The Influence of the Arctic Frontal Zone on Summer Cyclone Activity Today and in the Future

Alexander D. Crawford-Alley

University of Colorado at Boulder, alexander.crawford@colorado.edu

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written by Alex D. Crawford
has been approved for the Department of Geography

___________________________
(Mark C. Serreze)

___________________________
(Holly Barnard)

___________________________
(Peter Blanken)

___________________________
(John Cassano)

___________________________
(Jennifer Kay)

Date __________________________

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Of scholarly work in the above mentioned discipline.
Extratropical cyclone activity over the central Arctic Ocean reaches its peak in summer. Along with local genesis, previous research has argued for the existence of two major external cyclone source regions contributing to this summer maximum: the Eurasian continent interior and a narrow band of strong horizontal temperature gradients along the Arctic coastline known as the Arctic frontal zone (AFZ). This study incorporates data from an atmospheric reanalysis and an advanced cyclone detection and tracking algorithm to critically evaluate the relationship between the summer AFZ and cyclone activity in the central Arctic Ocean. Next, it uses the Community Earth System Model Large Ensemble to assess how the AFZ, Arctic cyclone activity, and the relationship between them respond to a global warming scenario.

Analysis of both individual cyclone tracks and seasonal fields of cyclone characteristics shows that the Arctic coast (and therefore the AFZ) is not a region of cyclogenesis. Rather, the AFZ acts as an intensification area for systems forming over Eurasia. As these systems migrate toward the Arctic Ocean, they experience greater deepening in situations when the AFZ is strong at midtropospheric levels. On a broader scale, intensity of the summer AFZ at midtropospheric levels has a positive correlation with cyclone intensity in the Arctic Ocean during summer, even when controlling for variability in the northern annular mode. Taken as a whole, these findings suggest that the summer AFZ can intensify cyclones that cross the coast into the Arctic Ocean, but focused modeling studies are needed to disentangle the relative importance of the AFZ, large-scale circulation patterns, and topographic controls.

Under a strong warming scenario, the AFZ remains a significant cyclone intensifier, and changes to the AFZ are largely restricted to June. The AFZ develops earlier in the year and appears stronger in June. This strengthening is accompanied by enhanced cyclogenesis along the east Siberian coast in June, but no change is observed for overall cyclogenesis or cyclone frequency and intensity in the Arctic. Cyclone-associated precipitation rises in all summer months, but this is likely driven by thermodynamic changes, not changes to cyclone development.
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CONTENTS

CHAPTER 1. INTRODUCTION 1

CHAPTER 2. OUR DEVELOPING VIEW OF CYCLONES IN THE ARCTIC 5

2.1. Dispelling the High Pressure Myth 5
2.2. Cyclone Seasonality 6
2.3. The Arctic Frontal Zone 9
2.4. The Arctic Ocean as Cyclone Graveyard 10
2.5. Trends and Interannual Variability in Arctic Cyclone Activity 10
2.6. Unanswered Questions 12

CHAPTER 3. REVIEW OF ARCTIC FRONTAL ZONE STUDIES 14

3.1. The Synoptic-Scale Cyclone Framework 14
3.2. The Arctic Air Mass Framework 16
3.3. A New Look at the Arctic Frontal Zone 18
3.4. Baroclinic Instability: Linking the AFZ to Synoptic-Scale Cyclones 24

CHAPTER 4. REVIEW OF CYCLONE DETECTION AND TRACKING METHODS 32

4.1. Introduction 32
4.2. 1830s-1980s: Manual Tracking 33
4.3. 1990s - Present: Automated Tracking 37
4.4. Ensembles and Sensitivity Studies 58
4.5. An Alternative: Eulerian Approaches to Synoptic Studies 60

CHAPTER 5. DATA & METHODS 62

5.1. MERRA, CFSR, and ERA Reanalyses 62
5.2. The CESM Large Ensemble 64
5.3. Ancillary Data 67
5.4. Defining AFZ Parameters 68
5.5. Clustering Analysis and AFZ Sectors 71
5.6. Cyclone Detection and Tracking 72
5.7. Seasonally Variable Northern Annular Mode (SVNAM) Index 83
5.8. Statistical Comparison of AFZ and Cyclone Characteristics 84
5.9. Climate Change Assessment for 2071-2080 86

CHAPTER 6. THE AFZ AND SUMMER CYCLONE DEVELOPMENT 88

6.1. The AFZ and Baroclinic Instability 89
6.2. Results from Cyclone Analysis 91
6.3. Discussion & Limitations 100
6.4. Conclusions 104

CHAPTER 7. THE AFZ IN A WARMING WORLD 106

7.1. Introduction 106
7.2. The AFZ in CESM-LE
7.3. RCP8.5: The Global Warming Context
7.4. The AFZ Under RCP8.5 (1990-2005 v. 2071-2080)
7.5. Observed Trends in the AFZ
7.6. Conclusions

CHAPTER 8. SUMMER ARCTIC CYCLONES IN A WARMING WORLD
8.1. Introduction
8.2. Validation of Arctic Cyclones in CESM-LE
8.3. Arctic Cyclones in RCP8.5 (1979-2005 v. 2071-2080)
8.4. Observed Trends in Summer Arctic Cyclone Activity (1979-2014)
8.5. Discussion of Limitations
8.6. Conclusions

CHAPTER 9. CONCLUSION

REFERENCES

APPENDICES

LIST OF ABBREVIATIONS

APPENDIX TO CHAPTER 5
5A.1. Input Parameters
5A.2. Outputs
5A.3. Limitations
5A.4. Comparisons to Other Algorithms
5A.5. Influence of Input Data
5A.6. Tracking Cyclone Centers Versus Cyclone Systems

APPENDIX TO CHAPTER 6
6A.1. Thermal Tide Relationship by Season
6A.2. Residuals versus Fitted Values for Regression Model for 700 hPa AFZ

APPENDIX TO CHAPTER 7
7A.1. Near-Surface Expression
7A.2. Seasonality of the AFZ
7A.3. Vertical Expression
7A.4. Spatial Variability of the AFZ
7A.5. Interannual Variability of the AFZ

APPENDIX TO CHAPTER 8
8A.1. Basic Cyclone Characteristics of CESM-LE
8A.2. Arctic Cyclone Origins
8A.3. The Relationship between the AFZ and Arctic Cyclones
TABLES

6.1. Coefficients for multiple linear regression models describing the
depthening rate of summer cyclones as they pass through the AFZ ........................................73
6.2. Correlations between AFZ strength and cyclone characteristics ........................................74
6.3. Correlations between AFZ strength and cyclone characteristics
after adjusting for SVNAM ........................................................................................................74
6.4. Correlation between AFZ strength and MAF strength in summer .........................................78

7.1. Mean summer temperature differences under RCP8.5 ..........................................................83

8.1. Changes in summer cyclone variables under RCP8.5 ............................................................108
8.2. Change in intensity of cyclones whose tracks intersect the
CAO+BCEL in June under RCP8.5 ..............................................................................................109
8.3. Change in intensity of cyclones whose tracks intersect the
CAO+BCEL in August under RCP8.5 ........................................................................................110
8.4. Trends in summer cyclone characteristics for 1979-2014 .....................................................114
FIGURES

1.1. Location map of Arctic region ........................................................................................................... 1

2.1. A typical depiction of the three-cell model ......................................................................................... 4

3.1. Mean June energy balance along the Alaska-Arctic Ocean coastline .................................................. 10
3.2. Mean July 2 m temperature gradient magnitude for 1979-2012 ....................................................... 13
3.3. Monthly climatology (1979-2012) of surface energy fluxes .............................................................. 14
3.4. Latitudinal cross sections of mean July meridional temperature gradient and zonal wind velocity ............................................................................................................. 16
3.5. Change in the geostrophic wind with altitude in three equivalent barotropic atmospheres .......................................................... 19

4.1. Example SLP pressure field .................................................................................................................. 35
4.2. Example of cyclone area interference .................................................................................................. 37

5.1. Mean July temperature gradient magnitude at 700 hPa for 1980 and 1994 ........................................ 51
5.2. Effect of minimum SLP gradient of cyclone center detection ............................................................ 54
5.3. Example of multi-center cyclone detection ......................................................................................... 57
5.4. Example of cyclone tracking methodology ......................................................................................... 59
5.5. Schematic of seven cyclone tracking events ......................................................................................... 60

6.1. Connection between AFZ and baroclinicity ....................................................................................... 66
6.2. Cyclone characteristics for 1979-2014 ............................................................................................... 67
6.3. Lysis minus genesis frequency, 1979-2014 ....................................................................................... 68
6.4. Distribution of cyclones entering the CAO+BCEL by source region .................................................. 69
6.5. Cyclone tracks originating in eastern Siberia, 1979-2014 ................................................................. 70
6.7. GPH at 500 hPa during summer cyclogenesis in the Kolyma Lowland .............................................. 76

7.1. Arctic warming under RCP8.5 .............................................................................................................. 84
7.2. Change in summer sea ice and snow cover under RCP8.5 ................................................................. 85
7.3. Change in near-surface AFZ strength under RCP8.5 ....................................................................... 87
7.4. Change in AFZ strength seasonality under RCP8.5 ...................................................................... 88
7.5. Change in June meridional temperature gradient and zonal wind velocity under RCP8.5 .................................................................90
7.6. Change in July meridional temperature gradient and zonal wind velocity under RCP8.5 .................................................................91
7.7. Change in August meridional temperature gradient and zonal wind velocity under RCP8.5 .................................................................92
7.8. Change in mean latitude of the migrating Arctic Front under RCP8.5 .........................................................................................94
7.9. Linear trends in AFZ strength at 2 m for 1979-2014 ..................................................................................................................95
7.10. Linear trends in summer sea ice and snow cover for 1979-2014 ......................................................................................95
7.11. Trend in June meridional temperature gradient and zonal wind velocity for 1979-2014 ............................................................96

8.1. Change in cyclone track density under RCP8.5 .....................................................................................................................102
8.2. Change in cyclogenesis frequency under RCP8.5 .............................................................................................................103
8.3. Linear trends in winter sea ice and snow cover under RCP8.5 ..............................................................................................105
8.4. Change in cyclogenesis frequency for CAO+BCEL cyclones under RCP8.5 .................................................................106
8.5. Change in cyclone tracks originating in eastern Siberia under RCP8.5 ................................................................................107
8.6. Change in AFZ-cyclone seasonal correlations under RCP8.5 .........................................................................................111
8.7. Change in AFZ-cyclone seasonal correlations under RCP8.5 after SVNAM adjustments ......................................................................112
8.8. Trend in cyclogenesis frequency for 1979-2014 ...................................................................................................................113
CHAPTER 1. INTRODUCTION

The Arctic climate system is strongly shaped by seasonal and spatial patterns of extratropical cyclone activity (Serreze et al. 2007; Serreze and Barrett 2008; Sorteberg and Walsh 2008; Iijima et al. 2016). As the Arctic Ocean becomes more accessible to shipping, resource extraction and other activities, the practical importance of understanding Arctic weather patterns grows (ACIA 2005; Dodds 2010). While recent years have seen strong advances in understanding the seasonal cycles (e.g., Serreze et al. 1993; Serreze 1995; Simmonds et al. 2008), interannual variability (e.g., Zhang et al. 2004; Serreze and Barrett 2008; Simmonds et al. 2008), and long-term changes (e.g., Serreze et al. 1993; McCabe et al. 2001; Vavrus 2013) in the characteristics, distribution and climate impacts of Arctic cyclones, much remains to be learned.

The central Arctic Ocean (CAO; inside gray outline in Figure 1.1) and the adjacent Beaufort, Chukchi, East Siberian, and Laptev Seas (BCEL) experience substantially more cyclolysis than cyclogenesis in all seasons (Serreze et al. 1993; Serreze and Barrett 2008). In other words, more cyclones fill and dissipate in the Arctic Ocean than form there (Reed and Kunkel 1960; Whittaker and Horn 1984). A detailed review of literature regarding Arctic cyclone activity can be found in Chapter 2, but the most important points are noted here.

Cyclonic activity is common over the Arctic Ocean in both winter and summer (Serreze et al. 1993; Zhang et al. 2004; Mesquita et al. 2008; Serreze and Barrett 2008; Simmonds et al. 2008), but the source regions for winter and summer storms are distinct. In winter, a majority of cyclones that migrate into the Arctic Ocean originate within the Icelandic Low region and elsewhere along the North Atlantic storm track (Serreze et al. 1993; Simmonds et al. 2008). These storms tend to migrate northeastward; accordingly, the highest winter cyclone frequency in the Arctic Ocean is observed in the Barents and Kara Seas (R and K in Figure 1.1). Relatively few storms find their way into the CAO. Summer stands in sharp contrast. Cyclone frequency in the North Atlantic and Barents Sea is significantly lower (Whittaker and Horn 1984; Serreze et al. 1993; Simmonds et al. 2008), and the cyclones that do form tend to be less intense than their winter counterparts (Serreze et al. 1993; Serreze 1995; Simmonds and Rudeva 2014).
Some summer cyclones develop locally within the Arctic Ocean (Serreze and Barrett 2008). Baroclinicity along the sea ice margin (Inoue and Hori 2011) and (later in the year) vertical enthalpy fluxes from the open water surface (Ledrew 1984) may both contribute to local cyclone development. However, it appears that many cyclones found over the Arctic Ocean in summer are generated over Siberia, especially in the Kolyma Lowland (KL in Figure 1.1), which lies in the lee of the Verkhoyansk and Gydan Ranges (SM) (Serreze 1995; Serreze et al. 2001; Sorteberg and Walsh 2008; Whittaker and Horn 1984). Hereafter, these ranges will be referred to collectively as the “Siberian mountains”.

Figure 1.1. Location map of Arctic region. The red buffer around the coastline denotes the AFZ. The bold gray line bounds the CAO [following the definition of Welsh et al. (1986)], the surrounding letters denote the coastal Arctic seas [Beaufort (B), Chukchi (C), East Siberian (E), Laptev (L), Kara (K), and Barents (R)], and the dark blue shading indicates the study area (CAO+BCEL). The letter pairs denote specific topographic features that are mentioned in the text [Mackenzie Range (MR); Kolyma Mountains (KM); Kolyma lowland (KL); Verkhoyansk and Gydan Ranges (SM); and Putorana Plateau (PP)].
Another regional mechanism that has been linked to summer cyclone development is the summer Arctic Frontal Zone (AFZ), a band of strong horizontal temperature gradients (Serreze et al. 2011; Crawford and Serreze 2015) and frequent near-surface weather fronts (Reed and Kunkel 1960; Serreze et al. 2001) that stretches along the Arctic coastlines of Eurasia and western North America from 41°E eastward to 234°E (126°W). The AFZ develops each spring and summer in response to differential heating of the atmosphere on either side of the coastline. Land, after losing its snow cover, attains higher skin temperatures than the combined sea ice/ocean surface to the north, and these thermal differences are transferred to the atmosphere via differences in the radiative and turbulent heat fluxes on either side of the coastline (Crawford and Serreze 2015). Prior studies have suggested that baroclinic instability associated with the summer AFZ contributes to cyclogenesis near the Arctic coastline (Reed and Kunkel 1960; Serreze et al. 2001; Serreze and Barrett 2008) and therefore influences both cyclone activity and precipitation over the Arctic Ocean (Serreze et al. 2001; Serreze and Barrett 2008). Although based on the common baroclinic instability model of extra-tropical cyclone development (Eady 1949; Farrell 1985; Pierrehumbert and Swanson 1995), this idea has not yet been rigorously tested. A more detailed review of AFZ literature is provided in Chapter 3.

