850 hPa relative
vorticity than when using MSLP fields. There are other factors, such as the selection of dataset and scheme specific parameters, which impact this result as well.

Ultimately, though, the observation that 850 hPa fields produce greater numbers of storms than using MSLP seems robust: the finding of Greeves et al. (2006) also showed higher track densities using 850 hPa relative vorticity for the western North Pacific than using when MSLP. This emphasizes the need to consider the parameter or storm definition being used to track cyclones when drawing conclusions about future estimates of storms and storminess. Greeves et al. (2006) does note that, although the frequency of events seen in a relative vorticity identification and tracking method is much higher than in a MSLP based identification and tracking method, the primary storm track locations remain similar in both methods. This fact, amongst others previously stated, re-emphasizes the need for a clear definition of a “storm.” Therefore, when contrasting studies of cyclone frequency and intensity, it is important to use methods that have employed the same gridded field or have used a consistent method for determining storm event databases.

Tracking methods that use relative vorticity often experience issues with high-resolution data due to noise from the background field, which can be defined as the spherical harmonic expansion of the relative vorticity field with a wave number less than 5 (Bengtsson et al. 2006). Blender et al. (1997) disregarded maxima in relative vorticity fields, as high-resolution data are too detailed and noisy to determine features similar to cyclones. Hodges (1994, 1995, 1999) uses lower resolution data fields to identify relative vorticity maxima. Alternatively, Hoskins and Hodges (2002, 2005) and Bengtsson et al. (2006) remove the background field before identification and tracking of features in order
to more accurately track the maxima and minima in the relative vorticity field as positive and negative anomalies, respectively. This improves the method’s ability to identify the features of interest, especially in lower tropospheric data fields (Bengtsson et al., 2006). Another issue that faces these methods is grid resolution. Pinto et al. (2005) and Raible et al. (2008) showed that the total number of cyclones and cyclone strength identified in a study are directly related to grid resolution. An objective scheme is more likely to identify a greater number of “features” when grid resolution is higher. The “features” identified are often more intense than in lower resolution data and can be tracked for a longer time period than in coarser resolution datasets. This idea is backed by Greeves et al. (2007), who showed that detected features were more frequent and more intense in a higher horizontal resolution grid than in a low horizontal resolution grid.

3.3 Statistical Prediction Methods

Classical statistical predictions of climate variables often begin with the use of some form of linear regression. In the simplest sense, linear regression (LR) relates an independent predictor variable (X) with a dependent predictand (Y). For simple regression, the relationship between the predictor and predictand that produces the least error is used to create a single straight line. Simple regression can be expanded by allowing for multiple predictors, where a coefficient is calculated for each predictor to estimate the linear relationship between each X in the presence of other predictors and the predictand Y. This process is known as multiple linear regression (MLR).

Not all linear regression models provide useful results, particularly when dealing with inherently non-linear systems or when the assumptions of the residuals taking a Gaussian form are “untenable” (Wilks 2011). Non-linear regression and modeling
methods in the climatological realm provide a secondary series of methods for statistical prediction. Generalized Linear Models (GLMs) are a means of non-linear regression that have previously been used in studies of teleconnection impacts on storminess, including seasonal extratropical storminess and tropical cyclone predictions (Elsner and Schnertmann, 1993; Elsner et al., 2001; Mailier et al., 2006; Seierstad et al., 2007). GLMs allow for the fitting of regression models to distributions from the exponential family, including normal, Poisson, and gamma, among others. GLMs are comprised of three primary elements: a linear predictor, a link function, and a response variable distribution (Meyers et al., 2010). The link function connects the linear predictor to the response variable through its natural mean, where the link differs based on the underlying assumption of the distribution of the response data. For example, in the case of the normal distribution, the linear predictor is related to the response through regression coefficients ($\beta$) as applied to each individual predictor ($x$) as shown in equation 3.1.

$$\eta_i = \beta_0 + \beta_1 x_1 + \beta_2 x_2 + \ldots + \beta_i x_i$$  \hspace{1cm} 3.1

The link function is known as the identity link, where the mean ($\mu$) is related to the response variable as shown in equation 3.2.

$$g(\mu_i) = \mu_i$$  \hspace{1cm} 3.2

The second form of statistical prediction used within this thesis is that of random forest ensemble regression (RF). RF is a form of non-linear regression analysis that operated using machine learning that is often compared other methods like neural networks (Cutler, 2013). RFs are based on classification and regression tree analysis (CART), which allow for the model to be run without formally specifying a link function. Variables used to create a fit are chosen automatically within the algorithm and vary.
between individual versions of the regression trees, known as weak learners. While CART algorithms are sensitive to small changes in the input data (a small change in data often changes the regression greatly), RFs are much more stable, primarily due to the combination of many solutions through an ensemble of regression trees. While a single tree may be sensitive, the forest is more stable due to the combination of many trees.

In this dissertation, RF analysis trains learners on bootstrap samples of the dataset through a process called “bagging” and combines them through averaging of regression trees (Dietterich, 2000; Brieman, 2001). This process is repeated multiple times to create many series of regression trees. The process improves upon individual CART algorithms by increasing output accuracy and increasing stability (less sensitivity) with respect to the data input. This means that the methods should be more robust with respect to cross-validation if there is legitimate skill in the analysis.

There are some advantages to the random forest method that make it advantageous to use in this work. First, unlike GLMs, random forests require no formal distributional assumptions to create a data fit, however, other assumptions tied to the bootstrapping method are made. The method also can automatically fit highly non-linear interactions, which allows for better modeling of non-linear processes. Accuracy of Random Forests is comparable to other machine learning methods (Cutler, 2013). However, the complexity of an ensemble of regression trees makes interpretation of the data difficult, particularly when looking for individual influences of specific predictors. The two statistical methods outlined here are used for winter season prediction of storminess indicators in the study area, and comparisons between the methods are made in Chapter 6.
Table 3.1. Three methods for extratropical storminess identification. Hodges and Serreze methods identify and track cyclones while the Atkinson method identifies storm events based on wind patterns.

<table>
<thead>
<tr>
<th>Method Reference</th>
<th>Variable Analyzed</th>
<th>Terrain Filtering</th>
<th>Temporal Constraint</th>
</tr>
</thead>
<tbody>
<tr>
<td>Hodges (1994,1995,1999)</td>
<td>850 hPa Relative Vorticity ($\zeta$)</td>
<td>&gt;1000 m</td>
<td>2-day minimum</td>
</tr>
<tr>
<td>Serreze (1995)</td>
<td>Mean Sea Level Pressure (MSLP)</td>
<td>none</td>
<td>2-day minimum</td>
</tr>
<tr>
<td>Atkinson (2005)</td>
<td>10m wind 0.995 sigma wind</td>
<td>none</td>
<td>6 hr minimum for wind events</td>
</tr>
</tbody>
</table>
Table 3.2. Hodges TRACK algorithm steps for identification and tracking of features, as applied within this study

<table>
<thead>
<tr>
<th>Step Name</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>Selection</td>
<td>User-defined specific region and thresholds</td>
</tr>
<tr>
<td>Segmentation</td>
<td>Divides the region into distinct areas, labels all points uniquely</td>
</tr>
<tr>
<td>Identification</td>
<td>Looks for points that have intensities larger than the user-defined thresholds - points below threshold are called ‘background points’</td>
</tr>
<tr>
<td>Matrix</td>
<td>Creation of binary feature maps, which are then converted to <em>hierarchy levels</em></td>
</tr>
<tr>
<td>Boundary Points</td>
<td>Takes care of ID feature points that are entering and leaving the region of interest - points re-filtered and labeled</td>
</tr>
<tr>
<td>Feature Detection</td>
<td>Identifies suitable points within the binary maps for tracking the storms</td>
</tr>
<tr>
<td>Correspondence</td>
<td>Links suitable points from first to second time frame, using a cost function</td>
</tr>
</tbody>
</table>
Chapter 4. Climatological Patterns of Cyclone Activity in the North Pacific and Alaskan Regions using the Twentieth Century Reanalysis (20CR)

Article Information

Chapter 4 consists of a manuscript submitted to the Journal of Climate. Figures and tables are the same as those submitted within the paper but have been renumbered for thesis consistency. References have been reformatted for thesis consistency.

Author Information

*N. J. Shippee¹

D. E. Atkinson²

¹N. J. Shippee, University of Victoria Climate Lab, Department of Geography, University of Victoria, PO Box 3060 STN CSC, Victoria, British Columbia, Canada, V8W 3R4

²D. E. Atkinson, University of Victoria Climate Lab, Department of Geography, University of Victoria, PO Box 3060 STN CSC, Victoria, British Columbia, Canada, V8W 3R4

*Corresponding author email: nshippee@uvic.ca

Author Contributions

Shippee developed the study and hypothesis, ran the tracking algorithms and computed the associated statistics, conducted the analysis, created the figures and tables, and wrote the manuscript. Atkinson provided the server time, reviewed and edited the manuscript.

Abstract

This study presents: 1) a comparison of two storm identification and tracking algorithms; and, 2) an evaluation of the general characteristics of extratropical cyclones in two
reanalyses for the North Pacific and subregions. First, two popular cyclone identification methods, one MSLP-based and one 850 hPa relative vorticity based, are applied to the NCEP/NCAR Reanalysis 1 (NCEP1). Results from this comparison show that the general patterns of cyclone characteristics are similar between the two methods, though with increased values of cyclogenesis density, cyclolysis density, and track density when using the relative vorticity based method. There is strong inter-method agreement of trends (both direction and magnitude) exhibited in the two methods’ datasets for the 1950 – 2010 period, with only one case where opposing statistically significant trends ($\alpha = 0.05$) are identified between the methods (Alaska Interior sub-region summer). Second, a comparison between storm tracks generated by the 850 hPa relative vorticity based method for NCEP1 and the 56-member ensemble of the Twentieth Century Reanalysis v2 (20CR) is conducted. There are distinct differences found between the 20CR and NCEP1 mean climatology for both cyclogenesis and cyclolysis density in storm formation and dissipation regions (east coast of Japan and Gulf of Alaska, respectively), with a strong positive bias in NCEP1 cyclone characteristics. Comparisons between the meridional and zonal wind components at 850 hPa from the reanalyses shows that this bias is likely related to a strong positive bias in wintertime zonal winds in NCEP1 with respect to 20CR. Additionally, inspection of time series of cyclone characteristics within the region shows homogeneity in the 20CR is found in the Gulf of Alaska and Bering Sea subregions after 1920. Poleward subregions, such as the Chukchi/Beaufort Sea and the Alaska Interior, are impacted by inhomogeneities into the mid 20th century and are not found to provide a reliable extension of the NCEP1 climatology.
4.1 Introduction

Extratropical cyclones (ETCs) have the potential to generate severe weather over the North Pacific, Alaska, and its adjacent seas. ETCs are of high interest in this region due to their ability to control both large scale and local weather processes such as precipitation, wind, and temperature change. These storms are often characterized by their ability to generate widespread impacts, including potentially dangerous sea states, coastal flooding and erosion (Blier et al., 1997, Lynch et al., 2008), and wind damage (Atkinson, 2005). Coastal zones possessing shallow littoral shelves are vulnerable to inundation and erosion due to wave action and storm surges, particularly during times when a protective sea ice buffer is absent (Mason et al., 1993; Blier et al., 1997). Such coasts typify much of the western and northern Alaskan coasts, as well as most of the Russian and western Canadian Arctic coastal zones.

In general, storms that impact the North Pacific and Alaska do not form locally, but rather move into the region from cyclogenesis areas farther to the west (Graham and Diaz, 2001; Mesquita et al., 2010). The North Pacific is a very active ETC region, exhibiting a storm track stretching from the east coast of Asia across the ocean basin into the Gulf of Alaska (Mesquita et al., 2010). The most intense cyclones in the North Pacific Ocean basin, however, often originate as tropical cyclones that generally form in the warm waters of the tropical Eastern North Pacific before tracking west and eventually “recurring” to the north and east (Simmonds and Keay, 2002; Mesquita et al., 2010). This recurring path often leads to the transition from tropical cyclones to ETCs before continuing across the North Pacific (Graham and Diaz, 2001). It is not uncommon for
late-life-cycle ETCs to re-intensify with assistance from favorable upper-atmosphere patterns within this region.

Most storms traversing the North Pacific tend to track towards the North American continent, eventually dissipating in areas off the western seaboard (Martin et al., 2000). In particular, ETCs often tend to stall and decay in the Gulf of Alaska, representing the primary area of cyclolysis, or ETC “death”, in the North Pacific (Mesquita et al., 2010). Other secondary centers of storm cyclolysis are found along Vancouver Island (Martin et al., 2000; Mesquita et al., 2010) and the central coast of British Columbia, and the Olympic Peninsula (Bengtsson et al., 2006; Mesquita et al., 2010). The North Pacific storm track exhibits secondary regions of track clustering, including into the northern Bering Sea and Bering Strait region, as well as into the Sea of Okhotsk, where the highlands of Kamchatka also tend to trap ETCs.

Cyclone impacts are a function of several elements. The first stems from the particular nature of the cyclone event and the general expanse of a given cyclone. Second is the concept of preconditioning – impacts from storms can often become more severe when multiple storms track over the same area in a short period of time, what can be termed “serial” cyclones. In these situations, any given storm in the sequence need not be particularly powerful, but the cumulative action of multiple, “moderate” storms can equal or exceed that of a single big storm (Mailier et al., 2006). Third, and particular to northern coastal regions, is timing of storm occurrence – whether or not storms occur during the presence of a protective buffer of coastal sea ice, which can constitute land-fast ice or a berm of piled-up slush or block ice (Hufford and Partain, 2004; Robinson,
Land-fast ice or slush berms help mitigate ETC related impacts such as storm surge and coastal erosion, particularly on the west coast of Alaska.

There are numerous previous studies that have examined various aspects of ETC activity and trends on a hemispheric basis (Hodges et al., 2003; Hoskins and Hodges, 2005; Bengtsson et al., 2006; Ulbrich et al., 2009; Hodges et al., 2011; Neu et al., 2013). One of the most recent studies is the community effort to inter-compare extratropical cyclone detection and tracking algorithms (IMILAST; Neu et al., 2013). IMILAST investigated the differences in a wide array of objective cyclone identification and tracking methods and investigated the implications of the choice of a specific method from a global to hemispheric perspective (Neu et al., 2013). Overall, relatively few studies have focused specifically on the North Pacific and Alaskan regions (Serreze et al., 1993; Graham and Diaz, 2001; Zhang et al., 2004; Mesquita et al., 2010; Simmonds and Rudeva, 2014). Some research has been conducted concerning storm track patterns and trends in the North Pacific, Bering Sea, and Alaskan regions using the Hodges (1994; 1995; 1999) cyclone tracking algorithm (Mesquita et al., 2008; Mesquita et al., 2010). However, there are a large number of objective storm tracking algorithms that exist and few studies that have directly contrasted the performance of different objective tracking methods using the same dataset in the North Pacific and Alaskan region other than to investigate specific storms (Mesquita et al., 2009).

Given the prevalence of storm activity and resultant impacts in the North Pacific and Alaskan regions and the limited storm research focus in this region, this study has two main objectives:

2. Determine by how much a cyclone climatology can be realistically extended before the start of NCEP1 for the study area using a longer term reanalysis, the Twentieth Century Reanalysis version 2 (20CR; Compo et al., 2011).

Objective 1 will compare the climatologies created using two objective tracking algorithms, applied to NCEP1 data. NCEP1 is used for two reasons: First, NCEP1 possesses a long record (60+ years) and grid spacing similar to 20CR. Second, the Serreze cyclone tracking algorithm is used operationally at NOAA’s Climate Prediction Center (CPC) and is run with NCEP1 to generate near real-time cyclone tracking for the Alaska region. Objective 2 will compare the climatologies from the 20CR and NCEP1 reanalyses generated using the Hodges objective tracking algorithm.

This paper continues with an overview of the data, algorithms, and the methodology used for comparing climatologies in Section 2. Section 3 presents results from each tracking method, the intercomparisons of the data, and the 20CR cyclone climatology. Section 4 discusses the results. A summary of the key findings is provided in Section 5.

4.2 Data and Methodology

4.2.1 Reanalysis Data

Two datasets are used in this study: NCEP-NCAR Global Reanalysis 1 (NCEP1; Kalnay et al., 1996) and the Twentieth Century Reanalysis version 2 (20CR; Compo et
al., 2011). NCEP1 assimilates surface, upper-air, and satellite data into the reanalysis using a T-62 (~209 km) horizontal resolution model with 28 vertical layers. 20CR differs from NCEP1 in that it only assimilates surface pressure observations and uses the 56-member Kalman ensemble assimilation system (Compo et al., 2011). The underlying model for the 20CR possesses many of the same characteristics as NCEP1, namely a T-62 horizontal resolution with 28 vertical layers. Differences between the data assimilation methods of the two reanalyses, particularly the reliance on only surface observations within the 20CR when compared with NCEP1, make the resultant storm tracks calculated from 20CR sensitive to changes in observational networks. Use of a wider range of data sources in NCEP1, such as upper air data and satellite data post 1979, make the reanalysis less sensitive to changes in any one ingest data source, though the effects of the addition of satellite data after 1979 also affect reanalysis homogeneity. Within the North Pacific and Alaskan regions, observational networks have undergone extensive change over the 140-year period covered by the 20CR.

Using the 20CR ensemble mean to evaluate cyclone activity utilizing tracking algorithms is not prudent for a few reasons (Wang et al., 2012). First, using the ensemble mean can possibly make the location of the cyclone center move further than what specific distance thresholds or limiting factors of a tracking algorithm, mainly due to variability within the ensemble members. This effectively causes undercounting of cyclones when using solely the ensemble mean. Second, Wang et al. (2012) showed that variability between ensemble members, particularly in the early portion of the 20CR period, make the mean conditions vary greatly at the 6-hourly temporal scale when compared to the 6-hourly variability within each individual ensemble member. For this
reason, the storm tracking algorithm is run on each of the 56 ensemble members, after which statistical properties storm tracks of each are calculated and, finally, averaged.

Global 6-hourly mean sea level pressure (MSLP) fields, spanning from 1948/49 – 2010, are used in the Serreze tracking algorithm. NCEP1 6-hourly u and v wind components at 850-hPa are used to calculate relative vorticity in the Hodges tracking algorithm. Global 6-hourly u and v wind components at 850-hPa from each member of 20CR ensemble are also used to calculate relative vorticity for the Hodges tracking algorithm, from 1871 – 2010.

4.2.2 The Serreze Identification and Tracking Algorithm

The Serreze (1995) cyclone identification and tracking method is configured to work with MSLP fields. The algorithm searches for minimum values in the MSLP field that are at least 2 hPa lower than the surrounding eight grid points. This check determines if a minimum value is enclosed by at least one isobar that is 2 hPa higher than the minimum value. Due to this, the Serreze method requires a pressure center to contain a closed isobar in order to be identified as a cyclone. If a closed cyclone is not found in the initial eight surrounding grid points, the algorithm will continue to search outward from the minimum value up to three grid points outward from the cyclone center. If a closed contour is found, the center of this area is then recorded as a cyclone. The Serreze method uses a nearest neighbor approach to track cyclones by assigning each storm a unique identifying number. At the following time step, a search for each cyclone occurs within an array of grid points, typically as far as 3 points in each direction, outward from the cyclone center location at the previous time step. Cyclones that are found within the new grid are considered a continuation of the original storm, and those that end are no longer
counted. Between successive time steps, cyclones are allowed to move a maximum of 1400 km, no more than 10 degrees of latitude, and have an absolute value of sea level pressure tendency (SLPT) change of up to 40 hPa in 12 hours. If any of these criteria are not met, a new identifier is attached to the cyclone and the tracking of the previous cyclone is terminated.

4.2.3 The Hodges Identification and Tracking Algorithm

The Hodges (1994, 1995, 1999) algorithm can utilize any variable in a gridded field to track cyclones. Typically, the 850-hPa relative vorticity field is used to find maxima, which are used as a proxy for surface cyclones due to the dynamical dependence of MSLP cyclones on low to mid-level vorticity. The Hodges method determines maxima or minima within a gridded field (e.g., relative vorticity) that are clustered closely to locate the centre of a feature, which is then identified as a storm. Features are then tracked using a cost function to determine the path of lowest energetic cost in the next time step. This cost function, therefore, determines the most likely location of the feature at the next time step. If a feature is found at the location suggested by the cost function, it is considered a continuation of the cyclone and its attributes are carried over to the next time step.

For the analysis of the 20CR ensemble members, individual storm tracks are matched between each of the 56 ensemble members of the 20CR using matching utilities from the Hodges algorithm. For this, spatial and temporal constraints are placed on tracks when matching. Mean separation between tracks in the ensemble members must be less than 5 degrees (~500 km) and must overlap temporally for at least 60% of their points. A
relaxation of these constraints, particularly the temporal matching requirement, would increase the number of tracks.

4.2.4 Study Area and Climatology Specifics

The study area includes the North Pacific, Bering Sea, and Alaskan regions (Fig. 4.1; Table 4.1). Additionally, sub-regions describing the Gulf of Alaska, Bering Sea, Chukchi and Beaufort seas, and interior Alaska are also utilized (Table 4.1). These sub-regions were defined to focus more specific attention on known areas of greater or lesser storm activity, allowing for separate calculation and comparison of trends between sub-regions. For example, the Gulf of Alaska is well understood as an area into which many mature-phase ETCs track (Simmonds and Keay, 2002; Mesquita et al., 2010). When compared to the Beaufort/Chukchi Seas region, the Gulf of Alaska is an area of much more frequent ETC activity for all measures of cyclone activity used in this study.

Cyclogenesis density, cyclolysis density, and storm track density are the measures used for the assessment of primary storm activity for this study. Output from the Serreze (NCEP1) and Hodges (NCEP1, 20CR) methods are analyzed over the entire study area and for each individual sub-region. Mean seasonal climatologies and trends are shown for each sub-region for select time periods. Seasons are defined in 3-month periods: January to March (JFM, winter), April to June (AMJ, spring), July to September (JAS, summer), and October to December (OND, fall).

4.3. Results

4.3.1. Serreze and Hodges NCEP1 Comparison

The patterns of cyclogenesis density in both the Serreze and Hodges track methods show distinct areas of storm initiation in the study region (Fig. 4.2 and 4.3,
respectively). In both algorithms, lee-side cyclogenesis regions are apparent on the eastern side of the Rocky Mountains in North America, with the maximum values of cyclogenesis density approaching 8 or more storms per year in this region (units: storms \((10^6 \text{ km}^2 \text{ season})^{-1}\)). Baroclinic cyclogenesis regions are found along the Kuroshio current boundary off the east coast of Japan, across the North Pacific, and into the southern Bering Sea. These regions persist throughout all seasons, though with varying intensities that peak in the winter (6 – 8 storms) and fall (3 – 5 storms) (Figs. 4.2a., 4.3a. and 4.2d., 4.3d., respectively). Spurious heat lows, particularly over the eastern Eurasian continent, are visible in the Serreze method in the warm season but not in the Hodges technique (Figs. 4.2c, 4.3c, respectively).

Cyclolysis density patterns distinctly show the Gulf of Alaska as the main center of cyclone “death” in the study region in both the Serreze and Hodges methods (Fig. 4.4 and 4.5, respectively). In general, 8 or more decaying storms (units: storms \((10^6 \text{ km}^2 \text{ season})^{-1}\)) are found in this region. This large center is also surrounded by other regions of relatively high cyclolysis densities, particularly in the Bering Sea and over the coastal regions of British Columbia. The Serreze method shows an extension of higher cyclolysis densities from the main region in the Gulf of Alaska into the Bering Sea and Norton Sound regions, with 4 – 6 storms per season. This extension exists in the Hodges method as well, though with a lower magnitude in all seasons (3 – 5 storms). Similarly, in both methods a secondary area of cyclolysis is found in the Sea of Okhotsk, particularly in the fall season (Figs. 4.4d., 4.5d.). This region of higher density is larger in the Serreze method (5 – 8 storms) than the Hodges method (3 – 5 storms). Additional regions of
cycolysis are evident in the Serreze method nearest the eastern edge of the study area over the Rocky Mountains; these regions are not as apparent in the Hodges method.

ETC track density (units: storms (10^6 km^2 season)^{-1}) peaks in the winter and fall (Figs. 4.6a., 4.7a. and 4.6d., 4.7d., respectively). In the fall season, both the Hodges and Serreze results show primary storm activity centers over the Gulf of Alaska and into the Aleutian Islands (Figs. 4.6d., 4.7d.). The maximum track densities (25 – 45 storms) in the fall and winter seasons are found in the Bering Sea, Aleutian Islands, and into the Gulf of Alaska. In both methods, all seasons show a large number of storms throughout the North Pacific, with the majority located along an axis stretching between 45-50°N latitude (Figs. 4.6 and 4.7). A smaller area of high track density stretches into the Bering Strait, most notably in the Hodges dataset (Fig. 4.7). Overall, compared to the fall and winter seasons, both methods indicate there are far fewer storm tracks occurring in the spring season across the study area (Fig. 4.6b., 4.7b.); this is particularly apparent in the Serreze method for the Gulf of Alaska sub-region. Furthermore, the summer season shows the fewest number of storm tracks in the study area in both methods (Figs. 4.6c., 4.7c.). When compared directly, overall track density values are greater in the Hodges method than the Serreze method year-round.

4.3.2. Hodges NCEP1 and 20CR Comparison

4.3.2.1 Cyclogenesis

Many of the prominent features found within the NCEP1 storm track climatology are also found within the 20CR storm track climatology. The two most prevalent areas of cyclogenesis found in NCEP1 are also found in the 20CR climatology (Fig. 4.8). The first is over the North Pacific off the coast of Japan, occurring in response to the baroclinic
gradient and intensified by the Kuroshio current. The second is over the continental interior of North America in association with lee-side cyclogenesis on the eastern side of the Rocky Mountains. Across the North Pacific, areas of maximum cyclogenesis shift equatorward in the winter. The strongest baroclinically generated cyclogenesis regions are found during the winter and fall seasons (Fig. 4.8a, 4.8d, respectively).

The differences between NCEP1 and 20CR cyclogenesis density (NCEP1 – 20CR values for the 1950-2010 period, units: storms (10^6 km^2 season)^{-1}) is highlighted regionally (Fig. 4.10). Greater values of cyclogenesis density using 20CR (up to +2.5 storms) are apparent in the Rocky Mountains in all seasons while NCEP1 values are higher (+1 to 2.5 storms) in the central North Pacific, south of the Aleutian Islands, over Japan, and into the Bering Sea. A more southerly shift of the baroclinically generated centers of cyclogenesis in the winter and fall is apparent within the 20CR climatology when compared with the NCEP1 results (Fig. 4.10a and 4.10d, respectively). In general, some of the greatest differences between the 20CR and NCEP1 cyclogenesis results are located in regions of complex topography and persist throughout all seasons.

4.3.2.2 Cycolysis

NCEP1 and 20CR both indicate the primary region of cycolysis is located in the Gulf of Alaska, with up to 8 storms per season in winter and fall. Both methods also identified secondary maxima of cycolysis densities extending into the Bering Sea (4 – 5 storms) and along the coastal regions of British Columbia (5 – 6 storms), particularly in the winter and fall seasons (Fig. 4.9a and 4.9d, respectively). The winter and fall seasons have higher cycolysis densities than the spring and fall (Figs. 4.9a, 4.9d, and 4.9b, 4.9c, respectively).
The maximum differences in cyclolysis density (NCEP1 – 20CR values for the 1950-2010 period, units: storms ($10^6$ km$^2$ season)$^{-1}$) are found in the primary “centers of action” such as the Gulf of Alaska and coastal British Columbia. Differences are particularly evident in the winter, summer, and fall (Fig. 4.11), where 20CR exhibits approximately 2 - 2.5 more storms than are identified in the NCEP1 climatology, and on the order of 2.5 storms or more along the British Columbia coastline. The NCEP1 climatology shows higher values of cyclolysis density than 20CR during winter and spring in mainland China and the southwestern United States (+1.5 to 2 storms).

4.3.3 Sub-regional Storm Characteristics in the 20CR

Time series of the frequency of cyclogenesis and cyclolysis in the study sub-regions were analyzed for the 1871 – 2010 period of the 20CR climatology. Individual ensemble member time series of annual cyclogenesis and cyclolysis frequency in each subregion were evaluated for inhomogeneities using RHTests version 4 (Wang et al., 2010). This analysis uses changepoints are defined as a shift in the mean of the time series that is statistically significant at the $\alpha < 0.05$ level. In this section, only “type-1” changepoints, which are statistically significant changepoints detected in situations that lack supporting metadata, will be discussed (Table 4.2). Examples of homogeneity testing results are shown for the Gulf of Alaska subregion (Fig. 4.12) and the Chukchi/Beaufort subregion (Fig. 4.13).

The cyclogenesis time series for the Gulf of Alaska sub-region has no significant change points within the 1871 – 2010. In the cyclolysis time series, two significant changepoints are found; one in 1910 and one in 1939. All changepoints within the Gulf of Alaska sub-region occur before the beginning of the NCEP1 time series.
Similar to the Gulf of Alaska, a relatively small number of changepoints are detected within the Bering Sea sub-region cyclogenesis and cyclolysis frequency time series. In the cyclogenesis series, two significant changepoints are identified occurring in 1899 and 1910. Within the cyclolysis series, two similar change points are found, one in 1882 and one in 1907. Following the 1910/1907 changepoint, the overall time series for both cyclogenesis and cyclolysis are homogeneous.

The cyclogenesis time series for the Chukchi/Beaufort sub-regions shows three significant change points, with the largest change detected in 1935 and associated with a significant change in the time series mean. The final cyclogenesis changepoint in this sub-region is found in 1974. The trend of the time series following each identified changepoint is a decrease number of cyclones being generated in the region. The cyclolysis time series has six significant changepoints, with three occurring post 1950 (1952, 1966, and 1977).

Within the Alaska Interior sub-region, there are two significant changepoints in the cyclogenesis time series, with the most recent of these occurring in 1992. Though not detected by the RHTests algorithm, a distinct change in the time series, particularly with respect to interannual variability, is observed between 1930 – 1940. Five changepoints are found within the Alaska Interior cyclolysis series, all occurring before 1946. Similar to the Chukchi/Beaufort sub-region, each changepoint is followed by a decreasing trend in the number of cyclones ending in the sub-region, including from the 1946 to 2010 period. Further discussion of the findings of the breakpoint analysis can be found in section 4.5.


4.3.4 Time Series and Trend Analysis

Time series trends for seasonal cyclogenesis and cyclolysis density within sub-regions are investigated for different time periods using a non-parametric Mann-Kendall test (Mann, 1945; Kendall, 1975; Gilbert, 1987). For NCEP1, three time series are examined: the extended period, and mid- and late 20th century periods (1950 – 2010, 1950 – 1978 and 1979 – 2010, respectively). For 20CR, trends are analyzed for the same periods as for NCEP1, with two additional periods for the first half of the 20th century: a) 1920 – 2010, b) 1920 – 1949, c) 1950 – 2010, d) 1950 – 1978, and e) 1979 – 2010. Periods c, d, and e allow for direct comparison with NCEP1. All trends are expressed as percent change per year (%yr\(^{-1}\)) over the period analyzed. Trends significant at the 5% significance level (\(\alpha = 0.05\)) are discussed. All trends for the Serreze and Hodges NCEP1 analysis are summarized in Table 4.2. All trends for the Hodges 20CR analysis are summarized in Table 4.3.

4.3.4.1 NCEP1 Trends

In the Gulf of Alaska sub-region, there were four significant trends found in the 1950 – 2010 period in the Hodges method (Table 4.2a). Cyclolysis increased in the fall (1.73%) and the winter (0.96%) and cyclogenesis decreased in the winter (-0.75%) and summer (-0.87%). In the 1950 – 1978 period, a single significant trend was exhibited: an increase of cyclolysis in the fall (1.33%). In the 1979 – 2010 period, three significant trends were identified. Cyclolysis was found to have increased in the winter (1.15%) and fall (1.68%), while cyclogenesis decreased in the summer (-1.93%). In the Serreze method (Table 4.2b), two significant trends were exhibited in the Gulf of Alaska for the 1950 – 2010 period. Cyclogenesis decreased in the winter (-0.53%) and in the spring (-
0.73%). During the 1950 – 1978 period, two significant trends are exhibited. Cyclolysis decreased in spring (-1.97%) and summer (-1.70%). One significant trend was found in the 1979 – 2010 period: an increase in cyclolysis in the fall (1.98%).

Using the Hodges method for the Bering Sea sub-region, the only significant trend exhibited in the 1950 – 2010 period was a decrease in cyclolysis (-0.33%). The 1950 – 1978 and 1979 – 2010 periods showed no significant trends for either cyclolysis or cyclogenesis. There were no significant trends found in the Serreze method analysis of the Bering Sea sub-region.

In the Hodges method analysis, the Alaska Interior sub-region exhibited one significant trend in the 1950 – 2010 period: an increase of cyclogenesis in the summer (1.09%). No significant trends were found in the 1950 – 1978 period. In the 1979 – 2010 period, four significant trends were identified. Decreases in cyclolysis were found in the winter (-1.21%) and spring (-2.32%) and decreases in cyclogenesis were found in the winter (-1.46%) and fall (-1.18%). Using the Serreze method, three significant trends were found for the Alaska Interior sub-region during the 1950 – 2010 period. Cyclolysis showed a decrease in the spring (-0.35%) and cyclogenesis showed decreases in the spring (-0.66%) and summer (-0.53%). No significant trends were found in the 1950 – 1978 and 1979 – 2010 periods.

Using the Hodges method, the Chukchi/Beaufort sub-region exhibited one significant trend in the 1950 – 2010 period: an increase of cyclogenesis in the winter (0.70%). In the 1950 – 1978 period, a single significant trend was identified: a decrease of cyclolysis frequency in the spring (-1.48%). There were no significant trends found in the 1979 – 2010 period. Using the Serreze method, two significant trends were found in
the 1950 – 2010 period: an increase of cyclolysis in winter (1.34%) and a decrease of cyclolysis in the summer (-0.57%). In the 1979 – 2010 period, a decrease of cyclolysis in summer (-2.33%) was also observed. No significant trends were identified for the 1950 – 1978 period using the Serreze method.