CAO cyclone frequency exhibits substantial variability from year to year (Serreze and Barrett 2008), and on multi-year scales (Zhang et al. 2004). This has been linked to the phase of the Northern Annular Mode (NAM). Summer cyclone frequency and intensity in the CAO tend to be greater when the NAM is in its positive phase (Serreze and Barrett 2008; Simmonds et al. 2008; Zhang et al. 2004). Because it also exhibits substantial year-to-year variation in strength (Ogi et al. 2004; Crawford and Serreze 2015), the AFZ is another potential influence on cyclone’s year-to-year variability.

Expanding on this current knowledge, the following study addresses three questions:

1) In what ways, if any, does the AFZ influence summer Arctic cyclone activity?
2) How does the AFZ respond to a global warming scenario?
3) How does the AFZ-cyclone relationship respond to a global warming scenario?

To address these questions, a new cyclone detection and tracking algorithm is developed building on work by Serreze (1995), Wernli and Schwierz (2006), Serreze and Barrett (2008), Hanley and Caballero (2012), and others. This algorithm offers increased sophistication when dealing with multi-centered cyclones (MCCs) and interactions
between storms. A review of prior algorithms is provided in Chapter 4, and the new algorithm is described in Chapter 5 (along with other data and methods). A rigorous assessment of how variability in the strength of the AFZ relates to the development of cyclones affecting the CAO is conducted by applying this algorithm to data from the Modern Era Retrospective Analysis for Research and Applications (MERRA) for the period 1979-2014 (Chapter 6).

In addition to research regarding the current Arctic cyclone regime, a growing effort exists to understand how global warming has and will impact Arctic cyclone activity. Several studies suggest that Arctic frequency and/or intensity should increase in a warming scenario (Schuenemann and Cassano 2010; Bengtsson et al. 2011; Jaiser et al. 2012). However, clear trends do not exist for either cyclone frequency or intensity in the Arctic Ocean since 1979 (McCabe et al. 2001; Serreze and Barrett 2008; Simmonds et al. 2008), nor even for the 21st century (Collins et al. 2013; Vavrus 2013; Akperov et al. 2015).

The next two chapters of this study build on these efforts. Chapter 7 addresses the second research question by examining how the AFZ changes in the Community Earth System Model Large Ensemble (CEMS-LE) under representative concentration pathway (RCP) 8.5, which is a high greenhouse gas emissions scenario. CESM-LE is then used to consider the final research question in Chapter 8, which evaluates how summer Arctic cyclone activity responds to warming in RCP8.5. Emphasis is placed on the degree to which this response can be linked to the concurrent response of the AFZ. Finally, Chapter 9 provides a brief summary of results and conclusions.
CHAPTER 2. OUR DEVELOPING VIEW OF CYCLONES IN THE ARCTIC

2.1. Dispelling the High Pressure Myth

Figure 2.1 shows a typical depiction of the three-cell model of general atmospheric circulation that is used in modern introductory earth science textbooks (e.g., Ahrens 2012). Key features include the persistent low pressure of the intertropical convergence zone at low latitudes, the subtropical high pressure zone where many deserts lie, and the polar front with its associated mid-latitude cyclones. The North Pole is depicted as the high pressure center of a permanent anticyclone. Although many of the features in this model are useful generalizations of atmospheric circulation, the illustration of the Arctic as an area of high pressure is a vast oversimplification, adequate for winter perhaps but largely inaccurate for summer. Such depictions are rooted in a 19th century understanding of general circulation suggested by Helmholtz (1888). Since then, our ability to observe and analyze Arctic circulation has grown substantially, and the modern view of the summer Arctic is quite different from that presented in Figure 2.1. Rather than a quiescent, stable region of high pressure, the summer Arctic is often dominated by low pressure and large extratropical cyclones.

How did our view of the summer Arctic make such a drastic transition? In one word: observations. Before the late 1940s, harsh conditions and persistent sea ice made direct observations of the high Arctic difficult and infrequent (Bradley and England 1979; Jones 1987; Serreze and Barry 2014). Without observations, groups that published weather maps for the Northern Hemisphere simply assumed a high pressure cap in the Arctic (Reed and Kunkel 1960). One of the first reliable data sources was the Soviet drifting station North Pole 1, which was analyzed along with data from an icebreaker by Dzerdzevskii (1945). Dzerdzevskii realized that cyclone activity was not only possible in the Arctic, but common, and he suggested that the mean pressure state for the high Arctic was less than 1005 hPa. By contrast, maps of climatological sea level pressure (SLP) published by the US Weather Bureau the next year indicated an average pressure of 1017 hPa (Reed and Kunkel 1960).

Throughout the 1950s, more drifting stations were deployed, and the scientific view of the summer Arctic began to change. Rae (1951) and Dorsey (1951) both noted the passage of cyclones in the Arctic, eroding the notion that the Arctic was a permanent high pressure cap. Similarly, in his review of cyclones and anticyclones in
the Northern Hemisphere, Klein (1957) showed storm tracks crossing north of 80°N in July. The idea of high pressure dominating the North Pole all year was not immediately dismissed, though. For example, Petterssen et al. (1956) described the Arctic as a center of high pressure fringed by frequent low pressure centers. However, Reed and Kunkel (1960) note that the US Weather Bureau maps consistently depicted the Arctic high pressure center as less and less pronounced in subsequent maps. With the arrival of a reliable observation network, a better and more nuanced understanding of Arctic circulation was becoming possible. For a more in depth look at the history of our changing view regarding Arctic circulation, see Jones (1987).

![Diagram of general atmospheric circulation](image)

**Figure 2.1.** A typical depiction of the three-cell model of general atmospheric circulation in an introductory Meteorology textbook (Ahrens 2012).

### 2.2. Cyclone Seasonality

Since the mid-20th century, a growing body of work has consistently supported the portrayal of the Arctic put forward by Dzerdzeevskii (1945) and Reed and Kunkel (1960). The two most distinct aspects of Arctic cyclones are a) the strong seasonality in Arctic cyclone frequency and b) the CAO’s role as a cyclone graveyard, a place where mature and dying cyclones collect.
Using a broad definition of the Arctic as north of 60°N, the area of highest cyclone frequency is the Icelandic Low (Whittaker and Horn 1984; Serreze et al. 1993; 1997; Simmonds et al. 2008), an area of semi-permanent low pressure in the North Atlantic that favors the development of cyclones because of its location downstream of a mid-tropospheric stationary trough (Blackmon et al. 1984) and horizontal thermal contrasts, such as between the North Atlantic Ocean and the relatively colder Greenland Ice Sheet (Tsukernik et al. 2007; Brayshaw et al. 2009). Many of the cyclones that develop (through either cyclogenesis or cyclone deepening) in the Icelandic Low track northeastward, and high cyclone frequency is also found in the Greenland, Norwegian, Barents, and Kara Seas, especially in winter (Whittaker and Horn 1984; Serreze et al. 1993; Simmonds et al. 2008).

The only other relative maximum in winter cyclone frequency is in Baffin Bay (Serreze 1995; Simmonds et al. 2008). The Aleutian Low, the Icelandic Low’s equivalent in the North Pacific, lies south of 60°N (Whittaker and Horn 1984), and passage of cyclones into the Arctic Ocean from the Pacific is impeded by the North American continent (Ledrew 1984). The Arctic Ocean itself is fairly quiet in winter. Although some cyclones track all the way from the North Atlantic, they are more than twice as common in the Norwegian Sea (Simmonds et al. 2008) and more than four times as common in the Icelandic Low region (Serreze et al. 1993). In other words, the characterization of the Arctic as quiet high pressure fringed by cyclones, as offered by Petterssen et al. (1956), is often accurate for the winter season.

Summer, however, presents a much different cyclone regime. Weaker equator-to-pole temperature gradients in summer enforce weaker mid-tropospheric standing waves and a weaker Icelandic Low. As a result, cyclone frequency in the North Atlantic and the Greenland, Norwegian, Barents, and Kara Seas is significantly lower in summer (Whittaker and Horn 1984; Serreze et al. 1993). Cyclones that do form tend to be less intense than their winter counterparts (Serreze et al. 1993; Serreze 1995; Simmonds and Rudeva 2014). Cyclogenesis is especially common along the east coasts of continents in winter, but in summer cyclones are more likely to form in the lee of mountainous areas (Whittaker and Horn 1984). Since the Icelandic Low and North Atlantic storm track are significant sources of Arctic cyclones, it would be logical to assume that cyclones in the CAO become even less frequent in summer when these features are diminished.

The opposite is true. Both Dzerdzeevskii (1945) and Reed and Kunkel (1960) depicted the central Arctic Ocean as a relative maximum for summer cyclone activity. This view has been echoed by several studies using
cyclone detection and tracking algorithms based on either SLP (Serreze et al. 1993; Zhang et al. 2004; Serreze and Barrett 2008) or relative vorticity (Mesquita et al. 2008; Simmonds et al. 2008).

Where do these cyclones come from? One major source region for the CAO maximum in cyclone frequency is the Eurasian continent (Reed and Kunkel 1960; Serreze 1995; Serreze and Barrett 2008). In winter, this area is under the influence of the cold Siberian High and experiences little cyclogenesis (Whittaker and Horn 1984). In summer, the warmer continent permits the formation of cyclones in the lee of mountains, and in particular the lee of the Siberian mountains (Whittaker and Horn 1984; Serreze 1995; Serreze et al. 2001; Sorteberg and Walsh 2008). Curiously, despite using a similar algorithm as Sorteberg and Walsh (2008), Mesquita et al. (2008) did not find a cyclogenesis region in the lee of these mountains.

Another notable fraction of cyclones contributing to the maximum develop locally in the Arctic Ocean (Serreze and Barrett 2008). As sea ice retreats, the exposed ocean begins to absorb energy from the atmosphere. This allows latent and sensible heat fluxes between the ocean and atmosphere, which can contribute to cyclone formation. The relative importance of these fluxes (and hence open water) was examined by Ledrew (1984), who found that these vertical fluxes were only important during late summer and early fall, after the atmosphere began to cool down. In June and July, the atmosphere is still warmer than the underlying ocean, so energy fluxes are directed downward. However, when cooling sets in, the atmosphere cools off more quickly than the ocean water and sea ice eventually forms. This leads to significant vertical enthalpy flux into the atmosphere.

Another local effect is baroclinicity along the sea ice margin. Cyclones often travel along the ice margin in both the Greenland Sea (Serreze et al. 1993; Tsukernik et al. 2007) and the Arctic Ocean (Inoue and Hori 2011). These may occur in part because of the thermal contrasts between cold sea ice juxtaposed with relatively warm ocean waters. Cyclonic circulation is a key mechanism by which such contrasts are counteracted (Charney 1947; Hoskins and Valdes 1990; Wallace and Hobbs 2006). However, assigning cause and effect with the sea ice margin and cyclone tracks is difficult since cyclones are also known to disrupt sea ice (Serreze et al. 1993; Simmonds and Rudeva 2012).
2.3. The Arctic Frontal Zone

These local effects may explain some summer cyclone activity in the Arctic Ocean, but both Dzerdzeevskii (1945) and Reed and Kunkel (1960) also considered regional baroclinicity. The 7000 km Arctic coastline stretching from the Kola Peninsula, through Siberia and Alaska, and to the Canadian Arctic Archipelago is roughly zonal in orientation, with land to the south and ocean to the north. In summer, this long stretch of coastline experiences particularly strong meridional temperature gradients (Serreze et al. 2011; Crawford and Serreze 2015) and a relative maximum in the frequency of near-surface weather fronts (Tsukernik et al. 2007; Reed and Kunkel 1960; Serreze et al. 2001). Extending throughout the troposphere, this narrow zone is a relative maximum in baroclinicity poleward of 60°N in summer (Reed and Kunkel 1960; Serreze et al. 2001). Near the surface, temperature gradients are often stronger than the polar front (Crawford and Serreze 2015). A full review of AFZ literature is presented in Chapter 3.

Reed and Kunkel (1960) identified the AFZ as a possible mechanism for cyclone development in the summertime Arctic. Any source of baroclinicity, be it a sea ice margin, general circulation patterns, or coastal contrasts, is a potential ingredient for cyclone formation. The AFZ is distinguished by its large spatial scale and roughly zonal orientation, which reinforces pre-existing equator-to-pole temperature gradients (part of the general circulation emphasized by Reed and Kunkel (1960)). Also, persistence of the AFZ both horizontally (along 7000 km and over 180° of longitude) and vertically (from the surface to the tropopause) facilitates the development of a jet-like feature that lies distinctly north of the polar front. Therefore, the AFZ has potentially broader impacts on cyclone development than other sources of baroclinicity, such as the sea ice edge.

Another distinguishing feature of the AFZ is its strong seasonality, which is more extreme than the Icelandic Low region, the major source of winter Arctic cyclones. In summer, the Icelandic Low is weak, but it still contributes some cyclones to the Arctic Ocean. The AFZ, on the other hand, is completely absent in winter and contributes nothing.
2.4. The Arctic Ocean as Cyclone Graveyard

Although Arctic cyclone behavior exhibits strong seasonality, one aspect never changes: the Arctic Ocean experiences substantially more cyclolysis than cyclogenesis (Serreze et al. 1993; Serreze and Barrett 2008). In other words, many more cyclones fill and dissipate in the Arctic Ocean than form there. This fact was observed by Reed and Kunkel (1960), who described the Arctic Ocean as a region where cyclones “collect and stagnate” (p. 496). Whittaker and Horn (1984) drew several cyclone tracks curving northeastward into the Arctic Ocean, but none exiting southward. Similarly, Serreze and Barrett (2008) observed that cyclones have a tendency to spiral counterclockwise around the CAO, commonly experiencing lysis near the North Pole.

The Arctic Ocean’s role as a cyclone graveyard has several interesting implications. For one thing, many Arctic Ocean cyclones are well-developed, mature systems (Serreze and Barrett 2008). Many of them are also already in the process of filling when they enter the Arctic Ocean. Such cyclones are typically large in area and slow in speed, but they tend to be less intense than burgeoning cyclones in the North Atlantic or Norwegian Sea (Serreze et al. 1993; Zhang et al. 2004; Simmonds et al. 2008).

Looking back at early speculation about the Arctic Ocean as a quiet area of high pressure and no cyclone activity, this idea has some logical basis. General circulation patterns have a greater tendency to create cyclones around 45-60°N (e.g., the Icelandic Low region) than the CAO, especially in winter (Lambert et al. 2002; Eichler et al. 2015). However, the Arctic is not bereft of cyclonic activity because many of these sub-arctic cyclones migrate northward. In summer, the Arctic is even less likely to be quiescent, with local cyclogenesis over both the Arctic Ocean and Arctic continents more than compensating for the decrease in storm frequency from the North Atlantic storm track.

2.5. Trends and Interannual Variability in Arctic Cyclone Activity

With the Arctic currently experiencing particularly rapid climate change in response to global warming (Serreze and Barry 2011), a prominent question in Arctic research has become: Will cyclone activity in the Arctic be affected? One potential mechanism of change is quickly declining summer sea ice extent (Serreze and Stroeve 2008). Recall how Ledrew (1984) only observed local enthalpy fluxes being important to cyclone development in
late summer and autumn, when open water in the Arctic Ocean is warmer than the overlying atmosphere. If the period of open water were to lengthen, it might lead to greater energy fluxes into the atmosphere, lower static stability, and more favorable conditions for autumn cyclone development (Bengtsson et al. 2011; Vihma 2014; Jaiser et al. 2012). Another mechanism that might lead to increased cyclone activity is a poleward shift in general circulation patterns (König et al. 1993; McCabe et al. 2001; Bengtsson et al. 2009; Schuenemann and Cassano 2010; Collins et al. 2013), including the polar jet stream. In both cases, the Arctic Ocean is projected to experience more cyclone activity.

Whether these projected changes can yet be observed is uncertain. Some studies have suggested an increase in Arctic cyclone activity over the latter half of the 20th century. Serreze et al. (1993) found that from 1952 to 1989, the area north of 65°N experienced an increase in cyclone frequency in winter, spring and summer. Similarly, Sepp and Jaagus (2010) found an increase in cyclone frequency and a decrease in mean cyclone central pressure north of 68°N from 1948 to 2002. Looking solely at winter, McCabe et al. (2001) observed an increase in cyclone frequency in the 60°-90°N band, a decrease in cyclone frequency in the 30°-60° band, and an increase in cyclone intensity in both bands.

However, other studies have found no change. For instance, Serreze and Barrett (2008) found no significant trends for the period 1958-2005. Simmonds et al. (2008) found only tenuous evidence for a trend using multiple datasets and analysis periods between 1957 and 2006, noting that trends were no longer significant when the time period was limited to the satellite record (1979-2006).

The future of Arctic cyclone activity is also uncertain based on results from several recent papers. Vavrus (2013) looked in greater detail at extreme cyclone events. He did not find strong evidence for an historical increase in extreme cyclone frequency, but at least one model showed a sudden increase in extreme cyclones part way through the 21st century. Using a regional climate model to project out to the end of the century, Akperov et al. (2015) observed no significant increase in overall cyclone frequency in either winter or summer, but they did find an increase in small and weak cyclones as well as an increase in intensity. Using the Bergen climate model, Orsolini and Sorteberg (2015) found an overall reduction in summer cyclone frequency in the Northern Hemisphere, especially south of 48°N but an increase in cyclone frequency between 48°N and about 70°N. Using an ensemble of seventeen models, Nishii et al. (2015) found increased cyclone activity in the CAO by 2100 under RCP4.5.
Part of the difficulty in identifying trends with confidence in the historical period is that reliable and consistent data do not exist prior to 1979 (Simmonds et al. 2008). Even atmospheric reanalyses, which have a consistent model and assimilation system, may experience artificial jumps in some output variables because of changing input data (Hodges et al. 2003; Sterl 2004; Ulbrich et al. 2009). This casts doubt on any trends calculated using data that goes back to the 1940s or 1950s.

Another difficulty is that if records span only a few decades, it may be difficult to distinguish trends from multi-year internal variability (McCabe et al. 2001; Simmonds et al. 2008). For Arctic cyclone activity, year-to-year variability is substantial (Serreze and Barrett 2008), and several multi-year cycles in activity are also expressed (Zhang et al. 2004). The main driver of multi-year internal variability is the Arctic Oscillation (AO). Several papers have observed greater cyclone activity when the AO is positive (Zhang et al. 2004; Serreze and Barrett 2008; Simmonds et al. 2008). Notably, the AO was more often positive in the latter part of the 20th century, so any trends observed from the 1950s to 2000s may simply be related to the behavior of the AO (McCabe et al. 2001). To complicate matters further, the AO can be defined as the leading empirical orthogonal function (EOF) for all calendar months (Thompson and Wallace 1998) or individually for each month (Ogi et al. 2004). The two methods differ in summer, which can explain some of the difference in correlations calculated by Serreze and Barrett (2008) versus Simmonds et al. (2008).

2.6. Unanswered Questions

Today, the basic characteristics of Arctic cyclone activity are well established. We know the seasonal cycles of cyclone frequency and intensity, and the shifting patterns of source regions for Arctic cyclones. However, the contribution of the AFZ is still a closed-door watermelon. Serreze et al. (2001) linked the AFZ to cyclogenesis occurring along the Kolyma Lowlands of Siberia, but this region also lies in the lee of the Siberian mountains. It has been suggested that variability in cyclone activity may be linked to the strength of the AFZ in addition to the phase of the AO (Serreze and Barrett 2008), but this has not been confirmed with observations. Crawford and Serreze (2015) established that near-surface AFZ strength experiences substantial year-to-year variability related in part to
variation in the rate of snow cover and sea ice retreat, cloud cover, and local wind regimes. But does this variability in the AFZ translate to variability in the frequency or intensity of cyclones?

Additionally, uncertainty remains regarding long-term changes to Arctic cyclone activity. Continually improving climate models may allow us to hone projections, but we still do not understand all the mechanisms behind internal variability, which hampers our ability to assess the validity of model output. For this reason, a better understanding of the role played by the AFZ in affecting Arctic cyclone activity is needed before the response of Arctic cyclones to Arctic warming can be fully understood.
CHAPTER 3. REVIEW OF ARCTIC FRONTAL ZONE STUDIES

3.1. The Synoptic-Scale Cyclone Framework

Historically, studies of northern high-latitude fronts have had two distinct motivations, and thus the terms “Arctic front” and “Arctic frontal zone” have been applied to two distinct but related atmospheric frameworks. The framework used here has its origin in attempts to explain why cyclone activity over the central Arctic Ocean reaches a maximum in summer. Dzerdzevskii (1945) observed the prevalence of fronts along the northern coast of Siberia and identified this “Arctic frontal zone” (hereafter AFZ) as the origin of cyclones tracking into the Arctic Ocean. Building on this research, Reed and Kunkel (1960) calculated frontal frequencies from early surface analyses and confirmed the presence of a high-latituide frontal zone broadly focused along the Arctic Ocean coastlines of Siberia and Alaska. This frontal zone was distinct from the polar frontal zone to the south. They noted the persistence of strong horizontal temperature gradients from the surface up to about 500 hPa, a sharp tropopause fold, and a jet-like feature centered at about 250 hPa.

These seasonal features were attributed to differential heating of the atmosphere over land and ocean surfaces (Figure 3.1). Several differences between these surfaces combine to make the ocean surface heat more slowly. First, the land surface, which loses its snow cover in spring, has a lower albedo than the Arctic Ocean (Brown and Robinson 2011), which maintains sea ice in some coastal regions as late as July or August (Stroeve et al. 2016). Therefore, the land surface absorbs more downwelling shortwave radiation in late spring and early summer than the ocean surface (green arrows in Figure 3.1). Second, water has a greater heat capacity than soil, rock, and vegetation (Jayalakshmy and Philip 2010), so it heats up more slowly given the same energy input. Third, much of the energy absorbed by the ocean surface in summer is used to melt sea ice instead of increasing the sensible heat content. Lastly, even when the sea ice is melted, surface warming is dampened by the distribution of energy throughout the ocean mixed layer (the top 10-30 m). Land, by contrast, lacks convective mixing1, transferring heat into the sub-surface column by conduction (black arrows in Figure 3.1). In other words, the ocean

\[1\] Assuming bioturbation and cryoturbation provide negligible “turbulent” fluxes.
surface absorbs less energy, and a smaller fraction of the energy it absorbs is used to increase its skin temperature. Consequently, the ocean surface stays relatively cool and heating of the atmosphere through upwelling longwave radiation and turbulent heat fluxes is limited (red, blue, and gold arrows). In fact, turbulent heat fluxes are often downward into the ocean-ice surface during summer. The heating contrast between air over the continent and air over the ocean creates strong horizontal temperature gradients (the AFZ) along the coast. Reed and Kunkel (1960) also suggested that the AFZ might be enhanced by coastal mountain ranges, such as the Brooks Range in Alaska, which prevent cold air from pushing inland.

![Figure 3.1. Schematic of the mean June energy balance along the Alaska-Arctic Ocean coastline based on representative grid cells in the Climate Forecast System Reanalysis (CFSR) for the period 1979-2009. Values for net shortwave (SW) and longwave (LW) radiation, the turbulent sensible and latent heat fluxes, and 1000 hPa air temperature are derived from 70°N, 154°W for Arctic tundra and 71°N, 154°W for the Arctic Ocean. To show how seasonal warming progresses over both surfaces, mean temperature is shown for both June and July. Energy fluxes into the surface are calculated as the residual of all other surface energy flux terms. Arrow size is roughly proportional to energy flux magnitude.](image)

This interpretation found strong support in the much later study of Serreze et al. (2001) based on data from the National Centers for Environmental Prediction/National Center for Atmospheric Research (NCEP/NCAR) atmospheric reanalysis. They applied a thermal frontal parameter algorithm, which is the horizontal gradient of the horizontal temperature gradient (see also Hewson (1998)) to 850 hPa temperature data, citing the strong influence
of model parameterization on the boundary layer as the reason for choosing 850 hPa rather than a lower vertical level for analysis. Using this method, Serreze et al. (2001) resolved a summertime frontal zone along the coasts of Alaska and Siberia. They also observed that the AFZ was strongest in areas with mountain ranges along the coast exceeding 1000 m elevation (Alaska and eastern Siberia).

Examining output from a cyclone detection algorithm, both Serreze et al. (2001) and Serreze and Barrett (2008) found that many of the cyclones that impact the Arctic Ocean originate in Siberia and attributed this cyclogenesis to the AFZ. Serreze et al. (2001) also noted that regions where the summer AFZ is strongest (eastern Siberia and Alaska) experience an especially high proportion of their annual total precipitation in summer. This suggests that the AFZ might play an important role in the Arctic hydrologic cycle.

### 3.2. The Arctic Air Mass Framework

The terms “Arctic front” and “Arctic frontal zone” are also used to describe the southern boundary of arctic air masses. (Arctic air masses originate over the Arctic Ocean and lie to the north of polar air masses.) Bryson (1966) used air mass trajectory analysis to identify the southern boundary of the Arctic air mass in North America. He located an Arctic front whose mean July location extended from the Mackenzie River Delta southeastward across Canada. This summer frontal zone was roughly co-located with the northern boundary of boreal forest.

In contrast to the frontal zone described by Dzerdzevskii (1945) and Serreze et al. (2001), this air mass boundary exists in all seasons, not just summer. In winter, Bryson (1966) observed that the mean location of the Arctic front pushed southward across most of Canada. The one exception was in the west along the Cordillera, where the frontal zone remained fixed near the Mackenzie River Delta. Whereas the summer position of Bryson’s AFZ was co-located with the northern boundary of the boreal forest, the winter position was co-located with the southern boundary. Noting this, Bryson (1966) suggested that the boreal forest’s extent might be controlled by the position of the AFZ.

The air mass framework has been revisited sporadically for North America (Barry 1967; Willis and Grice 1977; Scott 1992; Ladd and Gajewski 2010). Inverting Bryson’s (1966) hypothesis, some have suggested that heating differences between the tundra and boreal forest may influence or determine the location of the Arctic
The logic behind this hypothesis is similar to that depicted in Figure 3.1, with the boreal forest absorbing more solar energy and contributing larger turbulent energy fluxes than the tundra. However, Beringer et al. (2001) found that the surface heat flux difference in summer between tundra and boreal forest biomes is equal to the variation found within each biome. Additionally, the strongest contrast between the two biomes was observed in May, whereas the AFZ is not distinct until June. Comparisons between the mean July position of the Arctic front based on trajectory analysis and anomalies of the normalized difference vegetation index (NDVI) further question a vegetative control on the Arctic front; they instead suggest that variability in Arctic front position influences primary productivity (Ladd and Gajewski 2010).

One recent study has presented a perspective that combines these two frameworks. Liess et al. (2011) coupled climate and land surface models to project the impact of northward advance of the boreal forest on the AFZ. While they found the strongest frontal expression along the coastline, they also found that the temperature gradient and jet stream intensified slightly with northward advance of the tree line. They concluded that the land-ocean contrast may be primary to AFZ development, but vegetation contrasts can also enhance AFZ strength.

To summarize, in the framework motivated by cyclone behavior, the AFZ is defined as an essentially fixed geographic area of seasonal baroclinicity. In the framework motivated by air mass description, the AFZ has a constant function but no fixed location. An “Arctic front” always exists because it is defined as the boundary between arctic and polar air masses. This boundary will shift farther south in winter and farther north in summer. The regions of focus also differ for the two frameworks. The cyclone-motivated framework is usually applied to the Arctic Ocean basin whereas the air mass-motivated framework is more often associated with North America, and especially Canada. Hypotheses concerning the impact of Arctic fronts on biome boundaries and vice versa have been usually applied to the air mass framework; however, some studies (Lynch et al. 2001; Liess et al. 2011) have considered the role of vegetation in either.
3.3. A New Look at the Arctic Frontal Zone

Crawford and Serreze (2015) re-examined the basic structure of the AFZ using maps and cross sections of temperature, wind, and the four main terms of the surface-atmosphere energy balance (net shortwave and longwave radiation and the sensible and latent heat fluxes) from three atmospheric reanalyses: the National Aeronautics and Space Administration (NASA) Modern Era Retrospective-Analysis for Research and Applications (MERRA; Rienecker et al. 2011), the National Oceanic and Atmospheric Administration (NOAA) Climate Forecast System Reanalysis (CFSR; Saha et al. 2010); and the European Centre for Medium Range Weather Forecasting (ECMWF) Interim European ReAnalysis (ERA; Dee et al. 2011).

Figure 3.2. Mean July 2 m temperature gradient magnitude for 1979-2012 from a) ERA, b) MERRA, and c) CFSR, with the treeline (from the Circumpolar Arctic Vegetation Map) marked as a solid blue line and the mean July 15 sea ice edge (1979-2012; from the combined passive microwave record) marked as a dashed black line.
Figure 3.3. Monthly climatology (1979-2012) of surface energy fluxes from MERRA. Cross sections along 154°W in (a) July and (b) January and maps of (c) net longwave radiation, (d) net shortwave radiation, (e) latent heat, and (f) sensible heat fluxes in July are included. On maps, 154°W is marked with a dashed line. The Alaskan coast (C; black), and treeline (T; white) are also marked. Black curve in cross sections is net allwave radiation; other colors correspond to maps. Radiation fluxes are positive downwards and turbulent fluxes are positive upwards.
Strong near-surface (2 m level) temperature gradients were observed along the coastline of the Arctic Ocean in June through August in all three reanalyses (results for July are shown in Figure 3.2). These gradients were associated with sharp contrasts in all four terms of the surface energy balance (Figure 3.3). The net shortwave radiation flux was about half as strong for the Arctic Ocean in July than for Siberia or Alaska, a reflection of the differences in albedo. Likewise, the continents exhibited a substantially greater longwave radiation flux and greater turbulent sensible and latent heat fluxes than the Arctic Ocean surface. No such abrupt changes in energy fluxes or temperature were observed along the tundra boundary (or any other location north of 60°N). These findings, consistent in all three reanalyses, support the framework pioneered by Dzerdzeevskii (1945) and Reed and Kunkel (1960) and furthered by Serreze et al. (2001). For that reason, and because of the current interest in cyclone activity, the remainder of this study adopts the framework associated with synoptic-scale cyclones.