4.3.4.2 Twentieth Century Reanalysis (20CR) Trends

Using the Hodges method (Table 4.3), the Gulf of Alaska sub-region exhibited three significant trends during the 1920 – 2010 period. Cyclolysis increased in the winter (0.45%) and summer (0.43%) and cyclogenesis increased in the fall (0.86%). The 1920 – 1949 period shows one significant trend: a decrease of cyclolysis (-1.06%) in the spring. In the 1950 – 2010 period, only one statistically significant trend was identified: an increase of cyclolysis in the winter (0.57%). There are no significant trends in the 1979 – 2010 or 1950 – 1978 periods for the Gulf of Alaska.

The Bering Sea sub-region exhibited three significant trends during the 1920 – 2010 period. Cyclolysis and cyclogenesis both exhibited a decreasing trend in the summer (-0.27% and -0.68%, respectively) and, in the winter, cyclogenesis showed an increase (0.35%). In the 1920 – 1949 period, there are four significant trends identified: decreases in cyclolysis in the winter, summer, and fall (-0.90%, -0.85%, and -1.15% respectively) and decreases in summer cyclogenesis (-1.54%). In the 1950 – 2010 period, the only significant trend identified is observed in the winter as an increase in cyclogenesis (0.81%). The period 1950 – 1978 exhibited two significant trends: a decrease of cyclogenesis in the summer (-1.70%) and an increase in the fall (5.58%). There was only one significant trend identified in the 1979 – 2010 period: a decrease of cyclolysis in the fall (-0.88%).
For the Alaska Interior sub-region, significant trends for the 1920 – 2010 period are identified in nearly every season. In the following, several values were flagged as significant that are almost certainly are not representative of any natural variations and are discussed in more detail in section 4. Increasing trends of cyclolysis are found in the spring (0.66%), summer (1.44%) and fall (0.45%) and increases in cyclogenesis is observed in all seasons, winter (1.58%), spring (1.27%), summer (1.33%), and fall (0.93%). In the 1920 – 1949 period, significant increasing trends are observed in cyclolysis in summer (7.30%) and in cyclogenesis in the winter (6.62%), spring (28.98%), summer (128.75%), and fall (7.51%). The largest observed trends are found during this time period for this sub-region. In the 1979 – 2010 period, cyclogenesis was found to have decreased in summer (-1.55%) and fall (-0.92%). No statistically significant trends were exhibited in the 1950 – 2010 or 1950 – 1978 periods.

A similar pattern to that found in the Alaska Interior sub-region is exhibited in the Chukchi/Beaufort sub-region, particularly in the early period trends of 1920 – 2010 and 1920 – 1949. Both cyclolysis and cyclogenesis show increasing trends in all seasons during the 1920 – 2010 time-period, with increases for the winter, spring, summer, and fall measuring 0.81%, 2.54%, 0.89%, and 0.42% for cyclolysis and 2.14%, 1.29%, 1.45%, and 1.45% for cyclogenesis, respectively. The 1920 – 1949 period also showed statistically significant trends for both cyclolysis and cyclogenesis in all seasons. Cyclolysis exhibited an increase in the winter (22.02%), spring (154.36%), summer (49.96%), and fall (9.03%) and cyclogenesis exhibited increases in the winter (86.50%), spring (27.14%), summer (62.48%), and fall (28.73%). In the 1950 – 2010 period, two significant trends are found. Cyclolysis exhibited a decrease in the summer (-0.57%) and
cyclogenesis exhibited an increase in the fall (0.98%). In the 1979 – 2010 period, two significant trends are found: in the winter, a decrease in cyclogenesis (-1.56%) and, in the summer, a decrease in cyclolysis (-1.83%). In the 1950 – 1978 period, two significant trends are also found: an increase in fall cyclogenesis (1.78%) and a decrease in summer cyclolysis (-0.75%).

4.4 Discussion

4.4.1 Serreze and Hodges NCEP1 Climatology Comparison

There is a large degree of inter-method agreement in the location and frequency of cyclones (Figs. 4.6 and 4.7), regardless of the gridded field used by the algorithm. The Gulf of Alaska, where a high proportion of storms are nearing the end of their lifecycles, has fewer cyclones identified in the Serreze method in the winter (Fig. 4.6a). In all other seasons, there are fewer storms in this region identified in the Hodges method (Fig. 4.7) than in the Serreze method (Fig. 4.6). Greater cyclogenesis densities are also found in the Serreze method over more complex terrain, such as the Canadian and American Rocky Mountains, when compared to the Hodges method in all seasons. The greatest track density differences between the two methods are evident in the secondary areas of storm activity. For example, near the coast of British Columbia, lower values of track density are exhibited in the Serreze climatology with respect to the Hodges climatology. Furthermore, a greater track density is observed in the Hodges method at nearly all locations. This is not unexpected and is a by-product of the atmospheric field being evaluated (i.e., relative vorticity) and the associated time scales of the perturbations for vorticity versus MSLP (Hoskins and Hodges, 2002; Anderson et al., 2003; Mesquita et al., 2010). In particular, because the temporal and spatial scales of vorticity-based
features are on the shorter and smaller end of the synoptic scale, it would be expected that
the frequency of events in the MSLP-based approach would be lower than for the
vorticity-based approach used in the Hodges method. As such, many of the smaller scale
features that are found in the Hodges track density climatology, such as secondary lobes
of activity near the coast of Vancouver Island (British Columbia) and in the Bering Sea
and Bering Strait region, are not always seen as prominently or consistently in the
Serreze method results, particularly in seasons with reduced cyclone activity. As such,
this difference in track densities is most evident in the spring and summer months with
higher track density in the Hodges method than Serreze method when overall ETC
activity wanes in the Northern Hemisphere (Figs. 4.6 and 4.7).

The minimum values for all storm activity metrics are observed in the summer
season. This result is independent of the cyclone tracking algorithm and gridded field
because, as noted by Mesquita et al. (2008), this is a typical pattern for ETCs during the
summer season, when local baroclinicity within the study region is at a minimum
compared to the fall and winter seasons. Nevertheless, the Gulf of Alaska remains the
center of cyclolysis in the North Pacific for all seasons, and summer reductions of
cyclolysis density are minimal when compared to the seasonal differences in track
density and cyclogenesis density, as lysis counts are similar in all seasons. Meridional
shifts of track density between the fall and winter seasons are also evident within the
study region. These shifts, particularly of track density equatorward in the winter, reflects
mid-winter suppression of storminess within the North Pacific that has been noted in
previous studies (e.g. Nakamura, 1992; Penny et al., 2010). Differences between the
Serreze and Hodges climatologies are most visible with higher values of the storminess
metrics within the study area. As noted by Hoskins and Hodges (2002), MSLP tends to be dominated by larger-scale features and biased towards slower moving systems when compared to relative vorticity, which is less influenced by the background flow and allows for earlier detection of cyclones. This leads to the expectation of higher counts of cyclogenesis and longer lived storms in the Hodges method when compared directly to Serreze method.

Sub-regional trends observed in the two cyclone identification and tracking methods are in relative agreement. Most notably, there is strong inter-method agreement of trends (both direction and magnitude) exhibited between the Serreze and Hodges datasets for cyclogenesis within the Gulf of Alaska sub-region in the winter season for the 1950 – 2010 period. There is only one case where significant trends are identified in each method, but have differing signs. This occurred in the Alaska Interior sub-region during the summer, where a positive trend is observed in cyclogenesis density in the Hodges method (1.09%) and a negative trend is found in the Serreze method (-0.53%). As these two methods are analyzing different atmospheric fields, it is not surprising to find trends of differing magnitudes and signs between the two datasets in at least one sub-region. Much like comparing individual ensemble members, conclusions extracted from the comparison of the two cyclone identification and tracking methods’ trend results are, to some extent, generalizations, as some discrepancies between the methods and resulting analyses should be expected due to differences in the model fields and filtering parameters used within each algorithm.

One caveat related to the trend analysis of both the Serreze and Hodges NCEP1 cyclone statistics is the overall homogeneity of the reanalysis due to the effects of
changing data sources. This is particularly evident in the pre- and post-1979 satellite era split, where the addition of additional data can lead to “positive” trends that are related to changes in input data rather than changes in physical storm tracks. Some previous studies that used NCEP1 as a basis for analysis have shown positive trends in strong cyclones in the North Pacific across the 1950–2000 period (Graham and Diaz, 2001) and in the Chukchi/Beaufort Sea region of the Arctic (Sepp and Jaagus, 2010), which can be questioned due to the changes in data over the period. As such, trends assessed for the full period (1950–2010) are more sensitive to the homogeneity of the reanalysis, and should be treated with caution.

4.4.2 NCEP1/20CR Hodges track comparison

The ability of 20CR to provide a reliable longer-term ETC climatology for the study area is found to be somewhat limited due to data homogeneity issues, echoing the findings of Wang et al. (2012). This is particularly evident in the sub-regions closest to the Arctic. The temporal extent of reliable storm climatologies is limited to, at best, a start point near 1920 in the southernmost reaches of the study area and, at worst, the mid-1940s in the Alaska Interior and Chukchi/Beaufort sub-regions. In the Gulf of Alaska and Bering Sea sub-regions, relatively homogeneous cyclolysis time series are evident after the 1920-1925 period in all seasons, thus allowing for much more reliable trends to be analyzed from this point onward (Fig. 4.12). In other sub-regions (e.g. Chukchi/Beaufort), the time series does not become homogeneous until much later in the mid-20th century due to the lack of observational data in these regions in the early part of the 20th century. As such, trends for these sub-regions are not necessarily reflective of the true climate system signals for the 1920–1949 and 1920–2010 periods.
Wang et al. (2012) identified the Arctic and Alaskan regions as data sparse areas, particularly in the early 20th century. The result of this is a more inhomogeneous time series of storms in the high latitude regions of the study area. In this study, this higher degree of inhomogeneity is also evident in the sub-regions that are located in the northern part of the study area, such as the Chukchi/Beaufort and the Alaska Interior sub-regions, when compared with the Bering Sea and Gulf of Alaska sub-regions. As is shown in section 4.3, homogeneity test results show that the frequency of cyclolysis in the Alaska Interior exhibits five significant changepoints. For the same metric in the Chukchi/Beaufort region, a total of six significant changepoints exist, with the most recent occurring in 1977. Within the southern-most sub-regions of the study region (i.e., the Bering Sea and Gulf of Alaska), there are fewer changepoints identified in the cyclolysis time series, with a single changepoint in the Bering Sea (1907) and two changepoints in the Gulf of Alaska (1910 and 1939). The minimal changepoints detected and being only present in the early 20th century allows there to be value added by the 20CR storm tracks to the sub-regional climatologies for the Bering Sea and Gulf of Alaska sub-regions, particularly in the Bering Sea where the time series is homogeneous after approximately 1910.

The longer homogeneous 20CR datasets available in many of the sub-regions of the North Pacific and Alaskan regions can be utilized to temporally extend regional storm track climatologies prior to ~1950, when the NCEP1 dataset begins. In sub-regions such as the Bering Sea and the Gulf of Alaska, longer term homogeneous datasets are available, allowing for approximately 2-3 decades of additional data when compared to NCEP1 by extending the time series back to the 1920s. However, within historically
data-sparse regions such as the Chukchi/Beaufort and the Alaska Interior sub-regions, additional years of homogeneous data from 20CR are only valuable back to the 1940s. As was shown by Wang et al. (2012), the majority of temporal inhomogeneities within 20CR exist for the Alaskan regions prior to 1949. This is reflected within the trend results of this study, particularly for the specific sub-regions (i.e., most Northern sub-regions: Alaska Interior, Chukchi/Beaufort) where data scarcity in the early part of the 20CR dataset highly impacts the internal variability observed within the ensemble members. As such, trends from the Alaska Interior and Chukchi/Beaufort sub-regions are not stable for the time periods evaluated in this study until, at the earliest, the 1950 – 2010 period. While significant at the $\alpha < 0.05$ level, the increasing trend is highly influenced by changes in observational capacity reflected within the 20CR for the Alaska Interior and Chukchi/Beaufort sub-regions over the entire period. Using reanalysis for trend assessment has been cautioned against in the past (Kalnay et al., 1996; Bengtsson et al., 2004; Thorne and Vose, 2010), particularly due to changes in observational networks and the resultant changes of overall data availability. This is somewhat evident in this study, at least from the perspective of changes to the overall observational network, as evidenced by dataset inhomogeneities. As such, the most reliable statistically significant trends are only visible in the Alaska Interior and the Chukchi/Beaufort sub-regions in the 1950 – 2010, 1979 – 2010, and 1950 – 1978 sections of the 20CR dataset, and only appear for a few seasons, and the true reliability of these trends can be questioned due to their sporadic nature.

As was shown in section 4.3, there are distinct differences found between the 20CR and NCEP1 climatologies for both cyclogenesis and cyclolysis density. One
potential driver of these differences is the prescribed sea surface temperature (SST) and ice used by the different reanalyses. NCEP1 uses various prescribed SST data sources including the Comprehensive Ocean-Atmosphere Data Set (CODAS) and the Hadley Centre Sea Surface Temperature dataset (HadSST3), while 20CR uses the modern Hadley Centre Sea Ice and Sea Surface Temperature (HadISST) data set. The main difference between the two is spatial resolution, with HadSST3 being about 5 times coarser than the newer HadISST dataset. This difference allows for a much higher resolution of the Kuroshio region and associated SST gradients and could be a main driver of the differences observed between 20CR and NCEP1 datasets, particularly in the case of cyclogenesis density. Additionally, a comparison between the meridional and zonal wind components at 850 hPa as output from the reanalyses shows a strong negative bias in wintertime 850 hPa zonal wind in 20CR with respect to NCEP1, particularly on the western coastline of Japan (Fig. 4.14). This positive bias helps to explain the higher values of cyclogenesis density in the region west of Japan, where stronger winds would lead to higher values of vorticity and indicate a higher cyclogenesis density in this region.

Trends identified within the 20CR and NCEP1 climatologies can be compared by sub-region for the 1950 – 2010, 1950 – 1978, and 1979 – 2010 periods. For the 1950 – 2010 period, trends of similar magnitude are present in most sub-regions. Two trends significant at the 5% or higher level are found in both datasets. First, increases in Gulf of Alaska cyclolysis in the winter are observed in both the 20CR (0.57%) and NCEP1 (0.96%) climatologies for the 1950 – 2010 period. Second, decreases in cyclogenesis frequency in the fall are found in the 1979 – 2010 period for both 20CR (-0.92%) and NCEP1 (-1.18%). Overall, it is not expected that trends would be in complete agreement,
primarily due to the difference in assimilation methods of NCEP1 and 20CR. These results echo some of the findings of Wang et al. (2012).

Shifts in the various indicators of storm activity within the Chukchi/Beaufort sub-region, particularly in the 1970s, are evident in the cyclolysis and cyclogenesis time series (Fig. 4.13). These shifts are also noted by Atkinson (2005) with respect to wind event frequency and overall storminess in parts of the circum-Arctic. The time frame of the shifts in both the cyclogenesis and cyclolysis time series are also very similar to those found by Saveliava et al. (2000), where shifts in meteorological and hydrological parameters were found to be related to systematic shift in the location of storm tracks and major pressure features. This suggests that large scale changes in the time series, identified by the inhomogeneities within the time series, are not related to technical issues within that dataset but rather are guided by known physical shifts in the meteorological climate within the region.

4.5 Summary

In this study, a comparison of two cyclone identification and tracking algorithms is performed using 6-hourly NCEP1 MSLP (Serreze method) and 6-hourly NCEP1 and 20CR relative vorticity (Hodges method). Additionally, a comparison of the NCEP1 and 20CR climatologies for the North Pacific, Bering Sea, and Alaska regions is conducted. Analysis of changepoints in the long term 20CR time series within each sub-region is performed to detect significant shifts ($\alpha < 0.05$) in the mean which affect the homogeneity of time series.

Using the 20CR storm track dataset provides a temporal extension to the climatology of ETC characteristics, which has been found to be stable across most of the
North Pacific region after 1920. The use of temporal and spatial matching of tracks between ensemble members increases the confidence in the tracks and track statistics that are established. The addition of these years prior to ~1950 (when the NCEP1 dataset begins) considerably extends the homogenous climatological records of ETC activity within some sub-regions, such as the Gulf of Alaska and Bering Sea. There remain issues with inhomogeneities in some data sparse sub-regions, however, such as the Alaska Interior and the Chukchi/Beaufort. It must be noted that data sparsity over the Alaska Interior may not be due to lack of observational capacity but rather a lack of storm events to observe – storms do not tend to penetrate into the Alaska Interior because they are blocked by the mountains of the Alaska Range. Issues in the Chukchi/Beaufort sub-region likely arise because the region is an importer of ETCs from other sub-regions and not a place where they form (Sepp and Jaagus, 2011). Nevertheless, this echoes the findings of Wang et al. (2012), that many sub-regions in the North Pacific were sensitive to data scarcity throughout the early period of the 20th century.