Crawford and Serreze (2015) also confirmed the extension of strong, coastally fixed horizontal temperature gradients throughout the troposphere (Figure 3.4). Although most prominent near the surface, these temperature gradients are strong enough throughout a deep enough layer of the troposphere to encourage the development of a jet-like feature near the tropopause that is distinct from and north of the polar jet of the mid latitudes. This distinct frontal zone and jet-like feature exist only in summer, when incoming shortwave radiation is sufficient to create a large energy imbalance across the coastline. The jet-like feature is also only apparent along the 1700 km stretch of coastline from about the Kola Peninsula in Russia eastward to Banks Island in Canada, where the land-sea contrast is distinct and abrupt and the coastline has a generally zonal orientation. In places without a clear land-sea transition, such as the highly variable surfaces of the Canadian Arctic Archipelago, no deep frontal zone or jet-like feature develops (despite vegetation boundaries existing at these longitudes). It was also noted that the zonal orientation of the Arctic Ocean coastline means that the warm continent and cold ocean create temperature gradients that are primarily meridional, reinforcing the general equator-to-pole gradient of warm to the south and cold to the north. This reinforcing character being persistent across such a long arc (1700

\[ \text{\textsuperscript{2}} \text{ See the thermal wind equation (Equation 3.1) in the following section for a full explanation of why this occurs.} \]
km and 190° of longitude) likely enhances the ability of these surface contrasts to translate into a deep frontal zone that supports a jet-like feature aloft.

The last set of findings from Crawford and Serreze (2015) relevant to the current study is their examination of spatiotemporal variability in the near-surface expression of the AFZ. Despite having a reliable seasonal cycle of development in spring and summer and disappearance in fall and winter, the relative strength of the AFZ (as measured by the magnitude of the horizontal temperature gradient) from one July to the next varies substantially (e.g., from 4 to 8 K (100 km)\(^{-1}\) at the 2 m level along the coast of the eastern Kolyma Lowlands). This interannual variability is also spatially heterogeneous. The entire 190° of longitude does not experience above average AFZ strength at the same time; rather, anomalies in AFZ strength tend to occur regionally. For instance, in July 1997, the AFZ was stronger than normal along the Laptev Sea but weaker than normal along the Kara Sea.
Figure 3.4. Latitudinal cross sections of mean July meridional temperature gradient (a-c) and zonal wind velocity (d-f), averaged for the period 1979-2012 (from MERRA). Cross sections are along longitudes 120°E (a and d), 80°W (b and e), and 154°W (c and f). Also marked are the Arctic Ocean coastline (C), the treeline (T), the northern shore of Ellesmere Island (E), and the southern coast of James Bay (J).
Using the 34-year record of July 2 m temperature gradient anomalies, Crawford and Serreze (2015) performed a clustering analysis to determine whether the regionalization apparent in maps of temperature gradient anomalies was supported by statistical criteria. This clustering analysis grouped all locations along the Arctic Ocean coastline so that the temperature gradient anomalies of locations belonging to the same group were more similar than to locations belonging to different groups. In other words, the temperature gradient magnitudes of locations belonging to the same group tend to co-vary, whereas no (or only weak) correlation exists for locations belonging to different groups. No spatial information was included in the statistical analysis. However, spatial autocorrelation throughout the AFZ is strong, and all resulting sectors were characterized by contiguous locations. The resulting breaks in AFZ sectors were located at physically meaningful boundaries, such as the Bering Strait (between Sectors 7 and 8), the Verkhoyansk Range (between Sectors 4 and 5) and Khatanga Bay (between Sectors 3 and 4).

Finally, Crawford and Serreze (2015) used multiple linear regression models to determine which factors affect the relative strength of each sector of the AFZ each July. A spatial error term was included to account for spatial autocorrelation amongst AFZ sectors, but no serial autocorrelation was present (meaning anomalies to AFZ strength in one year have no influence on the anomalies in the following year). Results showed four significant factors: the timing of snow cover retreat from near-shore land and sea ice retreat from the near-shore waters, cloud cover, and the across-shore wind direction. Later retreat of snow cover leads to less absorption of solar radiation by the land surface, lower temperatures over land, and a weaker AFZ. Similarly, later sea ice retreat leads to less absorption of solar radiation by the water column and lower temperatures over the ocean. This results in a stronger AFZ since normal conditions are warmer air over land and colder air over the ocean. Cloud cover reduces the amount of incoming shortwave radiation and increases the amount of incoming longwave radiation. In July, the reduction in shortwave radiation is dominant, so increased cloud cover leads to cooling, but this effect is only significant for land. For this reason, cloudier conditions lead to a weaker AFZ. Finally, offshore wind anomalies mean less cold air advection and/or more warm air advection. Again, the effect is much greater on temperatures over land, which are more readily changed than temperatures over the ocean, so offshore anomalies lead to a stronger AFZ.
This study expands on the work of Crawford and Serreze (2015) by a) assessing the relationship between the AFZ and summer cyclone activity (Chapter 6) and b) comparing the behavior of the AFZ and its relationship with summer cyclone activity in two periods (1990-2005 and 2071-2080) under a global warming scenario (Chapters 7 and 8). In doing so, this study advances our understanding of the role the AFZ plays in the Arctic climate system and how that role might respond to global warming during the current century.

3.4. Baroclinic Instability: Linking the AFZ to Synoptic-Scale Cyclones

Since much of this study focuses on the relationship between the AFZ and summer Arctic cyclone activity, it is necessary to establish a firm physical basis for why such a relationship might exist. The reasons for suspecting a connection between the AFZ and cyclone development are rooted in common baroclinic instability models of extra-tropical cyclone development (Charney 1947; Eady 1949; Farrell 1985). Briefly, these models envision extra-tropical cyclones as the mechanism by which the potential energy stored in the equator to pole temperature gradient is released (Pierrehumbert and Swanson 1995). Notably, the models of Charney and Eady did not introduce the most basic concepts behind the development of extra-tropical cyclones, such as the understanding that extra-tropical cyclones develop along frontal boundaries between colder air toward the pole and warmer air to the south, that cyclonic circulation counteracts this frontal boundary by pulling warm air poleward and cold air equatorward, that this process involves the conversion of potential energy into kinetic energy, or that cyclones typically dissipate once the lifting of warm air ceases and the source of potential energy is cut off. Basic descriptions of this life cycle were established by the Norwegian School in the early 1900s (e.g., Bjerknes and Solberg 1922). The real contribution of Charney (1947) and Eady (1949) was providing a mathematical explanation for how such development could spontaneously occur; this mathematical explanation substantially improved our ability to accurately represent synoptic-scale cyclones in atmospheric models.

The inclusion of continuous vertical wind shear in their models was a key ingredient for both Charney and Eady. Simply put, vertical wind shear is a change in speed and/or direction of wind over a particular location as altitude increases. An important cause of vertical wind shear is described by the thermal wind equation,
\[(V_g)_2 - (V_g)_1 = \left(\frac{R}{f} \ln \frac{p_1}{p_2}\right) k \times \nabla (\bar{T})\]

(Equation 3.1)

where \((V_g)_1\) and \((V_g)_2\) are the geostrophic \(^3\) wind at pressure levels \(p_1\) and \(p_2\), respectively, \(R\) is the gas constant (287 J K\(^{-1}\) Kg\(^{-1}\)), \(f\) is the Coriolis frequency \((f = 2\Omega \sin\phi, \text{where } \Omega \text{ is Earth’s rotation rate and } \phi \text{ is latitude})\), and \(T\) is temperature (averaged for the levels between \(p_1\) and \(p_2\)). This equation helps show how on a large geostrophic scale, a meridional temperature gradient on a rapidly rotating planet facilitates the development of a strong, narrow jet stream aloft. More generally, it also shows how stronger horizontal temperature gradients lead to greater vertical wind shear.

The presence of vertical wind shear is an essential aspect of a baroclinic atmosphere. In a barotropic atmosphere, any difference in pressure at two locations is determined purely by a difference in mass. Higher pressure at a location means a greater mass of air is above it. In a baroclinic atmosphere, pressure differences can also arise from a difference in density, and therefore temperature. \(^4\) The following explanation of baroclinicity is strongly influenced by Section 7.2.7 in Wallace and Hobbs (2006) and their Figures 7.13 and 7.14.

To better understand the meaning of Equation 3.1, consider first an equivalent barotropic case (Figure 3.5), where the temperature (and thickness) field is aligned with the pressure field (and geopotential height (GPH) field) such that isotherms (and isopachs) are parallel to isobars (and contours). More specifically, consider first a situation in which the mass of the atmosphere is constant but its temperature varies horizontally (Figure 3.5a). Since the atmosphere’s mass is evenly distributed, the pressure at sea level \(p_1\) is also constant, and since there is no pressure gradient, the geostrophic wind has a velocity of \(0\) ms\(^{-1}\). However, because the atmosphere is cooler on the left side of Figure 3.5a and warmer on the right, it also is denser on the left side and less dense on the right. A balloon rising 1000 m through the cooler, denser atmosphere would rise above a greater percentage of the

\(^3\) “Geostrophic” means resulting from a balance between horizontal pressure gradient forces and the apparent horizontal Coriolis force.

\(^4\) Kept at constant pressure, warmer air occupies a greater volume, as seen by the ideal gas law \(pV = nRT\), where \(p\) is pressure, \(V\) is volume, \(n\) is the amount of gas (in moles), \(T\) is temperature, and \(R\) is the ideal gas constant. (A greater volume means a lower density.)
atmosphere’s mass than a balloon rising the same distance in the warmer, less dense atmosphere. Therefore, the balloon on the cooler, denser left side would be at a lower pressure. Considered another way, the balloon rising through the warmer atmosphere on the right would have to travel farther before reaching the same pressure level ($p_2$). The temperature difference leads to a pressure gradient, which can be observed as a slope from right to left along the $p_2$ surface. This slope induces a geostrophic wind (blowing with the warm air and high pressure to the right) even without any horizontal variation in total mass. Equation 3.1 shows that the stronger the temperature gradient, the more drastic the resulting pressure gradient, and the stronger the geostrophic wind ($V_g$) aloft.

**Figure 3.5.** Three examples of the change of the geostrophic wind with altitude in three equivalent barotropic atmospheres of the Northern Hemisphere. Colored surfaces represent planes of constant pressure. Orange indicates warm (low density, high thickness between pressure levels), and blue indicates cold (high density, low thickness). The white wedges in (b) and (c) represent the difference in total atmospheric mass. (Inspired by Figure 7.13 in Wallace and Hobbs (2006).)
Thermal wind shear may also influence winds in concert with mass differences. The lower two examples in Figure 3.5 show cases in which the pressure at sea level varies horizontally. In Figure 3.5b, the cooler column of air has less mass, so there it has lower surface pressure than the warmer column. As indicated by the slope of the $p_1$ surface, a surface pressure gradient induces a geostrophic wind. Since the air above the surface high is warmer and less dense than the air above the surface low, the slope of each successive pressure surface is a little steeper with increasing altitude, strengthening the geostrophic wind. In other words, a temperature difference leads to a density difference that reinforces the mass-driven pressure gradient and, by extension, intensifies the geostrophic wind.

The example in Figure 3.5c is similar to that in Figure 3.5b except now the temperature difference encourages a pressure gradient that is contrary to the mass-driven pressure gradient. In this case, the warm air column has less mass, so the pressure gradient is directed from cold to warm. The geostrophic wind near the surface ($V_{g1}$) is blowing with warm air to the left. However, even though the cold air has more mass, a balloon rising through it would still reach pressure level $p_2$ at a lower altitude than a balloon rising through the warm air because the cold air is denser. So despite the total mass difference, pressure level $p_2$ is tilted from warm to cold, and the geostrophic wind ($V_{g2}$) reverses direction, blowing with the warm air to the right at level $p_2$. Such reversal of wind direction is commonly observed at subtropical latitudes, where easterly trade winds may blow near the surface even when a strong westerly jet stream is present near the tropopause (as shown in Figure 3.4e-f).

Equivalent barotropic cases are useful for explaining the influence of thermal wind shear, but a fully baroclinic atmosphere has additional complexity. Just as the geostrophic wind follows along isobars (which relate to mass), the thermal wind shear follows along isotherms (which relate to temperature and density). In the equivalent barotropic cases described above, the isobars and isotherms were parallel to each other. The thermal wind shear never made the wind more zonal or meridional; it only influenced wind speed. In a fully baroclinic atmosphere, however, the isobars and isotherms intersect each other.

This has two important implications. First, the thermal wind shear affects both speed and direction. The process can be imagined as the thermal wind shear pulling on the geostrophic wind so that it better aligns with the isotherms. Such rotation of wind with altitude is called backing when cyclonic and veering when anticyclonic. Second, wind blowing across isotherms also means wind blowing from areas of colder air to warmer air (cold
advection) or from warmer air to colder air (warm advection). Cold advection is associated with backing, while warm advection is associated with veering.\(^5\)

One way in which baroclinic waves intensify is through a positive feedback on this advection. Advection causes a deformation of the temperature field, such as when poleward warm air advection pushes isotherms poleward. In doing so, advection can cause a previously meridional temperature gradient to adopt a zonal component. If the winds are primarily zonal, this causes a greater misalignment of isobars and isotherms than previously existed (i.e., the wind is now closer to blowing perpendicular to the isotherms). Greater misalignment means a more baroclinic state and more potential for advection.

The reason why this process is unstable is that the baroclinicity can spontaneously amplify in a positive feedback that continually taps into the potential energy stored by geostrophic flow (Charney 1947; Pierrehumbert and Swanson 1995). Notably, one of the main reasons Charney (1947) gives for investigating baroclinic instability is that barotropic models (and equivalent barotropic models) of the atmosphere showed no mechanism for spontaneous conversion of potential energy built up by geostrophic flow to kinetic energy. Only in the presence of substantial wind shear (instigated by substantial horizontal temperature gradients) do disturbances in the general flow amplify of their own accord.

This amplification of baroclinic waves also includes the development of circulation at the surface and tropopause, which further facilitates an acceleration of the conversion of potential energy into kinetic energy. According to the calculations of Charney and Eady, circulations that develop as a result of baroclinic instability typically have a spatial scale on the order of 1000 km (the synoptic scale) and alternate in the \(x\) direction between cyclonic and anti-cyclonic circulation. This scale and pattern matches the observed pattern of alternating synoptic-scale cyclonic and anticyclonic circulation associated with the polar fronts.

Baroclinic instability causes break-down into eddies at a synoptic scale because spontaneous amplification of baroclinic waves only occurs if the wind shear exceeds a certain threshold, and this threshold is dependent on wavelength (Charney 1947). However, the troposphere of the mid latitudes is always generally baroclinic (Eady 1949), so some degree of wind shear always exists. According to Charney (1947), a level of 1.5 m s\(^{-1}\) km\(^{-1}\) is usually

\(^5\) See Figure 7.14 in Wallace and Hobbs (2006) for an illustration.
exceeded throughout the mid latitudes in winter, equating to a length scale of about 6000 km. The spatial scale for spontaneously amplifying baroclinic waves can also be estimated by the internal Rossby radius of deformation, which is defined\(^6\) as

\[
L_{R,n} = \frac{NH}{f}
\]

(Equation 3.2)

where \(N\) is a measure of static stability called the Brunt-Väisälä frequency (where \(N^2 = \frac{g}{\theta}(\partial \theta/\partial z)\), \(g\) is acceleration due to gravity, \(\theta\) is potential temperature, and \(z\) is the vertical coordinate), \(f\) is the Coriolis frequency, and \(H\) is the scale height. Since \(N/f\) is typically on the order of 100 and the tropopause lies at a height of about 10 km, the internal Rossby radius is around 1000 km. In either case, the resulting numbers match the typically observed synoptic scale. At this scale, rotational effects become as important as buoyancy or gravity waves (Gill 1982), and the baroclinic waves induce synoptic scale circulation.