The general patterns of ETC characteristics are similar between the two different storm tracking algorithms when using the same reanalysis. The main difference in storm parameters (cyclogenesis density, cyclolysis density, and track density) is the higher values observed by the Hodges method compared to the Serreze method. This is due mainly to the different ends of the synoptic range that are utilized by each of the methods (relative vorticity vs. MSLP, respectively): the Hodges method should capture more storm events because of vorticity being at the higher frequency end of the synoptic temporal scale. Only two statistically significant trends at the 5% level are found in the same time periods (1950 – 2010 Gulf of Alaska Winter, 1979 – 2010 Alaska Interior Fall)
within the NCEP1 (Table 4.3) and 20CR (Table 4.4) time series. In general, inhomogeneities within the 20CR climatology are more prevalent in the poleward sub-regions; e.g., the Chukchi/Beaufort Region (Table 4.4).
Table 4.1. Regions and sub-regions used in this study.

<table>
<thead>
<tr>
<th>Sub-region</th>
<th>Abbreviation</th>
<th>Latitude Bounds</th>
<th>Longitude Bounds</th>
</tr>
</thead>
<tbody>
<tr>
<td>Gulf of Alaska</td>
<td>GA</td>
<td>55 - 63 °N</td>
<td>200 - 225 °E</td>
</tr>
<tr>
<td>Bering Sea</td>
<td>BS</td>
<td>50 - 68 °N</td>
<td>170 - 200 °E</td>
</tr>
<tr>
<td>Alaska Interior</td>
<td>AI</td>
<td>60 - 70 °N</td>
<td>200 - 240 °E</td>
</tr>
<tr>
<td>Chukchi/Beaufort</td>
<td>CB</td>
<td>68 - 75 °N</td>
<td>180 - 235 °E</td>
</tr>
</tbody>
</table>
Table 4.2. Type-1 Changepoints for the cyclogenesis and cyclolysis annual time series for the listed sub-regions from the Hodges 20CR for the 1871 – 2010 period.

<table>
<thead>
<tr>
<th></th>
<th>GA</th>
<th>BS</th>
<th>AI</th>
<th>CB</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Genesis</td>
<td>Genesis</td>
<td>Genesis</td>
<td>Genesis</td>
</tr>
<tr>
<td></td>
<td>Lysis</td>
<td>Lysis</td>
<td>Lysis</td>
<td>Lysis</td>
</tr>
<tr>
<td></td>
<td>N/A</td>
<td>1899</td>
<td>1884</td>
<td>1925</td>
</tr>
<tr>
<td></td>
<td>1910</td>
<td>1910</td>
<td>1882</td>
<td>1910</td>
</tr>
<tr>
<td></td>
<td>1939</td>
<td>1910</td>
<td>1907</td>
<td>1935</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>1917</td>
<td>1974</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>1946</td>
<td>1952</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>1966</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>1977</td>
</tr>
</tbody>
</table>
Table 4.3. Trends and significance of lysis density and genesis density by sub-region for the a) Hodges NCEP1 database and the b) Serreze NCEP1 database. Periods evaluated for trends are 1950 – 2010, 1950 – 1978, 1979 – 2010. Units are % yr\(^{-1}\). Trends that are significant at 95% are indicated by \textit{bold and italic} numbers. Trends that are significant at 99% are indicated by \textit{bold, italic, and underlined} numbers. Non-highlighted values are not statistically significant.

<table>
<thead>
<tr>
<th></th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>JFM</td>
<td>AMJ</td>
<td>JAS</td>
</tr>
<tr>
<td>Gulf of Alaska</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>lysis</td>
<td>0.96</td>
<td>0.38</td>
<td>0.42</td>
</tr>
<tr>
<td>genesis</td>
<td>-0.75</td>
<td>-0.38</td>
<td>-0.87</td>
</tr>
<tr>
<td>Bering Sea</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>lysis</td>
<td>0.18</td>
<td>-0.33</td>
<td>-0.02</td>
</tr>
<tr>
<td>genesis</td>
<td>-0.11</td>
<td>-0.43</td>
<td>0.30</td>
</tr>
<tr>
<td>Alaska Interior</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>lysis</td>
<td>-0.29</td>
<td>-0.30</td>
<td>0.34</td>
</tr>
<tr>
<td>genesis</td>
<td>0.11</td>
<td>0.32</td>
<td>1.09</td>
</tr>
<tr>
<td>Chukchi/Beaufort</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>lysis</td>
<td>-0.16</td>
<td>-0.18</td>
<td>0.44</td>
</tr>
<tr>
<td>genesis</td>
<td>0.70</td>
<td>1.05</td>
<td>-0.24</td>
</tr>
<tr>
<td>----------------</td>
<td>-----------</td>
<td>----------</td>
<td>----------</td>
</tr>
<tr>
<td></td>
<td>JFM</td>
<td>AMJ</td>
<td>JAS</td>
</tr>
<tr>
<td>Gulf of Alaska</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>lysis</td>
<td>0.09</td>
<td>0.07</td>
<td>-0.06</td>
</tr>
<tr>
<td>genesis</td>
<td><strong>-0.53</strong></td>
<td><strong>-0.73</strong></td>
<td>-0.28</td>
</tr>
<tr>
<td>Bering Sea</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>lysis</td>
<td>-0.06</td>
<td>-0.39</td>
<td>-0.10</td>
</tr>
<tr>
<td>genesis</td>
<td>0.28</td>
<td>-0.13</td>
<td>0.40</td>
</tr>
<tr>
<td>Alaska Interior</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>lysis</td>
<td>0.13</td>
<td><strong>-0.35</strong></td>
<td>-0.32</td>
</tr>
<tr>
<td>genesis</td>
<td>-0.39</td>
<td><strong>-0.66</strong></td>
<td><strong>-0.53</strong></td>
</tr>
<tr>
<td>Chukchi/Beaufort</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>lysis</td>
<td>0.42</td>
<td>-0.30</td>
<td>-0.26</td>
</tr>
<tr>
<td>genesis</td>
<td><strong>1.34</strong></td>
<td>0.03</td>
<td><strong>-0.57</strong></td>
</tr>
</tbody>
</table>
Table 4.4. Trends and significance of lysis density and genesis density by sub-region for the Hodges 20CR database. Periods evaluated for trends are 1950 – 2010, 1950 – 1978, 1978 – 2010, 1920 – 2010, and 1920 – 1949. Units are % yr\(^{-1}\). Trends that are significant at 95% are indicated by **bold and italic** numbers. Trends that are significant at 99% are indicated by **bold, italic, and underlined** numbers. Non-highlighted values are not statistically significant.

<table>
<thead>
<tr>
<th></th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>JFM</td>
<td>AMJ</td>
<td>JAS</td>
</tr>
<tr>
<td>Gulf of Alaska</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>lysis</td>
<td>0.57</td>
<td>0.16</td>
<td>0.11</td>
</tr>
<tr>
<td>genesis</td>
<td>-0.62</td>
<td>-0.25</td>
<td>0.05</td>
</tr>
<tr>
<td>Bering Sea</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>lysis</td>
<td>0.13</td>
<td>-0.18</td>
<td>-0.05</td>
</tr>
<tr>
<td>genesis</td>
<td>0.22</td>
<td>0.22</td>
<td>-0.38</td>
</tr>
<tr>
<td>Alaska Interior</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>lysis</td>
<td>-0.36</td>
<td>-0.33</td>
<td>0.29</td>
</tr>
<tr>
<td>genesis</td>
<td>-0.06</td>
<td>0.20</td>
<td>-0.14</td>
</tr>
<tr>
<td>Chukchi/Beaufort</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>lysis</td>
<td>-0.64</td>
<td>0.01</td>
<td><strong>-0.57</strong></td>
</tr>
<tr>
<td>genesis</td>
<td>0.89</td>
<td>0.65</td>
<td>-0.09</td>
</tr>
<tr>
<td>Hodges 20CR</td>
<td>1920 - 2010</td>
<td>1920 - 1949</td>
<td></td>
</tr>
<tr>
<td>----------------</td>
<td>-------------</td>
<td>-------------</td>
<td></td>
</tr>
<tr>
<td></td>
<td>JFM</td>
<td>AMJ</td>
<td>JAS</td>
</tr>
<tr>
<td>Gulf of Alaska</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>lysis</td>
<td>0.45</td>
<td>0.17</td>
<td>0.43</td>
</tr>
<tr>
<td>genesis</td>
<td>0.19</td>
<td>0.53</td>
<td>-0.14</td>
</tr>
<tr>
<td>Bering Sea</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>lysis</td>
<td>-0.06</td>
<td>0.01</td>
<td>-0.27</td>
</tr>
<tr>
<td>genesis</td>
<td>-0.01</td>
<td>0.00</td>
<td>-0.68</td>
</tr>
<tr>
<td>Alaska Interior</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>lysis</td>
<td>0.35</td>
<td>0.66</td>
<td>1.44</td>
</tr>
<tr>
<td>genesis</td>
<td>1.58</td>
<td>1.27</td>
<td>1.33</td>
</tr>
<tr>
<td>Chukchi/Beaufort</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>lysis</td>
<td>0.81</td>
<td>2.54</td>
<td>0.89</td>
</tr>
<tr>
<td>genesis</td>
<td>2.14</td>
<td>1.29</td>
<td>1.45</td>
</tr>
</tbody>
</table>
Figure 4.1. Map of study area with some key locations highlighted. Sub-regions in the Gulf of Alaska, Bering Sea, Chukchi and Beaufort Seas, and Alaska Interior are also used in this analysis.
Figure 4.2. Cyclogenesis density climatology for the a) Winter (JFM), b) Spring (AMJ), c) Summer (JAS), and d) Fall (OND) seasons using the MSLP-based Serreze cyclone identification and tracking algorithm with the NCEP1 reanalysis for the 1950 – 2010 period. Units: density of cyclone starting points (10⁶ km² season)⁻¹
Figure 4.3. As in Fig. 4.2 but for the 850 hPa relative vorticity-based Hodges storm identification and tracking algorithm.
Figure 4.4. Cyclolysis density climatology for the a) Winter (JFM), b) Spring (AMJ), c) Summer (JAS), and d) Fall (OND) seasons using the Serreze cyclone identification and tracking algorithm with the NCEP1 reanalysis for the 1950 – 2010 period. Units: density of cyclone ending points (10^6 km^2 season)^{-1}
Figure 4.5. As in Fig. 4.4 but for the 850 hPa relative vorticity-based Hodges storm identification and tracking algorithm.
Figure 4.6. Cyclone track density climatology for the a) Winter (JFM), b) Spring (AMJ), c) Summer (JAS), and d) Fall (OND) seasons using the MSLP-based Serreze cyclone identification and tracking algorithm with the NCEP1 reanalysis for the 1950 – 2010 period. Units: storms ($10^6$ km$^2$ season)$^{-1}$
Figure 4.7. As in Fig. 4.6 but for the 850 hPa relative vorticity-based Hodges storm identification and tracking algorithm.
Figure 4.8. Cyclogenesis density climatology for the a) Winter (JFM), b) Spring (AMJ), c) Summer (JAS), and d) Fall (OND) seasons using the 850 hPa relative vorticity based Hodges cyclone identification and tracking algorithm for the 20CR reanalysis matched tracks between the 56 ensemble members for the 1950 – 2010 period. Units: density of starting points (10^6 km^2 season)^1
Figure 4.9. As in Fig. 4.8 but for 20CR cyclolysis density. Units: density of ending points ($10^6$ km$^2$ season)$^{-1}$
Figure 4.10. Difference between climatological cyclogenesis density between the Hodges based NCEP1 and Hodges based 20CR for the a) Winter (JFM), b) Spring (AMJ), c) Summer (JAS), and d) Fall (OND) seasons for the 1950 – 2010 period. Positive (red) values indicate higher density in the NCEP1 while negative (blue) values indicate higher density in the 20CR climatology. Units: density of starting points ($10^6$ km$^2$ season)$^{-1}$.
Figure 4.11. As in figure 4.10, except NCEP1–20CR cyclolysis density. Units: density of ending points (10^6 km^2 season)^{-1}
Figure 4.12. a) Annual base anomaly series (reference period 1950 – 2010) with significant Type-1 changepoints of storms undergoing cyclogenesis, b) mean adjusted base series for Bering Sea region in the Hodges 20CR ensemble mean from matched tracks, c) Annual base anomaly series with significant Type-1 changepoints of storms undergoing cyclolysis and, d) mean adjusted base series for Bering Sea region in the Hodges 20CR ensemble mean from matched tracks.
Figure 4.13. a) Annual base anomaly series (reference period 1950 – 2010) with significant Type-1 changepoints of storms undergoing cyclogenesis, b) mean adjusted base series for Chukchi/Beaufort region in the Hodges 20CR ensemble mean from matched tracks, c) Annual base anomaly series with significant Type-1 changepoints of storms undergoing cyclolysis and, d) mean adjusted base series for Chukchi/Beaufort region in the Hodges 20CR ensemble mean from matched tracks.
Figure 4.14. Annual (a) zonal and (b) meridional 850 hPa wind differences between NCEP1 and 20CR for the study area. Units are ms$^{-1}$. 
Chapter 5 Seasonal Climatology and Trends of Strong Wind and Lull Events in the Circum-Arctic During the 1979 – 2010 Period Using a Novel Lull/Storm Wind Indicator

Article Information

Chapter 5 consists of a manuscript submitted to the Journal of Applied Meteorology and Climatology. Figures and tables are the same as those submitted within the paper but have been renumbered for thesis consistency. References have also been reformatted for thesis consistency.

Author Information

*N. J. Shippee¹
D. E. Atkinson²
J. Partain³

¹N. J. Shippee, University of Victoria Climate Lab, Department of Geography, University of Victoria, PO Box 3060 STN CSC, Victoria, British Columbia, Canada, V8W 3R4
²D. E. Atkinson, University of Victoria Climate Lab, Department of Geography, University of Victoria, PO Box 3060 STN CSC, Victoria, British Columbia, Canada, V8W 3R4
³J. Partain, National Atmospheric and Oceanic Administration, National Centers for Environmental Information, Anchorage, Alaska, USA.

*Corresponding author email: nshippee@uvic.ca

Author Contributions

Shippee developed the study and hypothesis, conducted the analysis, created the figures and tables, and wrote the manuscript. Atkinson provided the initial algorithm,
reviewed and edited the manuscript. Partain provided feedback and minor edits to the manuscript.

Abstract

Extratropical cyclones (ETCs) generate strong impacts in marine and coastal environments throughout the circum-Arctic region. Within a marine context, many impacts are caused by strong winds. For example, hazardous sea states, storm surge, and coastal erosion are common marine/coastal impacts from ETCs. Using a modified wind event identification algorithm as applied to reanalysis data, this study aims to determine the spatial and temporal characteristics of strong wind events in the circum-Arctic, defined as 45°N to the pole. In addition to wind events, a novel indicator is introduced that characterizes periods of favorable weather between strong wind events that last 48-hours or longer, termed lull events. Lull periods have been found to be an important consideration for northern marine operations – both economic and subsistence. This paper presents seasonal climatologies of wind and lull events and their associated trends over the 1979-2010 period using the ECMWF ERA-Interim reanalysis. Strong wind events often occur in locations and seasons where ETCs are prevalent. An interesting result shows opposing trends between 48-hour or longer lulls and lulls with a duration less than 48-hours in the North Atlantic (decreasing/increasing, respectively), suggesting an increasing frequency of wind events in this region. Additionally, combinations of lull and wind event indicators, termed lull/storm winds (LSW), were analyzed. Overall, LSW analysis shows that preferred areas of wind events and lull events are not always spatially coherent. This analysis shows that the North Atlantic is the most “stormy” region in the region north of 45°N.
5.1 Introduction

Extratropical cyclones (ETCs) are the primary source of high-impact weather in many areas of the North Pacific, Alaska, and Arctic. Depending on perspective or rationale, how a storm is defined may vary widely. Most research considers ETCs from a theoretical perspective; that is, storm definitions based on meteorological variables typically associated with cyclone development (Mesquita, 2009). These variables may include (but are not limited to) mean sea level pressure (MSLP), relative vorticity (\( \zeta \)), and measures such as pressure tendency. From the perspective of a person experiencing the impacts of an ETC, the way that a storm is defined is more a function of what the storm “brings” to a location. This is a more pragmatic view that relies less on theory-based indicators and more on the impacts generated by ETCs. These impacts might include high wind speeds, heavy precipitation, hazardous sea state and/or coastal inundation (Mason et al., 1993; Blier et al., 1997; Lynch et al., 2008). For example, a wind event assessment method developed by Atkinson (2005; referred to as A05 henceforth) provides a pragmatic perspective by allowing for the identification of different variations of “storms,” primarily related to the speed, direction, and duration of high wind events.

The pragmatic perspective allows for more direct consideration of the needs of end-users or stakeholders, making it a relevant means of storm identification in an operational and impacts-oriented context. Many regions of the circum-Arctic are home to sensitive ecological habitats and extensive fisheries, like the Bering Sea (ACIA, 2005). Indigenous communities rely on subsistence fishing and hunting, activities that can be particularly vulnerable to weather impacts. These regions are also vulnerable due to infrastructural constraints, which can result in an inability to provide quick response to
incidents. In the past quarter century, there have been many examples of marine incidents with socio-economic and ecological impacts, that have been exacerbated by the weather. One of the most notable examples occurred in 2004 near Dutch Harbor, Alaska, when the MV Selendang Ayu lost power during high winds and heavy seas, eventually leading to the loss of the vessel (Brewer, 2006). In another example, in November 2011, shipments of fuel for the city of Nome, AK were cancelled due to anomalously stormy conditions in the Aleutians and western Alaska (Medearis, 2011). Without an unprecedented, expensive wintertime delivery of fuel that required icebreaker support from the USCGC Healy (Medearis, 2012), the city of Nome would have run out of fuel.

Many regions of the circum-Arctic are serviced by tug-and-barge style marine transport that operates during the limited ice-free shipping season (AMSA, 2009). Tug-and-barge style providers are highly impacted by the marine state, with strong winds driving wave states that make towing barges unsafe. As a result, most shipping via tug-and-barge operations is conducted during periods of lower wind speed, which allows for safe transport of materials. Interviews with stakeholders in the marine transport sector have indicated that considering breaks between wind events, termed lull events, as much as the wind events themselves are important for marine operations (Eerkes-Medrano, personal communication, 2015). Depending on location and the geography of the marine landscape, the length of a typical safe operating window tends to vary between 36 and 48 hours. For the Alaska region, tug-and-barge operators require at least a 48-hour break between wind events to navigate between points of safe harbour or points of refuge. If a period of less than 48-hours of safe marine state is present in a forecast period, operations are halted until weather becomes more favorable.
This study has four main objectives:

i. Describe and evaluate a novel storm/lull winds indicator.

ii. Establish circum-Arctic seasonal climatologies of both wind and lull event indicators.

iii. Establish seasonal trends of the wind and lull event indicators for the 1979 – 2010 period.

iv. Analyze a combined indicator of wind events and lull events: a lull/storm winds (LSW) indicator.

The remainder of this paper is structured as follows: Section 5.2 introduces the LSW algorithm and how it differs from the A05 algorithm and overviews the data sources used in this study. Section 5.3 overviews the climatologies of both wind and lull events identified by the LSW algorithm. Additionally, trends analysis of both the wind and lull events is presented in Section 5.3. Section 5.4 provides a discussion of the results and concluding remarks. Section 5.5 presents future plans for the lull/storm wind algorithm.

5.2 Data and Methods

5.2.1 LSW Algorithm

The lull/storm wind algorithm (LSW) is built on an existing wind event identification algorithm by Atkinson (2005; A05). The original implementation of the A05 algorithm targeted wind events that could potentially produce waves and/or surges of a magnitude sufficient to impact a coastal region. For the purposes of the A05 study, “impacts” referred to sea state conditions with the potential to cause damage to coastal habitats and infrastructure, or that could potentially perform geomorphological work,
such as erosion. The wind speed and duration thresholds for event identification in A05 were set to 10 ms\(^{-1}\) for at least a 6-hour duration, independent of wind direction. These thresholds echoed those established in previous studies of impactful weather conditions for coastal geomorphological processes and sea states (e.g. Solomon et al., 1994, MacClenehan et al., 2001). The thresholds remain the same for this study with respect to wind event identification in the LSW algorithm. The use of a consistent 10 ms\(^{-1}\) wind speed threshold provides a standard value, allowing for comparison between older studies and this version of the algorithm. As part of the LSW algorithm, identification of lull events is completed as well.

The LSW algorithm is a two-stage approach in which occurrences of two types of events are identified: wind and lulls (Fig. 5.1). Both parts of the algorithm, wind and lull, identify events using a multistage approach that consists of a series of rules applied to the time-series wind speed trace. Following A05, wind events are first identified during an initial pass through the data by tagging any wind speed exceeding the 10 ms\(^{-1}\) threshold. Tags that are sequential are grouped to form a discrete event. It was recognized that for synoptic continuity the characterization of the full duration of a wind event must include short periods when the wind speed drops just below threshold, and periods preceding or following the main event that are also just below threshold. To account for this, the grouping process makes use of two additional features, termed “breaks” and “shoulder events” to fully describe a wind event. More specifically, a break is a temporary decrease in wind speed during an otherwise synoptically contiguous event. Allowing for breaks in events helps to prevent the separation, or “double counting”, of what would otherwise be a single event. The second feature, termed shoulder events, occur prior to the first or
following the last observations that are identified at the 10 ms\(^{-1}\) event threshold. These features are essentially the start, or the “ramp-up”, and end, or “ramp-down”, of the discrete event. For the purposes of the LSW algorithm, the \textit{wind event} break/shoulder values begin/end at 7 ms\(^{-1}\) in the LSW algorithm. For lull events, the event threshold is specified as 7 ms\(^{-1}\), with \textit{lull event} break/shoulder values of 10 ms\(^{-1}\) to allow for temporary (1 time step) increases in the wind speed during an otherwise continuous lull period and capture the lull event start and end in the LSW algorithm. In both cases, breaks are only allowed for a single 6-hr observation, as consecutive values exceeding the break threshold would end the event. The use of these specific thresholds may cause a small percentage of wind events and lull events to be occurring at the same time, primarily due to the shoulder event design. Wind and lull events are not expected to be mutually exclusive; there may be some period toward the end of a lull that is included in the “ramp up” of a wind event (the reverse is also possible). In this study, 6-hourly wind speed data are linearly interpolated to hourly wind speed values for use in the LSW algorithm. This action is performed to give a more precise approximation to the to the duration of both wind and lull events. For both types of events, an individual event database was developed, that included start and end day, duration, and grid point location. For wind events, additional information was recorded including the average core wind speed of the entire event, including shoulders, and the average wind speed of readings over the 10 ms\(^{-1}\) threshold.

\textit{5.2.2 Data Sources}

ERA-Interim (ERA-I) is one of the most recent reanalyses available from European Centre for Medium-Range Weather Forecasting (ECMWF), covering 1979 –
present (Dee et al., 2011). ERA-I uses a spectral model with a T255 (80 km) horizontal resolution and 60 vertical hybrid levels. For this study, initial forecast fields, available with a 6-hr temporal scale, from the reanalysis (forecast hour zero) of 10 metre u- and v-wind components are used to calculate wind speed at each grid point.

The LSW algorithm was run using 0.5° spatial resolution fields obtained from ECMWF over the circum-Arctic region with an extra southern extension, defined as north of 45°N. The LSW algorithm is applied to each reanalysis grid point over the time period from 1 January 1979 to 31 December 2010. Annual results were split into seasons using three-month ‘seasonal’ periods of January – March (JFM; winter), April – June (AMJ; spring), July – September (JAS; summer), and October – December (OND; fall). Lull event data underwent additional post-processing to identify lulls with durations exceeding 48-hours to match the prescribed needs of end-users. Additionally, locations with spuriously long lived lull events were removed, as they are outside the synoptic temporal scale of interest in this study. For this study, the climatology and trends of lulls with durations 48-hours or longer are shown. Lulls with durations of longer than the duration of a season (approximately 2160 hours) are filtered as they are outside of the synoptic temporal interest of this study. Comparisons with the trends and durations of all lull events are made within section 4, however, they are not the focus of the LSW algorithm.
5.3 Results

5.3.1 Climatology

5.3.1.1 Wind Events

The seasonal spatial climatological distribution of wind events (Fig. 5.2) reveals that the primary regions of activity are found in regions of sharp land/ocean contrast, with resultant differences likely due to changes in surface roughness parameterization within the reanalysis. In the Atlantic region, the axis of maximum activity is found in the Canadian Maritimes and the along the southeastern coast of Greenland. Over the 1979 – 2010 period, these regions average between 20 – 30 wind events per season in the winter and fall seasons (Fig. 5.2a and 5.2d, respectively). The winter season has a slightly higher number of wind events with respect to the fall season, though the difference is most notable in the areas near Greenland and Iceland and less so in the Canadian Maritimes. The spring and summer seasons (Fig. 5.2b and 5.2c, respectively) exhibit lower event counts overall in the Canadian Maritimes and coastal southeastern Greenland, with the total number of events decreasing by about 50% to 10 – 16 per season, with respect to the winter season.

In the Pacific region, much of the North Pacific, Gulf of Alaska, and the coastal regions of British Columbia and the US Pacific Northwest exhibit similar wind event frequency totals, with 20 – 30 events per season on average in the winter and fall seasons. Frequency maxima are typically located in the Gulf of Alaska in the winter and fall seasons, with winter maxima exhibiting an equatorward shift when compared with the fall. Poleward of the North Pacific region, a distinct meridional gradient is observed through the Bering Sea toward the Bering Strait, ranging from maxima near 24 – 26
events per season at the edge of the Aleutian Islands to 12 – 14 events per season at the southern opening of the Bering Strait. In the spring and summer seasons, a distinct decrease of close to 50% in the frequency of wind events is observed in comparison to the fall and winter seasons, with localized maxima (10 – 14 events) occurring to the south of the Aleutian Islands and decreasing along a similar meridional gradient from the Aleutians to the Bering Strait (local maxima of 8 – 10 events). When inter-compared, spring tends to show slightly more activity on average than summer. For the winter season, other more localized maxima are observed over Hudson Bay, the Great Lakes, and the Sea of Okhotsk, with between 14 – 20 events per season. These same regions experience a large reduction in the frequency of wind events during the spring and summer, with a decrease to 4 – 10 events during these seasons.

5.3.1.2. 48-hr+ Lull events

The climatology of lull events lasting 48-hours or longer (Fig. 5.3) reveals the highest frequency of occurrence is typically found in the spring and summer seasons (Fig. 5.3b and 5.3c, respectively), when there are fewer wind events observed (Fig. 5.2). In the North Atlantic region, the number of lull events is at a minimum during the winter and fall seasons (Fig. 5.3a and 5.3d, respectively), with an average of 2 – 5 events per season in total. The coastal regions of Newfoundland and Labrador see a small increase in the frequency of lulls (6 – 7 events) in these seasons. In the spring and summer seasons, the frequency of lulls increases and ranges from 10 – 13 events across most of the Atlantic, stretching from the Canadian Maritimes to continental Europe. A relatively high frequency of lulls also exists across the Hudson Bay region throughout the year, with a range of 5-9 events per season.
In the Pacific region, the number of lulls in the central North Pacific Ocean is at its lowest during the winter and fall seasons, and increases toward the pole. The highest frequency of lull events in the winter and fall seasons is found in the Gulf of Alaska, along the Alaskan coastline and in the Sea of Okhotsk, with a total of 8 – 10 events during both of the seasons, compared with only 2 – 5 events across most of the central North Pacific Ocean. In the spring and summer seasons, an increase in the number of lulls is observed across most of the North Pacific Ocean, stretching into the Bering Sea and Gulf of Alaska. In these regions, there is an average of 8 – 10 events per season. In all, coastal areas in the Pacific region tend to show the most inter-season consistency with respect to the frequency of lulls, averaging between 6 – 10 events per season.

For regions north of 60°N, on both the Atlantic and Pacific sides of the Arctic, the greatest number of lulls occur in the winter and summer seasons, with 6 – 9 events on average. The maximum frequency of lulls in the winter and fall is located near the Atlantic entrance to the Arctic. The location of maximum events shifts in the other seasons (spring and summer) closer to the Eurasian continent and with smaller local maxima values found in the Beaufort and Chukchi Seas.

5.3.2 Trends

5.3.2.1 Wind Events

Seasonal trends are calculated using a Mann-Kendall trend test (Mann, 1945, Kendall, 1975, Gilbert, 1987) over the 1979 – 2010 period. Using wind and lull event frequency, areas of trends significant at \( p < 0.1 \) are shaded and contoured, while areas of trends significant at \( p < 0.05 \) are indicated with black symbols. Wind events (Fig. 5.4) show positive trends in the winter (Fig. 5.4a) and spring (Fig. 5.4b) seasons across
portions of the Atlantic region. Winter trends show statistically significant increases in the frequency of wind events along the Newfoundland and Labrador coastlines. A similar trend is visible in the Sea of Okhotsk region and across small portions of the North Pacific and Gulf of Alaska. In the central Arctic, a decreasing trend is observed for the winter season, primarily north of Greenland on the Atlantic side of the Arctic. In the spring season, a larger area of increasing trends (compared to the winter season) is found in the North Atlantic, stretching from the Canadian coastline to the United Kingdom. Additionally, in the North Pacific, a more coherent region of decreasing trend is found south of the Aleutian Islands. In the summer (Fig. 5.4c), there are many discrete areas of both increasing and decreasing trends. The most organized area of increasing trends can be found off the coastline of British Columbia while the most organized area of decreasing trends can be found on the western edge of the Canadian Arctic Archipelago. In the fall, the primary regions of organized trends are found in the North Atlantic and central Arctic, with increasing trends in wind events to the south of Iceland and decreasing trends in the central Arctic. Large areas of increased wind event frequency can also be found throughout the Canadian Archipelago. For each season, the magnitude of the trend for specific regions (denoted by A. and B. in Fig. 5.4) have also been calculated (Table 5.1).

5.3.2.2 Lull Events

Overall, compared to the wind event trend analysis, there are fewer regions of contiguous spatial trends in the lull event data (Fig. 5.5). In winter (Fig. 5.5a), the strongest trend (decreasing) is found in the Eurasian sector of the Arctic. Areas of trends elsewhere are sporadic, with few areas of increasing trends observed throughout the study.
region. In the spring (Fig. 5.5b), decreasing trends are observed in the central Arctic and along the Eurasian coastline. In summer (Fig. 5.5c), a decreasing trend is seen in the circum-Arctic regions, in particular along the western boundary of the Canadian Arctic Archipelago and the Atlantic entrance to the Arctic Ocean. Fall (Fig. 5.5d) is the only season with an organized area of increasing trends, with a section of the Canadian Arctic Archipelago showing increased frequency of lulls. Similar to other seasons, an area of decreasing trends near 90°N is also observed in the fall, stretching from the northeast corner of Greenland to 120°E. As was completed for the wind event trends, the magnitude of lull event trends for each season in specific regions, denoted by A (Fig. 5.5a, Fig. 5.5b, Fig. 5.5c) and B (Fig. 5.5d), have been calculated (Table 5.2). For the lull table, not all seasons have contiguous areas of positive trends. Therefore, the seasons of winter, spring, and summer (Fig. 5.5a, Fig. 5.5b, Fig. 5.5c, respectively) only highlight one distinct area of decreasing trends, while the fall (Fig. 5.5d.) a larger organized area of increasing lull trends is shown.

5.4 Discussion and Conclusions

Overall, the wind event trends for the circum-Arctic region exhibit greater large scale spatial coherence than do 48-hour+ lull events for the 1979-2010 period. Close to the pole, a decreasing trend in both wind event frequency and 48-hr lull event frequency is found in nearly every season, with the most spatially coherent regions found in the winter and fall seasons. For nearly all the selected subregions, the trends observed in the wind event frequency are of greater magnitude than those of the lull event frequency. For example, sub region A (Fig. 5.4a; Table 5.1) exhibits a statistically significant decreasing trend (expressed in percent change per year) in the frequency of wind events (-0.39%),
compared to a decreasing trend in the frequency of 48-hour+ lulls (-0.25%) (Fig. 5.5a; Table 5.2). These combined decreasing trends over the Arctic would suggest a decrease in windiness over the 1979 – 2010 period.

The probability density function of lull event durations (Fig. 5.6) shows that the majority of all lull events occur with a duration of 48-hours+. Additionally, the cumulative distribution function (Fig. 5.7) exhibits the cumulative probability of lulls lasting less than 48-hours ranges 25 – 30% of all the lull event durations in the circum-Arctic region. These results show that when evaluating lull events of any duration, the majority of lull events that occur have a duration of 48-hours or greater. This indicates that the application of the 48-hour threshold is not overly data limiting, as most events are still retained within the analysis.

One interesting result from this study is that some regions with a decreasing trend of lull events lasting 48-hours+ have opposing trends when considering lulls of durations between 6-48 hours. This suggests a shift to more frequent short duration lulls, indicating an increase in windiness at these locations. This pattern is prevalent in the North Atlantic region during the winter season (Fig. 5.8). Wind event trends in this region show a similar increase in frequency in the winter season. The combination of these factors suggests an increase of windiness in this region and a corresponding decrease in the frequency of safe operational windows for shipping. This is an important result, as decreasing calm periods in the North Atlantic over the 1979 – 2010 period highlights a change in the distribution of both wind events and lull events in the region.

To better interpret the results of this study, the relationship between wind events and lull events, as identified by the LSW algorithm, must be considered. Event
identification is largely contingent on the thresholds applied within the LSW algorithm. For example, it is possible to have a lull event starting as a wind event is ending. If the wind speed decreases from 10 ms\(^{-1}\) to 7 ms\(^{-1}\) over a period longer than 6 hours, the wind event will end and the LSW algorithm will include the hours of the wind decrease in the total duration of the event. At the same time, a lull event is identified when the wind speed is below 7 ms\(^{-1}\), but the algorithm also allows for wind speed to go as high as 10 ms\(^{-1}\) over a short time period (e.g., 1 hr) during a lull event and includes the ramp-down of the wind event in the lull duration. Therefore, the decrease in wind speed to 7 ms\(^{-1}\) would also be included in the front shoulder event of the overall lull event duration.