Also notable is that, independent of scale, baroclinic instability is more likely whenever the wind shear is stronger. Looking back at Equation 3.1, this also means that baroclinic instability is more likely whenever the horizontal temperature gradients in the troposphere are stronger. This relationship between the strength of vertical wind shear and baroclinic instability is summarized by the Eady growth rate (EGR), defined more formally as

\[
EGR = 0.3098 \frac{|f| \frac{\partial V(z)}{\partial z}}{N}
\]

(Equation 3.3)

where \(f\) is the Coriolis parameter, \(V(z)\) is the vertical profile of the horizontal wind, \(z\) is the vertical coordinate, and \(N\) is the Brunt-Väisälä frequency. This equation is sometimes referred to as the maximum Eady growth rate equation because it describes the potential for the growth of baroclinic waves as opposed to realized growth.

Stepping back for a moment, it is useful to place both synoptic-scale cyclones and the AFZ in a larger context. At their core, both phenomena are related to a) the unequal heating of the Earth’s surface by incoming solar radiation and b) the tendency for the Earth’s atmosphere to work towards equilibrating the resulting energy imbalance. At the top of the atmosphere, the decline in annual insolation from equator to pole is quite gradual and smooth. In the absence of any other factors, one might expect energy transfer from equator to pole to also be smooth, with warm air flowing poleward and cool air flowing equatorward in a simple convection cycle.

However, the Earth is not nearly so simple. For one thing, its rapid rotation prevents the development a single simple convection cell. Convection is only dominant in the two Hadley Cells, which are confined to within roughly ±30° latitude. Moreover, balance between the pressure gradient force and apparent Coriolis force (geostrophic balance) favors the development of distinct jet streams that demarcate narrow zones of strong temperature gradients dividing somewhat homogeneous masses of air. This geostrophic balance also allows the accumulation of potential energy since it resists a smooth and simple poleward transfer of energy.

As a result, more punctuated and violent transfer of energy from equator to pole becomes focused along these broad frontal zones. Vertical wind shear induced by strong horizontal temperature gradients leads to instability in the standing waves of geostrophic flow. Instability leads to the breakdown of waves into synoptic-scale eddies, and these eddies perform most of the meridional transfer of atmospheric mass and energy in the mid and high latitudes. Through both horizontal and vertical motion, these eddies work to destroy the temperature gradients and convert the potential energy that has been built up in the system. Theoretically, once the baroclinicity is equalized and the pool of potential energy is consumed, the synoptic-scale eddies dissipate, permitting the reformation of the baroclinicity in a continuing cycle. In practice, of course, synoptic-scale cyclones exhibit more variability and messiness than suggested by such models, but the work of Charney (1947) and Eady (1949) is still the basis for more recent advances in baroclinic instability theory (Farrell 1985; Hoskins and Valdes 1990; Thornicroft et al. 1993).

Baroclinicity can be thought of as an intermediate step between the development of an energy imbalance by solar heating and subsequent equilibration by synoptic-scale eddies. It can also be considered as fuel for synoptic-scale cyclones. Zones of baroclinicity will develop spontaneously in general circulation models run for an
aquaplanet\(^7\) (Wallace and Hobbs 2006; Brayshaw et al. 2008). However, the location and strength of baroclinic zones can also be influenced by the disruption of large-scale flow by high topography and by surface contrasts, such as land-ocean contrasts (Hoskins and Valdes 1990; Brayshaw et al. 2009), ice-ocean contrasts (Tsukernik et al. 2007), and even ocean currents that separate thermally distinct water masses (Brayshaw et al. 2011). The thermal properties of Earth's various surfaces strongly impact how incident radiation is translated into skin temperatures, and if these skin temperatures are translated upward throughout the troposphere, they can help focus baroclinicity and facilitate the development of synoptic-scale cyclones.

The AFZ is a baroclinic zone that develops each summer because of a strong contrast between heating of the combined ocean-sea ice surface of the Arctic Ocean and largely snow-free land. It is persistent enough to develop a jet-like feature aloft that is north of and distinct from the polar jet (Reed and Kunkel 1960; Serreze et al. 2001; Crawford and Serreze 2015). Comparing winter to summer, the central Arctic Ocean experiences more cyclone activity when the AFZ is present (Serreze and Barrett 2008; Simmonds et al. 2008). In the context of the baroclinic instability model described above, these lines of evidence suggest that the AFZ likely plays an active role in the development of summer Arctic cyclones. The focus of this study is to critically evaluate the proposed AFZ-cyclone relationship both in the observational record and into the future.

\(^7\) An aquaplanet has a surface composed entirely of water.
CHAPTER 4. CYCLONE DETECTION AND TRACKING METHODS

4.1. Introduction

4.1.1. Why Study Extra-Tropical Cyclones?

Cyclones are a crucial component of Earth’s climate system. They are a key mechanism by which the atmosphere redistributes energy and moisture across the Earth’s surface (Wallace and Hobbs 2006). Moreover, they demonstrate the intimate links between large-scale atmospheric circulation and local weather patterns. Extra-tropical cyclones often form as the eddies that swirl off the meandering polar jet stream (Bjerknes and Solberg 1922; Wallace and Hobbs 2006). They are the turbulence and the chaos in Earth’s large-scale circulation. A large amount of the daily deviation from climatological means experienced at extra-tropical locations is induced by the influence of such eddies. The study of extra-tropical cyclones is thus essential to the field of climatology, for which a major focus is the relationship between large-scale circulation and local and regional climate (Barry and Perry 1973).

In addition to their role in the broader Earth system, cyclones are also integral to humans’ daily activities. The passage of weather systems is accompanied by changes in clouds, precipitation, winds, and temperature that impact the weather experienced by humans on a daily basis. When strong enough to earn the label “storm”, cyclones can severely impact human activities and infrastructure. Heavy snowfall can stifle traffic, delay flights, and cancel school days and events. High winds can topple trees and contribute to storm surges on the coast. Heavy rain may lead to flooding and substantial structural damage. Other hazards accompanying some storms, such as lightning, hail, and tornadoes, cause significant damage to human property and human lives. Even in the Arctic, with relatively few permanent settlements, cyclones have increasingly relevant as human activities such as shipping, resource extraction, and tourism, have become more prevalent (ACIA 2005; Dodds 2010).

Because of cyclones’ strong relevance to human activities, the ability to identify them in fields of atmospheric data, track their movement, and describe their characteristics is a valuable tool. A database of cyclones and their tracks can be used to better understand the climatology of cyclone behavior. It can also be used to project how the frequency and intensity of such weather hazards might change in the future. As discussed
below, researchers have been tracking cyclones for over a century, but the practice has experienced accelerated advancements over the past 25 years. The following review focuses on the development of cyclone tracking methods. Climatological knowledge of extratropical systems, which comes in part from the application of these methods, was discussed in Chapter 2.

4.1.2. Basic Characteristics of Cyclone Tracking

Most analyses of cyclone tracks include three basic steps: detection, tracking, and summarization. Cyclones are detected either by their centers or their areas at a particular snapshot in time. If cyclones are observed at two different times, systems that correspond to each other can be linked to begin forming a track. Repeating this detection and linking process over many time intervals yields tracks for many cyclones as they develop, mature, and eventually dissipate.

To satisfy the motivations described above, the final step is to calculate characteristics and statistics that summarize cyclone behavior for the place and period of interest. For instance, how many cyclones tracked through the North Atlantic Ocean in winter for the period 1980-1989? From where do cyclones impacting southern Alaska typically originate? Does cyclone frequency or intensity in the Southern Hemisphere exhibit any long-term trends?

The three-step process of detection, tracking, and summarization provides a simple framework for conceptualizing how to track cyclones. However, each of those three steps can be achieved in many ways. Since the advent of automated methods for tracking cyclones in the 1990s, a diversity of cyclone tracking methods has been developed by researchers across the world. These methods will be described in detail in Section 4.3, but first it is worthwhile to consider the historical context of cyclone tracking.

4.2. 1830s-1980s: Manual Tracking

4.2.1. Cyclone Tracking History

Interest in tracking storms was a major motivation behind the development of synoptic climatology in the mid-1800s. Early weather charts designed for mariners were used to trace the paths of hurricanes in the North Atlantic and the Bay of Bengal-Arabian Sea by Redfield (1831) and Piddington (1848), respectively. However,
comprehensive cyclone tracking suitable for building climatologies requires more rigorous data than were available in the mid-1800s. In particular, data needs to a) be available at regular time intervals, b) cover a sufficient period to calculate a climatology, and c) cover a large enough spatial domain to capture all cyclones relevant to the area of interest. Since individual cyclones are typically on the order of a thousand kilometers in size and travel several thousands of kilometers, even a local area of interest necessitates a regional or hemispherical domain to prevent edge effects (Serreze and Barry 1988; Blender et al. 1997).

As the production of weather charts became routine in several countries during the latter half of the 1800s, the requirement of regularly reporting atmospheric data was fulfilled. Additionally, physical understanding of cyclones increased. The passing of storm systems was related to various atmospheric features, such as winds and streamlines (Galton 1863; Abercromby 1883), pressure patterns (Abercromby 1878), and (for extra-tropical systems) fronts (Bjerknes 1919). Despite these advances, the only pre-1950s research on cyclone tracks frequently cited by modern analysts was by van Bebber (1891), who studied maps from the German Naval Observatory (Deutsche Seewarte) to identify and track surface pressure minima. Nearly sixty years later, Petterssen (1950) produced a comprehensive climatology of cyclone frequency and cyclogenesis for the entire Northern Hemisphere for the period 1899-1939. Since all analysis was performed manually, this process was a lengthy undertaking, requiring close inspection of closed low pressure systems on thousands of archived weather maps from the US weather bureau. Klein (1957) performed a similar analysis separately for the periods 1909-14 and 1924-37.

Analogous studies for the Southern Hemisphere did not appear until the 1960s (Karlesky 1963; van Loon 1965; Taljaard 1967).

Although using data with wide spatial and temporal coverage, all of these early efforts suffered from the relative paucity of data at high latitudes. The lack of observations in the Arctic and Antarctic required analysts to make assumptions when constructing weather charts (Haak and Ulbrich 1996). Barry and Carleton (2001) do not consider such charts reliable for high latitudes until the 1950s, when regular observations were begun. Accordingly, results of summer cyclone frequency in the Arctic differ considerably between the studies by Petterssen (1950) and Reed and Kunkel (1960), who used US weather bureau maps for the period 1952-56 (as discussed in Chapter 3).
Improvements to meteorological data acquisition networks, including increased use of radiosondes, buoys, aircraft, and especially satellites, led to improved weather charts throughout the 1970s and 1980s (Yarnal 1993; Barry and Carleton 2001). Accordingly, dozens of cyclone tracking studies were completed during these decades. In the Southern Hemisphere, these included Streten and Troup (1973), Carleton (1979), and Le Marshall and Kelly (1981). More work was done in the Northern Hemisphere, including studies by Reitan (1974), Chung et al. (1976), Colucci (1976), Zishka and Smith (1980), Hayden (1981), Overland and Pease (1982), Whittaker and Horn (1984), and Serreze and Barry (1988), among others. Moreover, maps of cyclone tracks became a standard part of NOAA’s Climatological Data, National Summary (Zishka and Smith 1980).

Despite the widespread use of computers in the field of synoptic climatology during the 1970s (Yarnal 1993), all of the studies mentioned above used manual tracking methods. For some studies (e.g., Zishka and Smith 1980; Whittaker and Horn 1984) much of the data was still in an analog form not readable by computer. But even those using computers to store and display data (e.g., Carleton 1979) used manual methods to relate expressions of the same cyclone at different observation times.

4.2.2. A Deeper Look at Manual Tracking Methods

The details of manual tracking methods can be better understood by comparing a few of these studies in greater detail. Zishka and Smith (1980) used the cyclone tracks published in NOAA’s Climatological Data, National Summary to build a climatology of cyclone activity in January and July of 1950-77. Whittaker and Horn (1984) aggregated weather charts at both 500 hPa and sea level from several sources to construct an updated climatology of the Northern Hemisphere for the four mid-season months of 1958-77. Similarly, Serreze and Barry (1988) combined surface weather charts and buoy data to focus on synoptic activity in the Arctic Basin for January and July of 1979-85.

One of the first issues presented to all these studies was the gridding of the input data. The US Weather Bureau data is gridded by latitude and longitude. This is problematic because the area of 1° latitude by 1° longitude is larger at the equator than the poles, where lines of longitude converge. A cyclone count of 20 centers at the equator represents a lower cyclone density than 20 centers near the poles. To address this issue, Whittaker and
Horn (1984) performed their analysis on the native latitude-longitude grid and then scaled results by latitude afterwards. However, as Zishka and Smith (1980) note, the smaller grid cell size at high latitudes means that smaller scale features can be identified. This biases the results toward more high-latitude cyclones. Serreze and Barry (1988) resolve this issue by reprojecting data into an equal-area grid before analysis. Although this method does address the bias, the reprojection leads to artifacts in the data that need to be smoothed out in post-processing. This smoothing effectively coarsens the spatial resolution of the results.

Whichever grid was used, cyclone centers were identified as closed isobars (at intervals of 4 hPa) around minima of SLP for each weather chart. Cyclone tracks were constructed by examining consecutive weather charts. Each cyclone center in time 1 was numbered. By visual inspection, the researcher attempted to associate each cyclone center from time 2 with a center from time 1. If a match was found, the center in time 2 inherited the cyclone number from time 1. Otherwise, the center in time 2 received a new number and a cyclogenesis event had occurred. “Visual inspection” could mean many things, such as looking at two pieces of paper side by side or at two images on a computer screen. A particularly useful method was to print out plastic overlays and align one atop the other (Zishka and Smith 1980).

In addition to the basic tracking method, Whittaker and Horn (1984) added two other processing steps. First, they removed any cyclone center from analysis that did not persist for at least 24 hours. Second, if an occluded semi-stationary cyclone center in time 1 appeared close to another cyclone center in time 2, they considered the nearby center a new independent cyclone even if the stationary low could have tracked to it. These choices modified the basic process to make results comply with our physical understanding of cyclone lifecycles. The typical lifespan of a cyclone is 5-7 days (Barry and Carleton 2001), so a minima in SLP that lasts less than 24 hours is more likely a diurnally produced heat low or other non-synoptic feature (Serreze et al. 1997). Occluding, quasi-stationary cyclones are usually in their final stages of maturity (Wallace and Hobbs 2006), so a sudden resurgence in propagation speed is unlikely.

After the detection and tracking steps, the final summarization step also requires some decisions. Cyclone frequency is of common interest to many cyclone tracking studies, but several metrics can be used. The simplest is to, for each grid cell, count the number of times a cyclone center appears (Klein 1957). If the grid cells are not already equal area, the count can be divided by area to obtain a cyclone density. However, this method may cause
a grid cell that experiences a few quasi-stationary cyclone centers to appear to have a greater frequency than another grid cell that experiences many fast moving cyclones. If a cyclone moves more than one grid cell per time interval, it will not be counted in the grid cells it merely passes through (Whittaker and Horn 1984). This is problematic because, especially if studying short time periods, areas of high cyclone frequency may artificially appear patchy (Lambert 1988). However, if the desired result is to observe areas commonly experiencing cyclone-associated weather, then cyclone center counts may be appropriate (ibid.).

Alternatively, cyclone frequency can be measured as the number of cyclone tracks that pass through a grid cell. Even if a cyclone spends multiple observation times in the same grid cell, it will only count once. Additionally, if a cyclone moves two grid cells between one observation time and the next, the track will still contribute to the frequency in the grid cell it passed through, not just the ones it appears in at finite intervals. This is the method employed by Zishka and Smith (1980) and Whittaker and Horn (1984).