Another important key to the interpretation of the distribution of lull events in relation to wind events is to realize that lull events are two-tailed: a very low lull event frequency can indicate two possible situations: almost no strong wind events or so many wind events that wind speeds almost never drop below the 10 ms\(^{-1}\) threshold. Conceivably, it would be possible to have fewer long duration wind events as well, though this was not reflected at any locations within the study area. This differs from the distribution of lull event durations (Fig. 5.6), as longer duration lulls are indicative of longer periods without wind events, and vice versa. For these reasons, the lull climatology on its own does not necessarily reveal “storm free” areas. Instead information about the corresponding wind climatology at a given location is required to determine overall storminess of an area. The occurrence of wind events is also not indicative of a lull event occurring shortly thereafter. While wind events and lull events occur in relatively similar numbers in some locations (e.g. Hudson Bay, Bering Strait), the timing of these events with respect to one another does not necessarily setup as a
“wind event – lull event – wind event” pattern. Similarly, some areas that exhibit a high frequency of wind events also tend to exhibit fewer lull events. Therefore, due to the thresholding, a lower number of lulls does not necessarily indicate a “stormy” location and, vice versa, an area with more lulls does not necessarily indicate a “non-stormy” location. Consequently, thresholding specifics in the LSW algorithm leads to a four category split of possible outcomes for the combined wind and lull event.

The first possible combination is “many lulls/many wind events”. This situation is common on the edges of climatologically stormy areas and toward the end of primary storm tracks. This is evident in regions such as the Gulf of Alaska due to the major cyclolysis centers located in the region (Mesquita et al., 2010). This differs from what is expected within the cores of storm tracks as the cyclones in areas of cyclolysis tend to be weakening rather than undergoing intensification. Wind events associated with these storms would still be strong, however, and are likely followed by lull periods more frequently than compared to areas of cyclogenesis.

The second possible combination is “few lulls/many wind events”. These regions are generally located near the core of the climatological storm tracks. Specific examples of this type of region are found in the Canadian Maritimes and in the North Pacific along the southern edge of the Gulf of Alaska and coastal British Columbia. A co-location with storm tracks also means this combination tends to occur during the active ETC seasons, i.e., winter and fall (Fig 5.2a and Fig. 5.2d, respectively). As these regions experience maximum amount of ETC activity during the fall and winter seasons, it is not unexpected to see low counts of lull events in these regions during those seasons. In the spring and summer seasons, wind events and lulls are more equally distributed, as the frequency of
wind events tends to be lower than what is found in the winter and fall. This more equal wind-lull event distribution can also be tied to the lower frequency of ETCs during the spring and summer months.

The third possible combination is “few lulls/few wind events”. This occurs typically across most continental regions. This combination was found to be very sensitive to algorithm thresholds; a study aimed at investigation of wind events occurring over continental regions would require lower thresholds in both the wind and lull identification parts of the LSW algorithm. For example, Malloy et al. (2015) show that long term mean annual surface wind speeds for much of the continental United States and Canada are between 3 – 6 ms$^{-1}$, which would only qualify within the lull criteria for this study. As the thresholds for this study are set for a specific marine context, the loss of information regarding the wind and lull events over continental regions is not unexpected.

The fourth possible combination is “many lulls/few wind events”. This pattern is unlikely to be observed largely because, in the LSW algorithm, the start and end of a lull event are governed by thresholds that look for the occurrence of wind events in proximity to lull events. That is, the LSW algorithm is, in general, designed to identify lull events that occur after a wind event, and not necessarily lull events that occur without the presence of a wind event preceding them, thus giving a more synoptically contiguous picture of lull events in the context of the overall storminess of a grid point. Additionally, the application of the 48-hour duration threshold to the lull data eliminates between 20 – 30% of the total occurrences of lull events in the circum-Arctic region (Fig. 5.7).
As previously noted, the regions with the highest frequency of wind events are generally located in the same regions where there are climatologically active ETC tracks. While a “wind event” as defined by the LSW algorithm does not necessarily indicate an ETC is occurring, regions predisposed to more frequent synoptic storm activity are more likely to exhibit more frequent wind events due to the periodic enhancement of pressure gradients required to generate strong near-surface winds. As such, wind events can be expected to be more likely to occur in regions prone to ETC activity and impacts. These same controls influence the frequency and mean duration of lull events. Areas of successive ETC activity can be expected to have periodic reduction of pressure gradients as systems progress through the region, which would lead to longer lulls between storms.

The two-tailed distribution of wind and lull events makes categorizing regional storminess using wind event or lull event data individually not of particular utility to end-users. Rather, the combined information from these two parts of the LSW algorithm allows for a more complete interpretation of the storminess (or lack thereof) at a given location. The juxtaposition of regions with strong wind events and major shipping routes indicates the need to also understand the timing of lull events with respect to the wind events in order to facilitate marine transportation activities. Using wind event outputs from the LSW algorithm, the percentage of time spent in wind event criteria per season (SP) by season (Fig. 5.9) is calculated using the following equation,

$$SP = \frac{\bar{e}_n \times \bar{d}_n}{d_{max}}$$

where $e_n$ is the climatological frequency of wind events at a given grid point $n$, $d_n$ is the climatological mean duration of wind events at that same grid point, and $d_{max}$ represents the total number of hours within a season. Lull percentage (LP) (Fig. 5.10) is defined
using the same equation, except with $e_n$ representing the climatological frequency of lull events (without the 48-hr thresholding applied) and $d_n$ representing the climatological duration of lull events. These indicators allow for a climatological “storminess” or “lulliness” definition to be assigned to a given grid box. Information about the overall “storminess” of a location can be determined by combining the SP and LP information at each grid point. The remainder of the time, spent outside of the lull or wind event criteria, can be deemed “non-windiness,” an effect that is experienced more regularly over continental regions and thus is outside of the scope of this study.

Results expressed in terms of the SP and LP indicators present wind and lull event data in a more usable form for end-users as it allows for easy determination of the percentage of time that is likely to be spent in adverse wind conditions during a given season. For example, in winter (Fig. 5.9a.), there are regions (e.g., central North Atlantic) that spend above 70 – 80 % of the time in strong wind event conditions and only 10 – 20 % of the time in lull event conditions. This would indicate that adverse marine impacts could be expected to be occurring (on average, 70 – 80% of the time) during the winter season versus lull periods (occurring 10 – 20% of the time) where operations are more easily conducted. This type of situation is evident in the central North Atlantic and the southern reaches of the North Pacific region, especially in the fall and winter seasons.

An interesting result of this expression of the wind and lull data is the fact that locations with high storminess percentage values are found to, in general, exhibit lower lulliness percentage values in the same time period. Regions that spend more equal time with the lull and storminess categories are likely to be more favorable for conducting socio-economic activities, such as shipping or subsistence hunting, particularly in the
case of the Canadian Arctic Archipelago. Areas spending the most time in the lull category are the most favorable for any activity, and usually appear in the spring and summer months. Expressing the data in the SP and LP form highlights the North Atlantic as one of the stormiest locations in the circum-Arctic, as the SP maximum values persist in this region, independent of time of year.
Table 5.1. Slopes, expressed as percent change per year, of the seasonal wind event frequency anomaly trendline for selected subregions, 1979 – 2010.

<table>
<thead>
<tr>
<th></th>
<th>Winter (JFM)</th>
<th>Spring (AMJ)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Location</td>
<td>Slope</td>
<td>Location</td>
</tr>
<tr>
<td>A</td>
<td>-0.39</td>
<td>A</td>
</tr>
<tr>
<td>B</td>
<td>0.42</td>
<td>B</td>
</tr>
<tr>
<td><strong>Summer (JAS)</strong></td>
<td></td>
<td><strong>Winter (OND)</strong></td>
</tr>
<tr>
<td>Location</td>
<td>Slope</td>
<td>Location</td>
</tr>
<tr>
<td>A</td>
<td>0.34</td>
<td>A</td>
</tr>
<tr>
<td>B</td>
<td>-0.41</td>
<td>B</td>
</tr>
</tbody>
</table>
Table 5.2. Slopes, expressed as percent change per year, of the seasonal 48-hour+ lull event frequency anomaly trendline for selected subregions, 1979 – 2010.

<table>
<thead>
<tr>
<th>Location</th>
<th>Slope</th>
<th>Location</th>
<th>Slope</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>-0.25</td>
<td>A</td>
<td>-0.22</td>
</tr>
<tr>
<td>B</td>
<td>N/A</td>
<td>B</td>
<td>N/A</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Location</th>
<th>Slope</th>
<th>Location</th>
<th>Slope</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>-0.21</td>
<td>A</td>
<td>-0.27</td>
</tr>
<tr>
<td>B</td>
<td>N/A</td>
<td>B</td>
<td>0.23</td>
</tr>
</tbody>
</table>
Figure 5.1. Schematic representation of the LSW algorithm for (a) wind event identification and (b) lull event identification with indications of the various components of the algorithm identified.
Figure 5.2. Wind event climatology from the LSW algorithm for the circum-Arctic region for the (a) winter (JFM), (b) spring (AMJ), (c) summer (JAS), and fall (OND), 1979 - 2010. Units are frequency of wind events per season.
Figure 5.3. Same as figure 5.2 except for 48-hour lull events.
Figure 5.4. Wind event trends from the LSW algorithm, 1979 – 2010. Locations with trends significant at the $p < 0.1$ level are contoured and shaded. Locations with trends significant at the $p < 0.05$ level are indicated a black dot. Magnitude of the trend is shown by the color, with increasing trends shown in red and decreasing trends shown in blue. Boxes labeled with letters A. and B have trends highlighted in Table 1.
Figure 5.5. Same as figure 5.4, except for 48-hour lull events. Boxes labeled with letters A. and B have trends highlighted in Table 2.
Figure 5.6. Probability distribution function (PDF) and distribution best fits of all lull events duration, 1979 – 2010. The 48-hour threshold value is indicated on the figure with a vertical red line.
Figure 5.7. Cumulative distribution function (CDF) and distribution best fits of all lull events duration, 1979 – 2010. The 48-hour threshold value is indicated on the figure with a vertical red line.
Figure 5.8. Subregion trends for (a) wind event frequency, (b) 48-hour lull event frequency, and (c) all duration lull frequency for the North Atlantic and Canadian Maritimes. Locations with trends significant at the $p < 0.1$ level are contoured and shaded. Locations with trends significant at the $p < 0.05$ level are indicated a black dot. Magnitude of the trend is shown by the color, with increasing trends shown in red and decreasing trends shown in blue.
Figure 5.9. Climatological mean percentage of time spent in wind event criteria (SP) by season, divided into (a) winter (JFM), (b) spring (AMJ), (c) summer (JAS), and (d) fall (OND). Darker shades indicate higher percentage of time spent in wind event criteria.
Figure 5.10. Same as figure 5.7 except climatological mean percentage of time spent in lull criteria (LP)
Chapter 6. The Potential for Seasonal Forecasting of Winter Storminess Indicators in the North Pacific and Alaskan Regions

Article Information
Chapter 6 has been prepared as a manuscript for submission to a journal.

Abstract
This study tests the statistical relationships and predictability of two measures of storm activity - cyclone track density (TDEN) and wind event frequency - in the North Pacific using teleconnection indices with local influence for the winter period of 1950 - 2012. Two statistical modeling techniques are applied to evaluate linear and non-linear methods of prediction for the region. For both measures of storm activity, the North Pacific index, Niño 3.4 index, and the AO index were found to be the best predictors. Using a 23-year hindcast period (1980 – 2012), the region of highest wind event anomaly prediction skill is located in the western Bering Sea, with hindcast correlation values as high as +0.5 and root mean squared skill scores (RMSESS) 25% higher than climatology. Highest TDEN predictive skill from the 23-year hindcast is found in the southeast region of the North Pacific, near the California coastline, with correlation and RMSESS as high as +0.7 and 30%, respectively. As a result of this study, it could be expected that modestly skillful seasonal outlooks of both wind event anomaly and TDEN anomaly could be produced using forecasted teleconnection indices from a dynamical seasonal prediction system.
6.1 Introduction

Extratropical cyclones (ETCs) influence the dominant weather patterns in many regions of the globe by affecting cloud cover, precipitation, winds, and temperature (Bengtsson et al., 2006; Rodionov et al., 2007; Mesquita et al., 2010; Wang et al., 2012). Many ETCs that occur within the North Pacific bring various impacts to the region through the generation of severe sea states, storm surges, localized coastal erosion, and strong winds (Mason et al., 1993; Blier et al. 1997; Graham and Diaz, 2001; Atkinson, 2005; Mesquita et al., 2008; Mesquita et al., 2010; Pingree-Shippee et al., 2016). Within the marine and coastal context, the frequency of these cyclones plays a key role in defining the “storminess” of a location. In addition to ETCs, strong wind events occur in the North Pacific, and can have similar impacts to those generated by ETCs in the region. It is important to note that wind events can occur both in association with and independent of ETCs. The primary reason for this is simply that wind events are driven by pressure gradients, which are generated in tropical and extratropical cyclones, but which are also generated between regions of non-cyclonic high- and low-pressure areas. In both cases, marine and coastal regions are vulnerable via weather impacts caused by the action of wind on the ocean.

One measure of cyclone activity is the density of storm tracks passing through a specific area, which translates into a track count per unit area (Hoskins and Hodges, 2002; Bengtsson et al., 2006; Mesquita et al., 2010). In the Northern Hemisphere, maximum values of storm track density are represented by features termed the “Aleutian” and “Icelandic” lows, the strength and positions of which provide an aggregate indication of many individual storm track locations, particularly in winter months (Graham and
Diaz, 2001; Rodionov et al., 2007; Zhu et al., 2007). In the North Pacific, the location and intensity of the Aleutian low has been linked to the variability of the storm track (Rodionov et al., 2007; Zhu et al., 2007). Years dominated by a zonal flow across the North Pacific, as measured by the North Pacific index (NP), tend to correspond to a stronger Aleutian Low while years with more meridional flow correspond to a weaker Aleutian Low (Rodionov et al., 2007; Zhu et al., 2007). Aleutian Low strength is also related to cyclone deepening rates, the spatial extent of storms, and the intensity of storms influencing the region (Zhu et al., 2007). Regardless of the apparent strength of the Aleutian Low, however, cyclone frequency, intensity and duration of ETCs in the North Pacific vary little from one year to the next (Zhu et al., 2007; Rodionov et al., 2007).

In addition to the NP index, cyclone activity can also be associated with the main large-scale climate patterns known to influence the North Pacific (Compo and Sardeshmukh, 2004; Rodionov et al., 2007; Zhu et al., 2007; Mesquita et al., 2010). Links between cyclone activity and teleconnections such as the Pacific/North American pattern (PNA), Pacific Decadal Oscillation (PDO), and El Niño-Southern Oscillation (ENSO) have been established (Overland and Pease, 1982; Graham and Diaz, 2001; Allan and Komar, 2002; Compo and Sardeshmukh, 2004; Franzke and Feldstein, 2005; Rodionov et al., 2007; Zhu et al., 2007; Mesquita et al., 2010). Compo and Sardeshmukh (2004) showed that prediction of the mid-tropospheric winter period storm track is possible with modest skill within the Pacific sector in almost all years, including those with weak or non-ENSO events. In general, changes in spatial patterns of cyclone frequency and storm track density are often viewed as modifications in the large scale flow, such as changes in
the Pacific Northwest storm track in association with the Southern Oscillation Index (SOI) (Allan and Komar, 2002).

Some statistical modeling techniques have been used previously to determine the influence of climate indices on storminess in the Northern Hemisphere. Both individual and combinations of teleconnections and their relationship to storm activity have been investigated in previous studies using generalized linear modeling (Mailier et al., 2006; Seierstad et al., 2007). Seierstad et al. (2007) found that in the Pacific Basin, the PNA and PDO were found to be the teleconnection patterns with the strongest association with storm activity, followed closely by the West Pacific Oscillation (WPO) and the East Pacific Oscillation (EPO). Other teleconnections, like ENSO, the Arctic Oscillation (AO), and the NP index are highly correlated with temperature and precipitation patterns for the North Pacific region (Allan and Komar, 2002). ENSO more typically affects larger scale atmospheric patterns of temperature, precipitation, and upper atmosphere features, such as the jet stream and upper level geopotential heights, which also help to drive extratropical cyclone activity.

Other work has suggested that multiple teleconnections have some correlation with temperature and precipitation patterns in the North Pacific and Alaska. In a study of the climate divisions of Alaska, Bienek et al. (2012) found significant (p < 0.05) correlations between many teleconnection indices and station temperatures during fall and winter months. In particular, the Gulf of Alaska, subdivided into the Northeast and Northwest Gulf, showed significant correlation with AO, NPI, PDO, Niño-3.4, SOI, and PNA in the winter months. In the fall months, the East Pacific/North Pacific Oscillation (EP/NP), NPI, PDO, and PNA each showed some influence on the subdivisions of the
Gulf of Alaska. In the Aleutians, the AO and PNA, along with PDO, were each found to be somewhat related to temperatures in the climate zone. Given that temperature changes in the fall and winter months in these regions are primarily driven by cyclones (given that heat transport into the region is large scale and, especially in winter months, driven solely by advection), it is reasonable to expect these teleconnections to also correlate well with storm activity for these same climate zones.