All of the issues just discussed were discovered and addressed during the manual tracking era, but they are still relevant today. As examined in the next section, the architects of automated tracking algorithms also struggle with data projection, the interplay of spatial resolution and temporal resolution, and how to measure cyclone frequency. One key difference is that the actual detection and tracking is automated by a computer, so more observation times can be analyzed much more quickly than by a human researcher.

4.3.1990s - Present: Automated Tracking

4.3.1. Historical Perspective

As implied above, two developments necessary for automated cyclone detection and tracking to develop were: 1) the automation power of computers and 2) the availability of gridded fields of atmospheric data in a digital format. However, two additional reasons for the shift to automated methods were: 3) manual tracking is time-intensive and 4) automated methods decrease subjectivity and increase consistency in detection and tracking decisions (Yarnal 1993; Haak and Ulbrich 1996).

The time-intensive nature of manual tracking was particularly important. Most of the manual tracking studies listed above considered only 5 to 15 years of data with time intervals of 12 or 24 hours. In the late 1900s,
however, digital data became both more widely available and higher resolution. New advances in the observation network (e.g., satellites and buoys) were a major reason (Barry and Carleton 2001). The introduction of global climate models provided an additional source of data that met all requirements for cyclone detection and tracking climatologies (König et al. 1993). This new abundance of adequate data made the benefit of less processing time outweigh the up-front costs of developing an algorithm for automation (König et al. 1993; Ulbrich et al. 2009).

Development of automated methods for cyclone detection and tracking began in the 1980s. Early automated tracking was developed by (Williamson 1981), but this method was mathematically complex and ran slowly as a result (König et al. 1993). It was also intended for weather prediction and has not been widely applied for synoptic climatology. Some algorithms (e.g., Lambert 1988; Serreze et al. 1993) were developed for cyclone detection but lacked a tracking component. Another early algorithm that was a true detection and tracking algorithm for synoptic climatology was an honors thesis at the University of Melbourne by Rice (1982), which inspired the now widely used algorithm developed by Murray and Simmonds (1991).

Regardless of the exact timing and origin of automated detection and tracking algorithms, they did not become common until a proliferation exploded around the world in the early 1990s. Between 1990 and 1995, separate algorithms were developed in Israel (Alpert et al. 1990), France (Le Treut and Kalnay 1990), Australia (Murray and Simmonds 1991), Germany (König et al. 1993), Japan (Ueno 1993), New Zealand (Sinclair 1994), the United Kingdom (Hodges 1994), and the United States (Serreze et al. 1993; Serreze 1995). Several other algorithms have been introduced since then, mostly from European researchers (e.g., Blender et al. 1997; Lionello et al. 2002; Wernli and Schwierz 2006; Hanley and Caballero 2012). Several more algorithms are referenced in the review paper by Ulbrich et al. (2009), and about two dozen are identified by the Intercomparison of Mid Latitude Storm Diagnostics (IMILAST) project (Neu et al. 2013).

No tracking algorithm is exactly the same, and along with the proliferation of algorithms has come a variety of research foci and a diversification of detection, tracking, and summarization styles. Rather than examine any one algorithm in depth, this section attempts to examine the broad diversity of these algorithms at each decision point.
4.3.2. Input Data: Analyses, GCMs, and Reanalyses

The first source of diversity is data selection. Formerly, input data comprised weather charts created by various weather bureaus and meteorology offices (e.g., Klein 1957; Reed and Kunkel 1960; Overland and Pease 1982). By the 1990s, ECMWF and other agencies provided digital versions of such charts as gridded fields of each variable in their atmospheric analyses (e.g., Alpert et al. 1990; Serreze et al. 1993; Sinclair 1994). These analyses were the main data source for observational studies, but GCMs provided an alternative data source for climate experiments and projections. Several studies have used cyclone detection and tracking algorithms to assess the accuracy of GCMs in recreating observed patterns (e.g., Lambert 1988; Le Treut and Kalnay 1990; Lionello et al. 2002). Additionally, application of cyclone detection and tracking algorithms to GCM output has been used to project the impact of global warming on cyclone activity (e.g., Pinto et al. 2007; Ulbrich et al. 2009; Mizuta 2012). More recently, cyclone detection and tracking algorithms have been applied to regional climate models, which are designed to better capture aspects of smaller spatial domains (e.g., Tilinina et al. 2014; Bromwich et al. 2015).

The biggest benefit of using atmospheric analyses to study cyclone behavior is their basis on actual observations. They are created by assimilating observations with the output of numerical weather prediction (NWP) models at sub-daily time intervals (Serreze and Barry 2014). This is preferable to merely interpolating observational data because a) our physical understanding of the atmosphere is adequate to create a superior product and b) our observational network involves an assortment of various collection techniques that have changed over time, so they would have to be combined in a more sophisticated manner than simple statistical interpolation anyway.

Three sources of uncertainty in analyses are a) uncertainty and bias in the atmospheric model, b) measurement errors in the observations, and c) biases in the assimilation system. From a statistical standpoint, that first type of uncertainty (arising from model output) is particularly difficult to define or constrain (e.g., with error bars) (Shackley and Wynne 1996). It arises in part from natural variability in the climate system, in part from imperfect understanding of the atmosphere that leads to imperfect representation, and in part from intentional estimation and simplification of known processes to make algorithms run faster on finite computer power (Flato et al. 2013). Because of the convoluted interaction of multiple sources of uncertainty, some of which is irreducible
(meaning we can never measure it because the amount of uncertainty is unknown), model uncertainty is an especially cumbersome limitation of climate science (Shackley and Wynne 1996; van der Sluijs 2012).

Error and bias arising from model uncertainty is mitigated by assimilation with observational data. In contrast, a long-term simulation on a GCM relies almost exclusively on its internal algorithms. GCMs avoid measurement errors and assimilation bias, which may seem like an advantage. On the other hand, they are not constrained by observations, and model uncertainty is a much greater part of the overall uncertainty in GCM output. In fact, it is typical for a GCM to experience “tuning” with additional parameters to prevent it from drifting off into an unrealistic preferred climate state (Flato et al. 2013). For this reason, assimilated analyses are often considered the “optimal blend” of observational and modeling approaches (Serreze and Barry 2014, p.19), providing complete coverage of space and variables in a more sophisticated manner than interpolation while still remaining constrained by observations.

Regional climate models are appealing because by limiting the spatial domain, computing resources can be devoted to finer spatial resolution or more sophisticated treatment of physical processes particular to that region. Theoretically, this should reduce model uncertainty in cyclone detection and tracking, especially for capturing high winds (Bromwich et al. 2015; Hewson and Neu 2015). However, regional climate models are dependent on their boundary conditions, and often many cyclones affecting a region of interest migrate into that region from elsewhere. Therefore, the model domain must be chosen carefully for such studies and should be much larger than the area of interest (Côté et al. 2015).

Despite the major drawback of model uncertainty in GCMs, they provide one more benefit for climatology: the error and bias is consistent even across long time intervals. By contrast, our observational network is constantly changing. For example, the modern era of weather satellites began with the launch of TIROS-N in 1978 (Netting 2012) and observation density has greatly increased for the upper 2000 m of the (ice-free) ocean since the beginning of the Argo Project in 2000 (Belbeoch 2015). Data assimilation techniques are also continually tweaked (Jung et al. 2006; Serreze and Barry 2014). These changes in observational network can lead to abrupt and artificial inhomogeneities in the time series of examined variables (Sterl 2004; Ulbrich et al. 2009). Although not important for meteorological predictions, these inhomogeneities are problematic for any climatology because they may be mistaken for physical trends (Bengtsson et al. 2004; Sterl 2004; Eisenman et al. 2014).
Problems with inhomogeneities have been partially reduced by the development of atmospheric reanalyses, which apply the short-term prediction techniques of analyses to historical observations using a consistent model and assimilation system (Dee et al. 2011). Changes to observational networks can still be problematic (Sterl 2004; Wang et al. 2006), but because the model and assimilation system are constant, fewer opportunities exist for inhomogeneities to appear.

The first atmospheric reanalysis was developed by the National Center for Environmental Prediction (NCEP) and National Center for Atmospheric Research (NCAR) (Kalnay et al. 1996). Almost immediately, this new dataset was incorporated in cyclone detection and tracking studies (Serreze et al. 1997; Sinclair 1997). Since then, reanalyses have become a common input for cyclone detection and tracking (e.g., Lionello et al. 2002; Zhang et al. 2004; Wernli and Schwierz 2006; Inatsu 2009) and are in fact the only observational data source discussed in the review by Ulbrich et al. (2009). The two most commonly used reanalyses are the NCEP/NCAR reanalysis and the family of reanalyses produced by ECMWF (ERA and its predecessors; Gibson et al. 1999; Uppala et al. 2005; Dee et al. 2011). The Japanese Meteorological Agency’s (JMA) 25-year reanalysis (JRA-25; Onogi et al. 2007) was used by Inatsu (2009). These and others have been compared in several sensitivity studies (Allen et al. 2010; Hodges et al. 2011; Tilinina et al. 2013).

4.3.3. Variation of Input Data: Resolution, Projection, and Study Area

Several variations exist for cyclone input data beyond the broad categories of analyses, GCMs, and reanalyses. For instance, each model and assimilation system has its own algorithms that may impact cyclone detection and tracking results. The gridded fields from any output may also vary in spatial resolution, spatial extent, temporal resolution, temporal range, and projection. Combined, these factors result in notable variation for cyclone climatologies derived from different datasets even when a single detection and tracking algorithm is used (Hodges et al. 2003; Raible et al. 2008; Tilinina et al. 2013).

8 Note that for models, the output resolution may vary independently of the underlying model resolution (Jung et al. 2006; Ulbrich et al. 2009).
4.3.3.1. Spatial Resolution

Spatial resolution has a strong impact on feature detection. If spatial resolution is too coarse, then some synoptic scale features might be too small for the algorithm to identify. Finer resolution permits the detection of more features (Blender and Schubert 2000), but finer resolution is not always superior. If the input data has too fine a scale, then an excessive number of features will be detected, including many that are smaller than the spatial scale of interest. These excess features will complicate tracking because the density of features is greater. They also require greater computer resources (Jung et al. 2006).

Several studies have examined the sensitivity of cyclone detection and tracking to different spatial and temporal resolutions using algorithms that associate cyclone tracks from different tracking methods based on similar track position, length, and time. Blender and Schubert (2000) applied a detection and tracking algorithm to the Hamburg climate model ECHAM4, which has a spatial resolution of about 1.125° by 1.125° and temporal resolution of 2 hours. They then applied the same algorithm to versions of ECHAM4 output with reduced spatial and/or temporal resolution and found that significantly more cyclones were detected in the finer resolution inputs. About 75% correspondence was recorded between results using the original data and results using data that had been degraded to ten times coarser.

In a similar study, Jung et al. (2006) considered the role of spatial resolution in the European Centre for Medium-range Weather Forecasting (ECMWF) 29rl model cycle, which is traditionally used for short- and medium-term weather forecasting. They performed two experiments. In one, they tweaked spatial resolution of the model that produced the SLP field used as input for cyclone detection and tracking. In the other, they ran the model as normal and then altered resolution of the SLP field output (mirroring Blender and Schubert (2000)).

The latter experiment showed that more cyclones are identified from finer spatial resolutions. The only types of cyclones that were robust to spatial resolution were especially long-lived (over 6 days) and far-moving (over 6000 km) systems. However, the former experiment showed that similar impacts occur if the resolution of the physical model is similarly altered. In other words, fewer cyclones are detected if important aspects such as orography and frontal zones are underrepresented in models. For both experiments, the Mediterranean, Arctic, and several high elevation regions were particularly sensitive to spatial resolution.
Highlighting the importance of spatial resolution, studies that compare cyclone and tracking results using different atmospheric reanalyses consistently find that the finer resolution data yields greater cyclone frequency (Hodges et al. 2003; Trigo 2005; Raible et al. 2008; Hodges et al. 2011; Tilinina et al. 2013). However, sensitivity to spatial resolution is greater at coarser resolutions. Results cease to be sensitive around a spectral resolution of T255 (roughly 0.75° latitude) (Jung et al. 2006; Tilinina et al. 2013). Weak and moderate cyclones (pressure > 960 hPa) seem to be more sensitive to the input data than strong cyclones (Pinto et al. 2005; Tilinina et al. 2013).

4.3.3.2. Temporal Resolution

Temporal resolution has received less attention in the literature but is arguably the most important variable for cyclone tracking. The ideal temporal resolution is fine enough so that features move only one or two grid cells each time interval (Zolina and Gulev 2002). The fewer feasible matches between features in one observation time and the next, the less likely the algorithm is to make incorrect tracks (Blender and Schubert 2000). If the temporal resolution is too coarse, there will be too many feasible matches between two observation times. A coarse temporal resolution can also lead to problems presenting results, as discussed in Section 4.3.6. The main down side to using finer temporal resolution is that it requires more data, which lengthens computing time. Therefore, the ideal temporal resolution is the coarsest possible temporal resolution that is still fine enough to track accurately.

In the study by Blender and Schubert (2000) introduced above, perfect correspondence was found between 93% of tracks when comparing the 2-h and 4-h resolutions. This indicates that even a small change in temporal resolution can have some impact on results. Increasing the time interval to 12-hr or 24-hr led to correspondence of 73% and 45%, respectively. Results were more robust to spatial resolution. Making the spatial resolution five or ten times coarser only decreased correspondence to 83% and 75%, respectively. Blender and Schubert (2000) concluded that although increasing either spatial or temporal resolution is beneficial, the temporal resolution needs to be no coarser than 12-hr to produce reliable results. Moreover, the benefits of halving the temporal resolution lessen as temporal resolution becomes finer.
4.3.3.3. Projection

Many gridded datasets used as input for cyclone detection and tracking algorithms are provided on grids of latitude and longitude, meaning that the area of grid cells is greater at low latitudes than high latitudes. Since more cyclones are detected with finer resolution data, this presents the potential for bias toward more cyclones being detected at higher latitudes (Sinclair 1997; Leonard et al. 1999). Additionally, tracking cyclones along the edges of a latitude-longitude grid can be difficult since adjacent locations on Earth may not be adjacent on the grid. Some studies that focus on mid-latitude regions ignore these issues (e.g., Lionello et al. 2002; Benestad and Chen 2006), but they are important to address for global or polar studies (Sinclair 1997).

Two steps are necessary to adequately reproject the data. First, data need to be regridded to a new projection. This new projection should alleviate both the area and edge problems described above, such as a polar stereographic projection (e.g., Murray and Simmonds 1991; Sinclair 1994) or polar Lambert projection (e.g., Hanley and Caballero 2012). Second, the noise generated by this process can be smoothed out (Sinclair 1994). Note, however, that smoothing results in a coarser effective spatial resolution (Serreze et al. 1997).

4.3.3.4. Spatial Extent

Cyclone detection and tracking has been performed for a variety of spatial domains, including the Mediterranean (e.g., Alpert et al. 1990; Trigo et al. 1999; Lionello et al. 2002), Arctic (e.g., Serreze et al. 1993; Serreze 1995; Serreze and Barrett 2008; Nishii et al. 2015), North Atlantic (e.g., Blender et al. 1997; Serreze et al. 1997), and North Pacific (Inatsu 2009). Other studies have considered cyclone climatologies of one or both hemispheres (e.g., Le Treut and Kalnay 1990; König et al. 1993; Simmonds and Murray 1999; Wernli and Schwierz 2006). The choice of spatial extent has two main impacts on cyclone detection and tracking algorithms.