In recent literature, climatic indices and snowpack measurements have been used to create statistical seasonal prediction of temperature, precipitation, and wintertime storminess indicators in the Atlantic basin. In these seasonal predictions, measures such as 10m wind speed and categorical significant wave heights based on the phase of the North Atlantic Oscillation (NAO) have been hindcast with success (Colman et al., 2011; Brands et al., 2012; Brands, 2014). Additionally, a study of seasonal predictability of the North Pacific storm track by Compo and Sardeshmukh (2004) show that some measures of storm track activity, like 500 hPa omega (or vertical velocity), are likely predictable in the Pacific Basin. Sigmond et al. (2013) suggest that some enhanced seasonal forecasting skill exists following sudden stratospheric warming events, though any skill was noted as likely conditional and not predictable more than 1-2 weeks in advance. Given this success using common statistical modeling techniques, the ability to use climate indices to hindcast storminess indicators in the North Pacific basin is evaluated in this study. As such, the following questions drive this study:

i. How well correlated with winter season storminess indicators are those teleconnection indices which are known to be related to North Pacific weather patterns?
ii. How predictable are winter storminess indicators using teleconnection indices in the North Pacific?

The remainder of this paper is structured as follows. Section 2 will overview data sources for teleconnection indices and subsequent variables used in the models. It will also provide a description of the models used within the study. Section 3 will present results of the correlations between different teleconnection indices and storminess measures in the study area. Section 4 will present results from the hindcast modeling and cross validation of storminess indicators using teleconnection indices. Finally, Section 5 will summarize and discuss the main findings of this study.

6.2 Data

Two main data sources are used for predictand data for the modeling efforts. The first is based on the wind event identification algorithm used in Atkinson (2005). This algorithm identifies strong wind events at given grid locations based on wind speed thresholds specified by the user. For this work, wind events with wind speeds of over 10 ms$^{-1}$ are identified. The 10 ms$^{-1}$ threshold has been established as a threshold known to generate impacts on sea state and drive coastal geomorphological processes and follows previous studies of wind events in the North (Solomon et al., 1994; MacClenehan et al., 2001; Atkinson, 2005). The second data source is extratropical cyclone track density (TDEN) as calculated from individual storm tracks output by the Hodges (1994,1995,1999) cyclone tracking algorithm. TDEN provides information on the number of cyclone tracks passing over a given location and can be used as an indicator of how much cyclone activity is occurring in an area. Both data sources utilize the NCEP Reanalysis 1 (NCEP1; Kalnay et al., 1996) dataset, using u and v wind fields from the
surface (10m) to calculate wind speed (for wind events) and at 850 hPa to calculate vorticity (for storm identification and tracking).

For predictors, one measure of ENSO, the Niño 3.4 index, and four teleconnections indices - PDO, NP, PNA, and AO - are used, reflecting the large-scale atmospheric conditions within the North Pacific (Table 1). Teleconnection indices for this study were obtained for the 1950 – 2012 period from the National Oceanic and Atmospheric Administration Climate Prediction Center (NOAA CPC) and the National Center for Atmospheric Research (NCAR) Climate Data Guide. Rather than providing specific definitions of these teleconnections within this paper, a review of individual teleconnection definitions can be found in Hurrell and Trendberth (1994), Panagiotopoulos et al. (2002), and Christensen et al. (2013). Further study of the relationships between teleconnection indices and mean sea level pressure (MSLP) based-measures of mean storminess can be found in Mailier et al. (2006) and Seierstad et al. (2007). The winter season is analyzed in this study, and is defined as the three-month period of January, February, and March (JFM). The study area (Fig. 6.1) stretches across the North Pacific from 30 – 80°N, 100 - 240°W.

Two types of statistical modeling approaches are used in this study to allow for comparison between various methods. The first type is known as Generalized Linear Models (GLMs). GLMs provide a regression framework where the probability distribution and relationship between the mean and variance are specified (Meyers et al., 2002; Wilks, 2011). GLMs have three parts: a linear combination of predictors, a link function, and a variance function and have been used previously to study teleconnection influence on an MSLP based measure of storminess in the North Atlantic and Pacific. All
indices of climate variability and storminess used in this study are detrended to remove the possibility of enhanced or decreased statistical association, which can be caused by time series with similar or opposing trends, respectively (Colman et al., 2011; Brands et al. 2012; Brands, 2014). Before use, storminess indicators are transformed into anomalies and the ascribed probability distribution for the GLM is a normal distribution.

The second modeling technique used in this study is that of “random forest” ensemble regression (Dietterich, 2000; Breiman, 2001; Hsieh, 2009). Ensemble regression melds results from many weak learners, defined as regression trees with lower correlation values, into one high-quality ensemble predictor. Ensemble regression takes a bootstrap sample from the data, fits a regression tree to that sample, and repeats the process many times to increase correlations. Combination of regression trees occurs through averaging. This process uses all predictors at each time step and improves accuracy over typical regression techniques. The ensemble regression method is known to be less sensitive to changes in the predictor/predictand relationship as the averaging over the forest of regression trees is more stable. One disadvantage of the method is that it does not necessarily provide information about the contributions of each predictor to the overall ensemble.

6.3 Results: Correlations

6.3.1 Teleconnections and Wind Event Anomaly

Correlations of the indices of climatic variability and the wind event frequency anomaly storminess indicator are shown in Figure 6.2. Niño 3.4 correlations with wind event anomaly (Fig. 6.2a) show regions of negative correlations over much of the North American continent (-0.2 to -0.4) with additional regions of negative correlations over the
Gulf of Alaska (-0.2) and into the Bering Sea (-0.2). Some isolated regions of positive correlations are found stretching across the central North Pacific from coastal Japan (+0.3) to the region bounded by 45 – 50°N to 135 - 165°W (+0.3 to +0.4).

PDO is found to be negatively correlated with wind event anomaly in the Bering Sea (-0.3 to -0.5), particularly near the Yukon-Kuskokwim (YK) Delta (Fig. 6.2b.). This area of negative correlation extends into the Chukchi and Beaufort Seas (-0.2 to -0.3). A secondary region of negative correlation is found in interior North America, primarily along the Rocky Mountains (-0.4). Two larger sectors of positive correlation are evident in the central North Pacific (+0.3 to +0.4), with the strongest values found in a zone along the 30°N latitude boundary of the study area (+0.5). Other smaller regions of positive correlation are visible in coastal Alaska along the Gulf of Alaska (+0.3) and along the North Slope of Alaska (+0.1 to +0.2).

The NPI shows positive correlations with wind event anomalies (Fig. 6.2c.) in a region stretching from the Kamchatka Peninsula (+0.5) across the western Bering Sea to the YK delta (+0.5). Smaller regions of positive correlation are exhibited over the Aleutians (+0.3) into the Gulf of Alaska (+0.4). Positive correlations also stretch along the axis of the Rocky Mountains from Alaska to the contiguous United States. Areas of negative correlations are found in a zonal region that stretches, in general, along 45°N across the zone 135 – 150°W (-0.5) and to the south along the 30°N border of the study area.

Wind event anomaly correlations with the PNA show very similar patterns to those exhibited by the PDO (Fig. 6.2d.). Areas of positive correlations are found in a relatively zonal region stretching across much of the central North Pacific (+0.2 to +0.5),
with a small extension equatorward to the study area edge at 30°N (+0.5). Large regions of negative correlations are found across the Bering Sea region (-0.5), extending from the Kamchatka Peninsula to the YK delta.

The AO shows smaller areas of strong correlation with wind event anomaly over the North Pacific with respect to the other teleconnections (Fig. 6.2e.). Stronger areas of positive correlation with wind event anomaly are visible in the North American continental region, with values as high as +0.4 over part of the Northwest Territories. Some smaller regions of negative correlation are found over the Bering Strait (-0.30) and off of the Kamchatka Peninsula (-0.4), along with more organized regions of negative correlation off of the coastline of Japan (-0.2 to -0.3).

6.3.2 Teleconnections and Cyclone Track Density (TDEN) Anomaly

Maps of the correlation of the indices of climatic variability and the wind event anomaly storminess indicator are shown in Figure 6.3. Statistically significant regions of positive correlation between Niño 3.4 and TDEN are found in the North Pacific region from the western coastline of North America to the central North Pacific and over central Alaska (+0.5) (Fig. 6.3a.). A small extension of positive correlations is also found in the Gulf of Alaska (+0.2 to +0.3). Some statistically significant negative TDEN correlation values are found in the western Bering Sea and the northern Chukchi/Beaufort Seas, stretching into the Arctic Ocean (-0.2 to -0.3).

In a similar pattern to Niño 3.4, statistically significant positive correlations between TDEN and PDO are found across much of the North Pacific (+0.4 to +0.5), particularly in the region bounded by 35 – 45°N, 135 – 190°W (Fig. 6.3b.). A secondary region of positive correlation is also found in central Alaska (+0.3 to +0.4). Significant
negative correlations are found throughout parts of the Bering Sea, Beaufort Sea, and British Columbia (-0.2).

The correlations between TDEN anomaly and the NPI are strong across much of the study area (Fig. 6.3c.). Statistically significant negative correlations are found at a large number of grid points in the study area including much of the North Pacific (stretching between 135 – 190˚W) and central Alaska (-0.5). Areas of strong positive correlation are found in parts of the North Pacific and into the Bering Sea (+0.5), particularly in the region encompassed by 42 – 50˚N, 170 – 195˚W.

Overall, the correlations between the PNA and TDEN (Fig. 6.3d.) are nearly the inverse of the patterns exhibited between the NPI and TDEN (Fig. 6.3c.). The PNA exhibits significant positive correlations across the North Pacific (+0.5) from 35 – 45˚N and 135– 190˚W. Strong positive correlations are also found across much of the Alaska Interior (+0.5). Significant negative correlations are found along the British Columbia coastline and in parts of the Bering Sea (-0.3 to -0.4).

In general, the correlations between TDEN and AO are confined to much smaller spatial areas than those observed for the first four teleconnection indices (Fig. 6.3e.). For example, significant negative correlations between the AO and TDEN are found over a smaller portion of the North Pacific, in the region from 30 – 45˚N and 120 – 165˚W (-0.3 to -0.5). Areas of positive correlation are confined to the interior North America (Yukon and coastal BC) between 50 – 60˚N and 115 – 135˚W (+0.3 to +0.4).

6.4 Results: Statistical Prediction

Using the GLM and ensemble regression approaches described in section 2, both wind event anomaly and TDEN anomaly are hindcast using the teleconnection indices
observed in the same season as predictors at each of the 1197 grid points within the study domain. The models are initially trained on a 40-year dataset, from 1950 – 1989. The full hindcast period is thus the 23-year period from 1990 – 2012. Yearly storminess indicators are hindcast based off the all available data prior to the date of prediction, such that a prediction for 1990 includes all data from 1950 – 1989, a prediction for 1991 includes all data from 1950 – 1990, and so on until the final hindcast year of 2012. All data in all cases are detrended and have the local climatological mean removed. Root mean squared error skill score (RMSESS) is applied as a skill measure to assess the model performance against climatology (based on the 1950 – 1989 training period), which is used to represent a forecast with zero skill. RMSESS will indicate value added by using either GLM or ensemble regression forecasting techniques by giving the percentage by which the model outperforms climatology.

In the wind event anomaly hindcast, the maximum values of hindcast correlation as predicted by the GLM are found in the region near Kamchatka Peninsula, stretching into the western Bering Sea (Fig. 6.4 a.,b.), with hindcast correlation ($r_{\text{hind}}$) and RMSESS values of +0.5 to +0.7 and 20-30%, respectively. The maximum values predicted by the ensemble regression are found in similar but more spatially distinct areas, such as the central Bering Sea (Fig. 6.4 c.,d.), with the $r_{\text{hind}}$ values between +0.3 to +0.5. RMSESS for the GLM method is higher in the regions with maximized $r_{\text{hind}}$, particularly on the western edge of the Bering Sea and near the Kamchatka Peninsula. Some locations without significant $r_{\text{hind}}$ show RMSESS values that are slightly better than climatology, but are not noted as regions with potential forecast skill (e.g. 15 – 20% in the Gulf of Alaska).
Hindcasts of the TDEN indicator produce much higher $r_{hind}$ and RMSESS values across the study area than the wind event frequency anomaly indicator. Values of $r_{hind}$ ranging between +0.5 to +0.7 are found across much of the North Pacific, with the maximum potential skill from the GLM method found in the region bounded by 30°-45°N, 130°-165°W (Fig. 6.5 a.,b.), with $r_{hind}$ and RMSESS values of +0.7 and 30%, respectively. The maximum values as predicted by the random forest ensemble regression (Fig. 6.5 c.,d.) are found in the same region, again with $r_{hind}$ and RMSESS values of 0.5 to 0.7 and 15 – 25%, respectively. Both the $r_{hind}$ and RMSESS values for the random forest ensemble regression are more spatially constrained than in the GLM-based hindcast, though the primary area of TDEN correlations in the southeastern portion of the study area are well predicted as previously highlighted.

6.5. Discussion and Conclusion

In this work, the potential for seasonal predictability of two storminess indicators (cyclone track density and wind event frequency) using winter teleconnection indices is investigated. The two statistical techniques (generalized linear models and random forest ensemble regression models) are evaluated for their ability to hindcast the two storminess indicators within the North Pacific region. Results from this study suggest that modest skill in the hindcasts for both storminess indicators when using the GLM methods, particularly in the case of TDEN prediction. The local correlations (Figs. 6.2 and 6.3) are indicative of a similar pattern in the best predictors within the GLMs. Teleconnections associated with the strongest relationship with wind event anomaly were the NP and the AO. The NP showed the highest correlations with the wind event anomaly (Fig 6.2c.),
particularly in the western Bering Sea and Sea of Okhotsk. The GLM prediction of TDEN had the strongest relationship with Niño 3.4, NP, and AO.

The correlations with various teleconnections identified in this study reinforce the findings of Compo and Sardeshmukh (2004), with TDEN correlation with teleconnection patterns being stronger than correlations observed between the same indices and wind event anomaly. Teleconnections with centers of action in the study area, such as the NP, PDO, and PNA, have strong correlations with both of the storminess indicators, though in very different spatial patterns. Significant correlations with the NP index show that there is a strong relationship between Bering Sea wind event anomaly patterns and the strength of the wintertime Aleutian Low. Likewise, TDEN anomaly in the region off of the California coastline is correlated well with a strong Aleutian Low.

Comparing the results from the two storminess indicators, the spatial patterns of higher correlation and skill vary greatly between wind event anomaly and TDEN anomaly. Highest skill regions of TDEN anomaly prediction were found in the southeastern part of the study region (30 – 45°N, 120 - 155°W), though patterns of skill show a much more expansive area than those found with wind event anomaly. The highest skill regions of wind event anomaly prediction were confined in both the GLM and random forest ensemble regression methods to the western Bering Sea. For both storminess indicators, the regions with the maximum hindcast skill reflects the patterns of local correlation with various teleconnections (Figs. 6.2, 6.3)

Strong correlations between TDEN and the Niño 3.4 index suggest that the subtropical sea surface temperature (SST) anomalies have a strong relationship with the local anomaly of TDEN, particularly in the southeastern portion of the study area, where
the positive phase of ENSO typically drives more cyclones towards the California coastline. This relationship between TDEN anomaly and Niño 3.4 is reinforced by the highest correlated teleconnections in the GLM, where the primary significant teleconnections were the Niño 3.4, the NP index, and the AO.

Regions highlighted by significant $r_{\text{hind}}$ and RMSESS for TDEN also show a strong local correlation with the NP index (Fig. 6.2). Previous studies have shown that the strength of the Aleutian Low is correlated with the amplification of the Pacific storm track (Zhu et al., 2007). The strength of the Aleutian Low has also been correlated with the phase of ENSO, where an intensification of the Aleutian low has been shown to be related to the El Niño phase and a weakening of the Aleutian Low has been shown in La Niña events (Neibauer, 1988; Rodionov et al., 2007; Zhu et al., 2007). As such, one would expect the NP index is correlated with the lagged tropical Pacific SST changes, such as those in the Niño 3.4 region (Neibauer, 1988; Trendberth and Hurrell, 1994; Rodionov et al., 2007). Therefore, ENSO plays a strong role in the phase of the NP index (Rodionov et al., 2007; Zhu et al., 2007). This suggests that even with strong correlations between TDEN and the NP index, the relationship between the predictability of the TDEN anomaly is heavily reliant on the phase of ENSO. This relationship is likely the reason for the higher forecast skill seen in GLM predictions of TDEN than wind event anomaly, as local correlations of Niño 3.4 with TDEN are stronger than those with wind event anomaly. In general, the local correlations (Figs. 6.2, 6.3) between each teleconnection index and TDEN are found to be strongest in the region of highest hindcast skill as measured by both $r_{\text{hind}}$ and RMSESS (Figs. 6.4, 6.5).
GLM skill for wind events (Fig. 6.4 a.,b.) is marginally higher and encompasses a larger area than that found in the random forest ensemble regression (Fig. 6.4 c.,d.). For example, the GLM method shows modest skill in the western Bering Sea and near the Kamchatka Peninsula that is missing in the random forest method. However, the overall pattern of the \( r_{\text{hind}} \) correlation and RMSESS for 23-year hindcast of wind event anomaly is similar for both methods, though the random forest method highlights smaller spatial regions in the central Bering Sea, particularly in the \( r_{\text{hind}} \) assessment.

In the case of both storminess indicators, the GLM method outperformed the random forest ensemble regression as shown by both \( r_{\text{hind}} \) and RMSESS values, particularly in the case of TDEN predictions. This finding is not completely unexpected, as the architecture of random forests is likely to degrade the final prediction due to the inclusion of weak learners that are potentially poor predictors. Also, the use of bootstrapping methods, such as bagging, reduces the initial sample size such that the averaging of the learners in this case likely decreases the skill of the final prediction.