First, researchers may choose to optimize cyclone detection and tracking for a particular region. For instance, reprojecting input data is less important for the mid-latitude Mediterranean Sea than the high-latitude Arctic (see above). More careful consideration of how to handle cyclone splits and merges may be desirable in

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9 Consider trying to track a cyclone that goes from (89°N, 90°E) to (89°N, 90°W) on a Plate Carrée or Mercator grid.
complicated regions such as the North Atlantic (Tsukernik et al. 2007) or North Pacific (Inatsu 2009). The Mediterranean region has received much focused attention in part because of its unique characteristics (Trigo et al. 1999) and in part because algorithms have low agreement in that area (Neu et al. 2013).

Second, the choice of a sub-hemispherical spatial extent can lead to spurious results, particularly along the edges of the study area. For instance, if a cyclone appears near the edge in a certain observation time, it may not be obvious whether that cyclone was generated at that time or merely migrated into the study area (Serreze 1995; Gulev et al. 2001). In a Lagrangian approach, where life cycle events of each storm system are counted and analyzed, this poses a major issue that can only be truly solved by performing the analysis on an entire hemisphere and then subsetting the results. However, the time required to process a cyclone detection and tracking algorithm can be greatly reduced by limiting the spatial extent. This tradeoff has led to a variety of spatial extents for the same region of interest. For example, the Mediterranean has variably been examined for the domains 0 - 60°N, 0 - 60°E (Alpert et al. 1990), 24.75 - 50.625°N, 15.75°W - 45°E (Trigo et al. 1999), and 25 - 55°N, 10°W - 40°E (Lionello et al. 2002).

4.3.3.5. Numerical Prediction and Assimilation in Reanalyses

Spatial resolution and extent, temporal resolution, and projection all have direct impacts on the detection and tracking of cyclones, but results may also vary because of differences in how the input data were generated. Section 6.3.2 describes some of the differences between analyses, reanalyses, and GCMs, but variation also exists amongst reanalyses. The most obvious and important differences are the spatial and temporal resolution of reanalysis output (Raible et al. 2008; Hodges et al. 2011; Tilinina et al. 2013). Spatial resolution within the numerical weather prediction model is also important (Jung et al. 2006). However, many other modeling and assimilation decisions may also impact reanalysis output, and therefore cyclone detection and tracking results.

These indirect effects are more difficult to assess because of the many complex interactions operating within models, but several general statements can be made about the second generation of atmospheric

\footnote{Many of the same principles discussed here for reanalyses are also applicable to the differences that arise when using different climate models for cyclone detection and tracking.}
reanalyses (more specifically, CFSR, ERA, and MERRA) based on sensitivity studies by Allen et al. (2010), Hodges et al. (2011), and Tilinina et al. (2013). Encouragingly, these reanalyses show strong agreement for the basic features of cyclone behavior. The main mid-latitude storm tracks for both hemispheres are consistent, as are major regions of cyclogenesis and cycloysis. Seasonal changes from winter to summer, such as average cyclone intensity and preferred storm track positions, are also consistent.

Track matching shows better correspondence for strong cyclones than weak cyclones; accordingly, more variance amongst reanalyses exists for the summer season, when cyclones tend to be weaker. Correspondence between reanalyses is also better over oceans than continents. It is worst at high elevations because of the need to extrapolate SLP for high elevations and because of the representation of complex topography in each reanalysis.

ERA and CFSR show better agreement than MERRA. When using the same detection and tracking algorithm, MERRA yields significantly higher overall cyclone frequency than either ERA or CFSR despite having the same temporal resolution and a comparable spatial resolution (Tilinina et al. 2013). The distribution of central pressure in MERRA seems to have different tail-behavior, as well. Hodges et al. (2011) identified more and deeper “deep cyclones” (central pressure < 940 hPa) in MERRA, although Tilinina et al. (2013) note fewer “very deep cyclones” (central pressure < 920 hPa). Both Allen et al. (2010) and Tilinina et al. (2013) suggest that MERRA is less similar to ERA and CFSR because it uses a non-spectral resolution. However, other differences amongst the three reanalyses exist. For instance, each one assimilates different assemblages of satellite data (Hodges et al. 2011).

Whatever the reason for these discrepancies, they are considered minor for most regions (excepting areas of high elevation in summer), but results of any study will still be more robust if multiple datasets are assessed instead of one (Raible et al. 2008).

4.3.4. Step 1: Detection

4.3.4.1. Identification of Cyclone Centers

Two main frameworks exist for detecting cyclones: 1) identifying cyclones as minima in SLP (e.g., Serreze 1995; Wernli and Schwierz 2006) or GPH (e.g., Alpert et al. 1990; Blender et al. 1997), and 2) identifying cyclones as maxima in vorticity (e.g., Hodges 1994; Sinclair 1997). Using SLP or low-level GPH (1000 or 850 hPa is typical)
continues the tradition of defining cyclones as synoptic-scale systems of low pressure (Barry and Carleton 2001). Using vorticity, on the other hand, recognizes a more general definition of cyclones as any circulation of air that rotates in the same manner as the Earth (counter-clockwise in the northern hemisphere and clockwise in the southern hemisphere) (American Meteorological Society 2012). Each method highlights a different aspect of cyclonic systems, and each has its own strengths and weaknesses.

The most basic approach to identifying minima in SLP or GPH is to search through every grid cell in the input field and identify grid cells that having a lower pressure than their eight neighbors (Figure 4.1) (e.g., Serreze 1995; Blender et al. 1997; Wernli and Schwierz 2006). After that, several steps can be used to refine the collection of identified centers. For instance, a minimum SLP or GPH gradient in the area around each candidate center is often enforced to eliminate heat lows and other weak minima that would not be classified as synoptic scale cyclones (e.g., Blender et al. 1997; Trigo et al. 1999; Hanley and Caballero 2012). Mountainous regions are sometimes masked from analyses to remove spurious minima that occur because SLP and low-level GPH must be identified by extrapolation at high elevations (e.g., König et al. 1993; Serreze 1995; Hanley and Caballero 2012).

![Figure 4.1](image.png)

**Figure 4.1.** An example SLP pressure field (in hPa). Minima are highlighted in red, with their eight nearest neighbors in orange. The green grid cell is not a minimum because it is only lower than three of its eight nearest neighbors.

More complicated methods have also been employed to identify minima from SLP. For instance, some methods use interpolation to refine the location of identified minima (e.g., Alpert et al. 1990; Murray and Simmonds 1991; Sinclair 1997). Benestad and Chen (2006) used a calculus-based approach combined with a
Fourier approximation of the SLP field to identify minima. Such steps are particularly helpful when working with coarse spatial resolutions but are not necessary when the spatial resolution of the input data is much smaller than the size of the features being identified (Hodges 1994; Blender et al. 1997; Wernli and Schwierz 2006). Recall that detection and tracking algorithms are much less sensitive to spatial resolution when the resolution is at about 0.75° latitude or finer (Jung et al. 2006; Tilinina et al. 2013).

Low-level vorticity has been calculated for either 1000 hPa (e.g., Sinclair 1994; 1997) or 850 hPa (e.g., König et al. 1993; Hodges 1994), but at either level centers are typically identified as maxima in vorticity in the Northern Hemisphere (where cyclonic vorticity is positive) and minima in vorticity in the Southern Hemisphere (where cyclone vorticity is negative). Relative vorticity is the most common type employed (e.g., Hodges 1994; Mesquita et al. 2008), but geostrophic relative vorticity (e.g., Sinclair 1994) and potential vorticity (e.g., Kew et al. 2010) have also been considered. Strength parameters and elevation masking techniques may be applied to limit the number of storms identified, as with SLP or GPH methods.

SLP and GPH methods are often based directly on the manual methods that preceded automation (e.g., Le Treut and Kalnay 1990; Serreze et al. 1993), so they are often intuitive and straightforward. However, using these methods may not be the best in all situations. For instance, SLP and GPH methods are less accurate in detecting and tracking small and fast-moving systems (Sinclair 1994; Hodges et al. 2003). Similarly, it has been claimed that vorticity methods are better at identifying cyclones in the early stages of development (Sinclair 1994; Rudeva et al. 2014). Many SLP and GPH methods also restrict themselves to closed lows (but see Murray and Simmonds 1991), which means the quantity of cyclones being identified may be biased toward stronger systems (Sinclair 1994).

These issues are most pronounced at very low latitudes, where individual cyclonic systems may be difficult to distinguish within a background field of low pressure that exhibits little variation (Hodges 1994). The benefits of using vorticity were also greater in the 1990s, when spatial resolution was coarser. Today, with reanalysis data available at sub-degree resolution, the SLP and GPH methods can more easily identify finer-scale features (Trigo et al. 1999). Moreover, vorticity methods may be too liberal in their identification of cyclones, including small-scale or weak features that would not be considered synoptic scale (Sinclair 1994; Blender et al. 1997). Also notable is that in their review of cyclone detection and tracking methods, Neu et al. (2013) found no
significant difference in cyclone lifespan between vorticity and SLP or GPH methods, which is inconsistent with the supposition that vorticity methods will identify cyclones earlier in their life cycle.

Another reason why some researchers prefer SLP methods is that observed SLP measurements are assimilated into atmospheric reanalyses, while vorticity is purely an analyzed field (Benestad and Chen 2006). In other words, vorticity is more dependent on model algorithms and therefore more prone to model errors (Sinclair 1994). The robustness of SLP observations adds more trustworthiness to results derived from these data (Benestad and Chen 2006). Despite all of the these methodological differences, a review of fifteen cyclone detection and tracking algorithms by IMILAST found no systematic differences in the results between the two main detection methods (Neu et al. 2013).

4.3.4.2. Identification of Cyclone Areas and Multi-Center Cyclones

Although a cyclone may be identified using a central point, synoptic scale systems frequently exceed 1000 km in diameter (Barry and Carleton 2001). Therefore, identifying cyclone area can be a useful additional step for understanding the full effects and characteristics of cyclonic systems. One method is to use the last closed contour (or isobar) around a cyclone center as the limits of its area (Trigo et al. 1999; Wernli and Schwierz 2006). Others identify the edge of a cyclone where the Laplacian of SLP reaches zero (Allen et al. 2010; Simmonds and Rudeva 2012). Such methods are straightforward and easily derived from the data, but they can result in interference from nearby centers (Wernli and Schwierz 2006; Hanley and Caballero 2012).

Such interference can be observed in Figure 4.2, which shows 2 hPa isobars for 0900Z 25 September 1989. Cyclone centers are shown in red and cyclone areas based on the last closed isobar are shown in black. Cyclones 4-6 in the Norwegian Sea are all minima in a broad region of especially low pressure. However, none of them exhibits a very large area. This is because the area between two isobars is only assigned to a cyclone center if it is the only center enclosed by the outer isobar. The 978 hPa isobar encloses both cyclone 5 and cyclone 6. Similarly, the area of cyclone 2 is limited by the 982 hPa isobar, which encompasses cyclone 2 and 3. Meanwhile, cyclone 1 enjoys the largest area of any cyclone in the field because it is isolated from other systems. All other cyclone centers exhibit both a lower central pressure and a stronger SLP gradient in a 1000 km radius. One result
of interference is a bias toward smaller cyclone areas in regions of higher cyclone center density (Wernli and Schwierz 2006).

Figure 4.2. Cyclone centers (red points) and areas based on last closed isobar (black) at 0900Z 25 September 1989. Brown 2 hPa isobars indicate SLP.

One way to mitigate this issue is to remove weak centers from consideration, as described above. Another way is to allow one cyclone area to contain multiple centers. Doing so acknowledges that extratropical storms often contain more than one center, especially when two systems split apart or merge together, as is occurring between cyclone centers 5 and 6 in Figure 4.2 (Inatsu 2009; Hanley and Caballero 2012). In their algorithm, Hanley and Caballero (2012) combine two centers into a multi-center cyclone (MCC) if the number of isobars between the deepest center and the last closed isobar at least doubles when the two are combined rather than treated individually. In Figure 4.2, both center 5 and center 6 have three closed isobars when treated individually, but if they were combined they would have seven before interference from center 4. Since the number of additional shared isobars (four) is greater than the number of separate isobars (three), they could be combined into a MCC.

When using vorticity, one method for defining cyclone area is to use the change in sign of vorticity as the boundary (Sinclair 1997), which is analogous to identifying inflection of the SLP field using its second derivative (Simmonds and Rudeva 2012). Another method is to identify areas for which vorticity exceeds some threshold (e.g., $10^5 \text{ s}^{-1}$) and then identify minima within each area (Hodges 1994). One convenience of such a method is how
it more naturally allows for the identification of MCCs. Inatsu (2009) identified MCCs in a similar way, only using areas for which the meridional wind component at 850 hPa exceeded 10 m/s after applying a temporal filter to the wind field to highlight variation on the synoptic scale (a period less than 10 days).

Simmonds and Keay (2000) and Rudeva and Gulev (2007) use a similar concept, defining points along the cyclone’s edge where $\partial \text{SLP}/\partial r = 0$, where $r$ is the radial distance along 36 radii extending from a cyclone center at 10° angular intervals. The area is calculated as the region enclosed by these 36 points. This method is more complicated than using a last closed isobar or vorticity threshold because it involves interpolating the SLP field to a new coordinate system for each cyclone center of each observation time (Rudeva and Gulev 2007). Although the added sophistication may be beneficial, interpolation is computationally intensive.

Related to finding cyclone areas, Lionello et al. (2002) assigned every grid cell in the Mediterranean basin to a different cyclone by treating the SLP field like elevation and then determining the steepest descent path for each grid cell. Descent paths that ended at the same minimum were grouped into the same cyclone region. Since all grid cells are assigned to a cyclone, even those under high pressure systems, this method presents less information about how much area a particular cyclone impacts, but it does provide another means of differentiating between dominant and subordinate centers.

4.3.5. Step 2: Tracking

The basic goal of cyclone tracking is just like any other feature tracking: the same feature must be identified in multiple observation times as it moves over the Earth’s surface. However, cyclone tracking includes greater complexity and uncertainty than other features because cyclones change rapidly. Over the course of a few hours, every measurable property of a cyclone may change, including its size, shape, location, and intensity. One cyclone will often split apart into two individual systems, and often two individual systems will merge together into one cyclone. This complicated reality has led researchers to develop a variety of tracking methods to accompany the variety of detection methods.
4.3.5.1. Nearest Neighbor and Maximum Speed

Many cyclone tracking methods are based on a nearest neighbor approach (e.g., Serreze 1995; Blender et al. 1997; Trigo et al. 1999), which says that, assuming the time interval of the data is sufficiently short, the most likely continuation of the track for a cyclone in time 1 will be the nearest location that has a cyclone in time 2. All tracking methods must also consider the genesis of new cyclones and the lysis of mature ones. The easiest way to identify genesis is if a cyclone has no match in the previous observation time. Similarly, lysis can be identified if a cyclone has no match in the subsequent observation time.

However, this simple approach requires at least a few additional considerations. For instance, cyclones cannot travel infinitely fast, so most algorithms will use a finite search region for nearest neighbors (e.g., Alpert et al. 1990; Serreze 1995; Lionello et al. 2002) or set some maximum speed parameter (e.g., Sinclair 1994; Hanley and Caballero 2012). How exactly this is operationalized varies by algorithm, but the end result is that the maximum propagation speed allowed for cyclones (when defined) ranges from 80 km hr$^{-1}$ (Blender et al. 1997) to 167 km hr$^{-1}$ (Wernli and Schwierz 2006). In other words, with a 6-hr temporal resolution, a cyclone in time 2 must be no more than 480 km or 1000 km from a cyclone in time 1 in order to be considered the continuation of a cyclone track.