Finally, in this study, all predictions of winter storminess indicators are tied directly to teleconnection indices for the simultaneous season. Therefore, the relationship specified within this study is indicative of forecast skill when using teleconnection indices with negative lead time; that is, a lead time requiring knowing the teleconnection indices prior to forecast. Assuming that a dynamical seasonal forecasting system could reproduce a stable ENSO signal and the other teleconnections used in this study, it could be expected that modestly skillful seasonal outlooks of both wind event anomaly and TDEN anomaly could be produced.
6.6 Future Work

Each of the indicators used as predictands in this study use near surface conditions to establish storminess. As suggested by Compo and Sardeshmukh (2004), additional skill may be available when using atmospheric fields that are higher in the atmosphere, such as 500 hPa omega, which provide more potential predictability in the winter season. Future efforts would use the same statistical techniques applied to mid-tropospheric based storminess indicators within the study region to investigate if seasonal predictability can be increased. Ultimately, near-surface storminess indicator predictability may be increased by combining information about wind events and track density into a single indicator.

Additional future efforts would look into more statistical methods for increasing predictable skill within the region, including neural networks and other non-linear methods. It is likely that the best prospects for predictable skill will be found in dynamical seasonal forecasting models with a greater understanding of the physics of storminess processes on a seasonal scale. Finally, an important point is that work from this study is not intended to replace any output from current seasonal forecasting systems already in place at major modeling centers. Instead, the goal is to supplement and enhance current forecasting efforts by strengthening the understanding of the relationships between known indices of climatic variability in the North Pacific region and storminess indicators. The methods described in this study would require predicted teleconnections to be calculated from a suite of dynamical forecasting systems.
Table 6.1. Teleconnection/Climate indices included in both GLM and random forest regression models. Indices were downloaded from NCAR/UCAR Climate Data Guide (http://climatedataguide.ucar.edu) and NOAA Climate Prediction Center (http://cpc.ncep.noaa.gov) in March 2014.

<table>
<thead>
<tr>
<th>Teleconnection</th>
<th>Abbreviation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Niño 3.4</td>
<td>N34</td>
</tr>
<tr>
<td>Pacific Decadal Oscillation</td>
<td>PDO</td>
</tr>
<tr>
<td>North Pacific Index</td>
<td>NP</td>
</tr>
<tr>
<td>Pacific – North American Pattern</td>
<td>PNA</td>
</tr>
<tr>
<td>Arctic Oscillation</td>
<td>AO</td>
</tr>
</tbody>
</table>
Figure 6.1. Study area map with key locations highlighted.
Figure 6.2. Spearman rank correlation coefficient (R) between winter (JFM) teleconnection index values and wind event anomaly for (a) Nino 3.4, (b) PDO, (c) NP index, (d) PNA, and (e) AO. Grid locations with significant correlation (Alpha = 0.05) are highlighted by white symbols. Time period for correlation analysis is 1950 – 2012 and R is multiplied by 100.
Figure 6.3. Same as Figure 6.2, except for cyclone track density (TDEN) anomaly.
Figure 6.4. Hindcast skill obtained from the use of GLM and Random Forest ensemble regression with wind event anomaly as the predictand. Spearman rank coefficient with hindcast ($r_{\text{hind}}$) with significant positive correlations ($\alpha = 0.05$, $r_{\text{hind}} > +0.337$) for the a) 23-year GLM hindcast and c) 23-year Random Forest hindcast. As shown are the root-mean squared skill score (RMSESS) values above zero with reference to climatology for the a) 23-year GLM hindcast and the d) 23-year Random Forest hindcast.
Figure 6.5. As in Figure 6.4, except for TDEN.
Chapter 7 Conclusion

7.1 Introduction

This dissertation presents three analyses that explore the use of different representations of how a storm is defined and various perspectives of storminess. Two of these analyses focus on the North Pacific and Alaskan regions while one presents an analysis of new storminess indicators from the circum-Arctic perspective. The first paper investigates the ability to increase the temporal coverage of the North Pacific extratropical cyclone climatology back to the beginning of the Twentieth Century Reanalysis (20CR) in 1871. This first paper also compares two frequently used storm tracking algorithms with varying atmospheric fields for cyclone identification and tracking that have been used within the North Pacific region. The second paper applies an established wind event identification algorithm to modern reanalyses for the circum-Arctic. This analysis is coupled with a new, user-influenced algorithm that also identifies breaks between wind events, termed “lulls”, to establish the Lull/Storm Wind (LSW) algorithm that evaluates the frequency of wind event and lull periods for the circum-Arctic. The third paper applies two statistical techniques, that of Generalized Linear Models (GLMs) and “random forest” ensemble regression (RF), to predict winter season wind and cyclone activity indicators using teleconnections known to influence the climate of the North Pacific as predictors. Each of these papers have enhanced the knowledge surrounding North Pacific and circum-Arctic storminess regimes and their predictability.

The work for this dissertation was guided by the overarching hypothesis: The definition of a storm impacts the ability to define the statistical relationships between
measures of climatic variability (e.g. teleconnections) and the storminess indicator. From the objective tracking algorithm perspective, it was found that objective algorithms operating with different core definitions of a storm managed to generate similar climatological results for storm frequency, track density, cyclogenesis density, and cyclolysis density statistics. Storm tracks from the relative vorticity based Hodges algorithm are smoother, longer lasting, and slightly more numerous than those from the Serreze sea level pressure based algorithm. When compared to the results from the lull/storm wind indicator, storm track density shows higher correlations with major teleconnection patterns and better predictability with statistical modeling techniques, outperforming the wind event frequency measures for the North Pacific. This suggests that statistical prediction of storm track density is more likely to succeed than predictions of wind event frequency using the same measures of climatic variability. Additionally, statistical relationships between extratropical cyclones and measures of climatic variability are stronger than with the pragmatically defined lull/storm wind indicator.

7.2 Main Research Results and Key Points

1. It was shown that regardless of the: a) storm identification and tracking algorithm or, b) atmospheric field of analysis (MSLP or 850 hPa relative vorticity), similar climatological frequencies of track density, cyclolysis density, and cyclogenesis density are found in all seasons. This is important as differences in objective definition of a storm from the tracking perspective are not found to highly influence the final climatology of storms. Winter cyclolysis and track density is slightly lower in the Serreze MSLP based method than the Hodges relative vorticity based method. Track density is higher in the Hodges method in nearly all
seasons, indicating that more, longer lived disturbances are found in the Hodges method than the Serreze method. This was demonstrated to be a result of the relative time scales of disturbances in the vorticity field when compared to sea level pressure. Additionally, this is likely related to the requirement of a closed isobar at the 2 hPa level in the Serreze method that differs greatly from the cost function used in the Hodges method (Chapter 4; Thesis Objective 1)

2. The 20CR storm track dataset provides a temporal extension to the climatology of homogeneous ETC characteristics, beginning as early as 1920 for some subregions of the North Pacific and Alaskan regions. In the Gulf of Alaska and Bering Sea, homogeneous time series of cyclogenesis and cyclolysis are found back to 1920, which is a considerable extension over the previous longest reanalysis period for the region (NCEP1, beginning in 1948, though with homogeneity issues in the long period). There are issues related to inhomogeneities for both cyclogenesis and cyclolysis densities in the northern subregions of the study area, including the Alaska Interior and Chukchi/Beaufort Sea, where lack of data coverage in the early 20th century impacts the homogeneity of the storm track dataset. (Chapter 4; Thesis Objective 1)

3. Interviews with marine tug-and-barge operators in the Alaskan and Canadian North indicate the importance of wind, particularly wind speed, for safe operations in the North. Through the interview process, it was established that the occurrence of both wind events, which can shut down operations, and lull events, which allow for safe travel between points of safe harbour, were key to understanding the impacts of weather on the tug-and-barge industry. The
occurrence of lull periods lasting 48-hour or longer between strong wind events was key to the ability to navigate between points of safe harbour and is a novel result from this work. (Chapter 5; Thesis Objective 1)

4. A circum-Arctic climatology of wind and lull event frequencies using the Lull/Storm Wind (LSW) algorithm for the 1979 – 2010 period based on the ERA-Interim reanalysis was established based on user-influenced criteria for storm identification from an Eulerian perspective. Winter and fall are the primary stormy seasons in the North Atlantic and North Pacific, with decreases in wind event frequency by 50% in the spring and summer with respect to the winter and fall. Lull event frequency is highest in the spring and summer, though it is a more complicated interpretation due to the effects of wind events and local storminess impacts. Areas with the highest climatological mean frequency of wind events tend to also have the largest number of lull events (due to short wind events and short lulls). When expressed as percentage of a season spent “stormy” or “non-stormy/lully,” locations with high storminess percentages exhibit lower lulliness percentages at that same time. Regions that spend time in the near 50% storm and lull percentages seasonally are generally better climates for socio-economic activities due to extended periods of calmer weather in between storms, which is evident in the Canadian Arctic Archipelago where tug-and-barge type transportation occurs regularly (Chapter 5; Thesis Objective 2)

5. Lull periods are an important consideration for northern marine operations in any facet, including economic and subsistence based, particularly in the tug-and-barge shipping operations that resupply much of the North. Lull periods of 48-hours or
longer are ideal for shipping, especially within the Alaskan region. Analysis shows a decreasing trend for lulls of this duration in the North Atlantic and the central Arctic in many seasons, indicating less frequent lull periods following wind events. (Chapter 5; Thesis Objective 2)

6. In the North Atlantic region, opposing trends related to lull frequency were found when delineated by a duration threshold of 48 hours or longer, and when no duration threshold is applied. The frequency of all lull events shows a statistically significant increasing trend while, when imposing the 48-hour duration threshold, the frequency of events shows a decreasing trend. This implies a change in the mean duration of lull events to shorter periods. In light of increasing frequency of short period lulls (6 – 48 hrs) and decreasing frequency of longer period lulls (48 hrs or longer), an overall increase in wind event based storminess in found in the North Atlantic over the 1979-2010, due to a decrease in the length of the lull that follows a wind event. (Chapter 5; Thesis Objective 2)

7. Two forms of statistical models, Generalized Linear Models and random forest ensemble regression, were used to generate skillful hindcasts of two storminess indicators (wind event anomaly and cyclone track density anomaly) using teleconnection indices in the winter as predictors. GLMs using selected teleconnections (Niño 3.4, the NP index, and the AO) produce areas of hindcast skill in the western Bering Sea for wind event anomaly (RMSESS between 15 – 25%) as the primary storminess indicators and. Skillful predictions of cyclone track density anomaly hindcasts (RMSESS between 30%) are found for much of
the study area when using GLMs, with the highest skill found along the west coast of California. (Chapter 6; Thesis Objective 3)

7.3 Conclusion

The results from this dissertation show that climatological patterns of storminess depend more on the perspective by which storminess is being assessed and less on the individual algorithms being employed within each perspective. A dependence on approach perspective suggests that the way a storm is observed differs based on the needs related to the situation. For example, when using the theoretical and tracking perspectives, the choice of storm tracking algorithm results in small differences in the general characteristics of extratropical cyclones within the North Pacific, but does not modify the overall climatological patterns (i.e. locations of maxima/minima for various metrics). The Hodges relative vorticity based method tends to identify more cyclones and begin cyclogenesis earlier than the Serreze method for the study area, which is related to both the timescale of the tracked field and the atmospheric field used.

In the case of the 20CR, frequency of cyclogenesis and cycloysis events were found to be temporally homogeneous in the Bering Sea and Gulf of Alaska after the mid-twentieth century. Interannual variability in the Bering Sea is apparent in both the cycloysis and cyclogenesis time series, with shifts from lower activity to higher activity in the mid-1970s. Previous work has noted a climatic shift of the Alaska region as observed in the PDO at that time (Mantua et al., 1997; Mantua and Hare, 2002). Trends in both sea surface temperature and sea surface temperature gradient have been shown to be positive in the Bering Sea during the 1979 – 2010 period (Mesquita et al., 2010). However, no statistically significant trends in cyclogenesis are found within the Bering
Sea subregion at this time in either the 20CR or NCEP1 reanalyses (Table 4.3, Table 4.4). This echoes the findings of Mesquita et al. (2010) that indicated increases in SST do not necessarily translate into an increase of storm activity in the region. These results are important, and emphasize the findings that variability of storminess indicators in the 20CR time series (particularly cyclogenesis and cyclolysis densities) are controlled less by SST and SST gradient and more by interannual variability of the climate. In the Chukchi/Beaufort subregion, three distinct shifts in the mean of the time series are observed in the 1960 – 1980 period; one in the cyclogenesis time series (1974) and two in the cyclolysis time series (1966, 1977). These shifts coincide with large scale changes to the preferred storm track in the region during these periods, as discussed in the studies of Saveliava et al. (2000) and Atkinson (2005). These shifts explain the cyclical variability in the frequency of cyclolysis and cyclogenesis within the region, as shown in Figure 4.13a and Figure 4.13c.

From the pragmatic perspective, storminess indicators that are rooted in weather conditions that create impacts describe storm climatology differently than cyclone identification algorithms. This result is important in that the locations where wind events occur are not always associated with the center of a cyclone, mainly due to the dependence on local pressure gradients and the large scale weather patterns. This means that wind events can occur in periods that might otherwise be considered “non-stormy” times. This is important from the perspective of someone on the ground that is concerned with conducting socio-economic activity, as periods of what otherwise might be considered fair weather can still lead to conditions that are relatively hazardous.
The establishment of new storminess indicators relating climatological data to the needs of end-users is important in the context of the North. A need for knowledge of periods of breaks in stormy periods with respect to wind events (i.e. lulls) is driven by the marine shipping interests in the Arctic, where resupply of many villages is undertaken by tug-and-barge style shipping. Tug-and-barge style shipping is highly vulnerable to periods of storminess due to the flat-bottom design of shipping vessels and cargo. Outside of the typical means of defining a storm, this analysis allows for the understanding of periods of lulls (48-hr or greater periods of calmer weather), which allows for movement between points of safe harbour. As wind event related storminess can occur without the presence of an extratropical cyclone, these lull periods are crucial to the understanding of climatological patterns that are favourable for socio-economic activity in the North.

The seasonal predictability of storminess indicators such as cyclone track density and wind event frequency within the North Pacific was shown to have modest skill when using teleconnection indices with local influence in the region. Track density anomaly was well hindcast by two structurally different statistical methods, showing modest skill when compared to climatology in the southern parts of the study area near the California coastline. Wind event anomaly prediction showed skill along the western Bering Sea and Kamchatka peninsula when using Generalized Linear Models, but very little skill using other statistical techniques. In hindcasts of both metrics, the North Pacific index, Niño 3.4, and the Arctic Oscillation were the teleconnections that showed the highest correlation and influence over predictability in the region. The potential predictability shown provides evidence that statistical models of both storminess indicators may be able to be used to supplement seasonal outlooks of storminess from dynamical seasonal
forecasting systems in the winter season in regions where the skill was demonstrated to be better than climatology, particularly if these dynamical systems are able to skillfully predict teleconnection indices. For example, the Canadian Seasonal to Inter-Annual Prediction System currently predicts ENSO relatively well and could potentially provide dynamical input for this type of forecasting system.

7.4 Future Work

Each of the preceding key points provides an impetus for future work within both the storminess definition and seasonal predictability areas of research for the North Pacific and Alaskan regions. Future work points that have been unveiled by this work are:

- As noted in Chapter 5, the established LSW algorithm shows potential for becoming a short- to mid-term forecast product by incorporating forecast model data. To do this would mean expanding the current circum-Arctic LSW domain equatorward to incorporate the eastern and western seabords of North America, Gulf of Mexico, and the central North Pacific (including Hawaiian Islands).
- In order to establish the LSW indicator as a potential forecast product, a pseudo-operational environment is needed to establish the mechanics of an “end-to-end” implementation, including assessment of indicator skill. This has been discussed with the NOAA/NWS Alaska Region Climate Test Bed in Anchorage, Alaska.
- One future research direction is the evaluation of the physical mechanisms responsible for patterns of cyclone activity and storminess indicators in the North Pacific and circum-Arctic regions, which have been identified in this dissertation. The exploration of these mechanisms could provide new insights explaining
why changes have occurred in cyclones and wind/lull events throughout the 20th century.

- The results returned by Lagrangian storm tracking algorithms often do not include information about associated windfields – extent, location, or magnitude. In some Lagrangian schemes, it is possible to add winds and location to tracks, but this is typically not part of the basic form of the methodology. Creating an integrated method that combines both outputs from tracking algorithms and the LSW algorithm would allow for a new means of storm analysis. Additionally, this could provide information about cyclones that generate strong wind events and those that do not, which is also of great interest to marine stakeholders in the North.

- Another interesting question for further evaluation of the seasonal predictability of storminess indicators relates to the magnitude of skill required to provide useful forecasts to end-users. The 30% improvement over climatology shown in Chapter 6 provides modest skill, but is only found in small portions of the study area, and none of those are found directly within the Alaskan region except for a very small region of the Bering Sea. Therefore, for end-users in the Alaskan region, how much skill over climatology is necessary to provide useful information for operators.
References


Simmonds, I., R. J. Murray, and R. M. Leighton, 1999: A Refinement of Cyclone Tracking Methods with Data from FROST. *Australian Meteorological Magazine* (Special ed.), 35–49.


164