The vast difference in maximum speeds allowed has been criticized in reviews of cyclone tracking methods. Neu et al. (2013) consider cyclone propagation speeds of 110 km hr$^{-1}$ to be extreme cases but certainly possible, and they note that many algorithms are too conservative with their speed parameters. Intuitively, the faster the maximum allowed speed, the more likely it is that a cyclone track will be extended. Therefore, increasing the maximum allowed speed means longer tracks with longer lifespans. If the maximum speed allowed is too conservative, then rapidly moving cyclones will be artificially limited to shorter tracks. On the other hand, if the maximum allowed speed is too liberal, then two cyclones with no relationship may be incorrectly linked with the same track. Rudeva et al. (2014) consider a value of 170 km hr$^{-1}$ unreasonable, and the largest used in the algorithms reviewed here is 167 km hr$^{-1}$ (Wernli and Schwierz 2006). Taking all of these studies together, the ideal limit seems likely to fall between 110 and 170 km hr$^{-1}$, but may vary by the purpose of research and the study area.
4.3.5.2. More Complicated Methods

The nearest-neighbor approach can be modified to address several other issues. As discussed in Section 4.3.3.2, if the temporal resolution is too coarse, then the distance travelled by a single cyclone between two observation times will exceed the distance between two cyclones existing in the same observation time (Murray and Simmonds 1991). Using a nearest neighbor approach in such situations would lead to inaccurate results because the nearest neighbors between two observation times may not really be manifestations of the same cyclone. Unfortunately, because of the substantial changes that occur to cyclone size, propagation speed, and separation from other systems throughout a cyclone’s lifespan, the danger of incorrect matching between two observation times is still present (although much reduced) when using fine temporal resolution. Additional information about the cyclones detected in consecutive observation times is often desired to decrease the likelihood of making a wrong match.

One way to reduce the number of mismatches is to reduce the number of features being tracked. Vorticity methods and higher spatial resolution SLP fields often lead to the detection of extrema that are not features of interest. Eliminating some features in the detection stage, such as by enforcing a strength parameter, may be useful (as described in Section 4.3.4). However, any elimination step risks removing some features that really are of interest, and no limit on detected features will completely prevent mismatches by nearest neighbor approaches. Complex synoptic situations can still arise, especially when multiple cyclones occur in close proximity (Gulev et al. 2001).

Therefore, some researchers prefer to complicate the tracking stage instead of or as well as complicating the detection stage. Two basic approaches are common. The first is to consider the other characteristics of cyclones besides location. For instance, if the cyclone in time 1 has already been tracked for several time intervals, its past motion may be used to predict its next position in time 2. In some algorithms, the nearest neighbor search is conducted on the projected location instead of the prior location (e.g., Wernli and Schwierz 2006; Hanley and Caballero 2012). Other algorithms also include projections of central pressure and/or vorticity based on past tendencies (e.g., Murray and Simmonds 1991; Sinclair 1997; Pinto et al. 2005). When multiple characteristics are incorporated, a probability for every potential match is calculated, and then the combination of matches for all
cyclones that creates the highest global probability for the given observation time is used to determine matches (Murray and Simmonds 1991; Sinclair 1994; Allen et al. 2010).

The second common approach is to include background information not specifically related to the cyclone in question. For instance, cyclones tend to propagate more rapidly in their earlier stages of development, so projections that use past motion may be down-weighted (Hanley and Caballero 2012). The background wind field, especially at the 700 hPa “steering level” (Wallace and Hobbs 2006), can also be used to project how cyclones should propagate (e.g., Alpert et al. 1990; Sinclair 1994; Terry and Atlas 1996). Incorporating the actual wind field at each observation time is computationally expensive, so some algorithms consider climatology instead. In general, cyclones are more likely to travel west to east and south to north in the mid and high northern latitudes, so several algorithms use an asymmetrical search area that allows greater propagation in climatologically favored directions (e.g., König et al. 1993; Serreze 1995; Lionello et al. 2002).

A possible limitation of the methods just described is that they all treat cyclones as points, when in fact they extend over large areas. A few attempts have been made at alternative frameworks. For instance, Inatsu (2009) defines cyclone areas using the meridional wind component and then tracks cyclones by searching for area overlap between two observation times. Kew et al. (2010) perform a similar method for potential vorticity anomalies. Such an approach, of course, requires that the distance travelled each time interval is smaller than the size of the features. Hewson (1997) identifies cyclones by both their central pressure and their associated fronts and so has additional information extending over more space. In their review, Neu et al. (2013) identified some outlier behavior for both the Inatsu (2009) and Hewson (1997) algorithms; however, other reasons besides tracking method were cited. Inatsu (2009) used a different pre-processing than other methods that led to smoother data, and Hewson’s (1997) method was the only algorithm included that used a 12-hr time interval. Therefore, it is uncertain whether their different tracking methods had a substantial impact on results.

4.3.5.3. Merging and Splitting

Another issue beginning to receive more attention is the merging and splitting of cyclones. Traditionally, merging and splitting has been treated as a nuisance to overcome while tracking. For instance, if a single cyclone in
time 1 splits into two systems by time 2, a decision must be made regarding which of the two candidates in time 2 is an extension of the track from time 1. The pressure tendency from the previous observation time can be used to help determine the best match (e.g., Murray and Simmonds 1991; Sinclair 1994). Alternatively, some algorithms will preferentially extend the path to the center with lower central pressure (e.g., Serreze et al. 1993). Hanley and Caballero (2012) introduce a more elaborate method that biases the selection to make the longest possible tracks. Merging presents a similar problem, except that it involves matching two cyclones in time 1 with a single cyclone in time 2.

Three recent cyclone detection and tracking algorithms have taken greater interest in splitting and merging, treating them as meaningful events rather than tracking obstacles (Inatsu 2009; Kew et al. 2010; Hanley and Caballero 2012). Doing so provides richer information about cyclone interactions, especially regarding the manner of cyclogenesis (which may involve a split) and cyclolysis (which may involve a merge). However, it also assumes that a distinction can be made while tracking between a splitting event and cyclogenesis that occurs nearby but independent from a pre-existing cyclone, and likewise between a merging event and cyclolysis that occurs to a nearby but unrelated cyclone.

4.3.6. Step 3: Summarization

No cyclone detection and tracking exercise is complete without summarizing characteristics. Summarization may be for a single case study or an aggregation of many cyclones, but in either case, a variety of variables are usually considered to capture cyclone behavior in the study area. Several standard variables of interest include: spatial distribution of cyclone center frequency and density, cyclone track density, event frequency (genesis, lysis, splitting, and merging), and cyclone-affected areas; and basic statistics regarding central pressure, deepening rate, depth, propagation speed and direction, track length, and lifespan. All of these variables can be examined separately for different regions and seasons, and temporal trends are sometimes calculated, especially to examine the relationship of cyclone characteristics to global climate change (e.g., Gulev et al. 2001; Zhang et al. 2004).
4.3.6.1. Cyclone Frequency Measures

Over time, several measures of cyclone “frequency” and “intensity” have been used; however, the current standards are demonstrated in review papers by Ulbrich et al. (2009) and Neu et al. (2013). One measure of frequency is the raw cyclone center count, or the number of cyclone centers that appear in a grid cell summed over all observation times in a particular period. In other words, it measures how frequently the area is being impacted by cyclones. This method is problematic when the grid projection is not equal area, so a cyclone center density is often considered instead. Cyclone center density is the number of cyclone centers occurring per time interval per unit area. For instance, if the unit area is defined as a circular region with radius 200 km centered on each grid cell, then the unit area is $40,000\pi$ km$^2$. Often this is reported as a percent density. A cyclone center density of 5% means that a cyclone occurs in that $40,000\pi$ km$^2$ region in 5% of all observed fields. If the unit area is large enough, the cyclone center density can exceed 100%, which means on average more than one cyclone center exists within the unit area per observed field.

One drawback of using cyclone centers as the measure of frequency is that cyclones are an areal feature, not a point feature. For this reason, it may be preferable to use cyclone areas to calculate the frequency with which any location is impacted by cyclones (Wernli and Schwierz 2006). However, as discussed in Section 4.3.4.1, accurately measuring cyclone areas is difficult, although incorporating multi-center cyclone identification in the detection stage may improve accuracy (Hanley and Caballero 2012).

Another issue occurs when temporal resolution is coarse enough so that cyclone centers move more than one grid cell each time interval (as mentioned in Section 4.3.3.2). Using a cyclone center count in such cases will underestimate the cyclone frequency in areas with fast cyclone propagation. According to Zolina and Gulev (2002), temporal resolution needs to be about 2-3 hours to limit underestimation below 5% (using 2.5° latitude by 2.5° longitude input data).

Several methods can be used to alleviate this issue. The simplest is to smooth the data, counting the total number of cyclone centers within a certain distance of each grid cell (e.g., Serreze et al. 1997; Hanley and Caballero 2012; Neu et al. 2013). Another method is to interpolate cyclone positions at a finer temporal resolution (Zolina and Gulev 2002).
Alternatively, cyclone track density can be used as a measure of frequency. Track density is measured as the number of tracks that travel through a grid cell or unit area during a given period (Neu et al. 2013). In other words, whereas cyclone center density measures how frequently an area is impacted by a cyclone, track density measures how many unique cyclones impact an area. In comparison to track density, center density tends to highlight regions where cyclones move more slowly or become stationary (Sinclair 1994). Since cyclones typically travel more slowly as they mature (Hanley and Caballero 2012), this also means that lysis regions will tend to be more strongly highlighted by center density than track density. The question of whether to use track density or center density is not new, and discussion of the merits of each date back at least to the 1950s (e.g., Klein 1956).

4.3.6.2. Cyclone Intensity Measures

Cyclone intensity measures are similarly varied. Following the practice of weather forecasting, some studies use the convenient measure of central SLP (e.g., Trigo 2005; Benestad and Chen 2006; Jung et al. 2006). However, since the background SLP field tends to be lower at higher latitudes, this method will present a bias toward more intense cyclones at higher latitudes. This can be addressed by subtracting a background SLP field first (e.g., Lionello et al. 2002; Hodges et al. 2003). Another approach is to measure the difference in SLP between the center and some defined distance, such as the cyclone edge or a 1000 km radius (e.g., Blender et al. 1997). This measure of depth is independent of spatial and seasonal variations, but it relies on either accurate identification of the cyclone edge or else an arbitrary radius.

The Laplacian of SLP or the geostrophic relative vorticity (which are proportional to each other) can also be used (e.g., Murray and Simmonds 1991; Serreze et al. 1997). The Laplacian is independent of the background field, which is beneficial, but it is also scale-dependent. This means that cyclones of vastly different sizes are not directly comparable. For this reason, Simmonds and Keay (2000) propose using cyclone depth (D),

\[ D = \frac{1}{4} \nabla^2 p R^2 \]

(Equation 4.1)

which is proportional to both the Laplacian of pressure throughout the cyclone area (\( \nabla^2 p \)) and the cyclone's size (the effective radius (R) the cyclone would have if its area were circular). Although Simmonds and Keay (2000)
conclude that this Laplacian-depth measure is the best measure of intensity, it is not without drawbacks. Namely, it requires confidence in the cyclone size measure, which currently has no standard. Perhaps reflecting the continued uncertainty over the best intensity measure, subsequent papers by the Simmonds research group have each included multiple intensity measures, including Laplacian of central pressure, Laplacian-depth, cyclone radius, and even simple central pressure (e.g., Simmonds et al. 2008; Simmonds and Rudeva 2012).

Also of interest is how quickly cyclones intensify (“deepen”) or diminish (“fill”). Based on the suggestion of (Roebber 1989), it is now standard to measure deepening rates (DR) as the change in central pressure ($\Delta p_{cent}$) per time interval ($\Delta t$) scaled by a reference latitude ($\phi_{ref}$), or

$$DR = \frac{\Delta p_{cent}}{\Delta t} \frac{\sin \phi}{\sin \phi_{ref}}$$

(Equation 4.2),

where $\phi$ is the latitude of the cyclone (Serreze et al. 1997; Trigo et al. 1999; Jung et al. 2006; Rudeva et al. 2014).

4.4. Ensembles and Sensitivity Studies

With so many algorithms available with such rich variety of methods, one may question how consistent results will be for two different algorithms and whether one algorithm is superior to another. Several papers have examined these questions for a variety of algorithms, datasets, and parameters. More recently, the IMILAST project is a comprehensive and systematic comparison of about two dozen detection and tracking algorithms developed since the early 1990s (Neu et al. 2013). The importance of input data (spatial and temporal resolution, spatial extent, and projection) was already discussed in Section 4.3.3. This section focuses on the differences that occur in results from cyclone detection and tracking algorithms even when the input data is consistent.

The assemblage of results from available automated cyclone detection and tracking algorithms exhibit many similar characteristics. For instance, the main storm tracks in the North Pacific and North Atlantic Oceans are common features (Ulbrich et al. 2009; Neu et al. 2013), with cyclones typically deepening on the western sides of the basins and reaching maximum intensity on the eastern sides (Ulbrich et al. 2009). In winter, the Asian
continent and Mediterranean also experience relative maxima in cyclone frequency (ibid.). Cyclogenesis is common in or near areas of complex orography (Ulbrich et al. 2009; Rudeva et al. 2014).

Algorithms show less consistency around complex topography than over open oceans (Raible et al. 2008; Neu et al. 2013), which is not surprising considering problems extrapolating SLP from high elevation (Sinclair 1994). The fact that some algorithms mask high elevation regions adds to variability, especially since some legitimate cyclones do originate at high elevation (Neu et al. 2013). Exploring elevation impacts even further, Rudeva et al. (2014) performed a sensitivity study with or without an elevation mask on over a dozen algorithms. They found that when elevation is masked, the tracks of some cyclones may be truncated to the point of no longer satisfying lifespan thresholds, which creates a bias toward lower cyclone frequency. Many cyclones experience their maximum deepening rates early in their development, so this track truncation also impacts the histogram and distribution of maximum deepening rates. Intensity measures are more robust to elevation since cyclones typically experience their greatest intensity late in their life cycles and over oceans. Raible et al. (2008) found that intensity was a more robust measure than frequency in other comparisons as well.

Many tracking algorithms, and all those considered by Ulbrich et al. (2009), exhibit greater cyclone frequency in winter than summer. However, the larger study by Neu et al. (2013) showed several algorithms showing summer had greater cyclone frequency. The difference in cyclone seasonality comes from disagreement over weak and short-lived cyclones (Raible et al. 2008; Neu et al. 2013). These marginal cases may be discarded for a variety of reasons, including strength parameters, coarse spatial resolution, and maximum distance between centers (Neu et al. 2013). Rudeva et al. (2014) found about 800 to 1000 cyclones reliably detected each year by all algorithms. Better agreement exists amongst algorithms for winter than summer, again because winter exhibits larger, deeper cyclones that are more easily identified (Neu et al. 2013; Rudeva et al. 2014).

The cause for many of the differences observed amongst different algorithms is often unclear and difficult to predict. For instance, although anticipated by earlier studies (e.g., Sinclair 1997), differences in lifespan or genesis/lysis locations between SLP/GPH and vorticity methods were not found by either Neu et al. (2013) or Rudeva et al. (2014). Neu et al. (2013) concluded that there were too many parameters and too many choices to be made in these algorithms to distinguish the impact of any individual parameter or choice.
However, some general statements can be made. The amount of variability in results when using different algorithms is about the same as the variability in results when using different reanalyses (Raible et al. 2008). That variability is worst for marginal cases (weak, short-lived cyclones) (Leonard et al. 1999; Raible et al. 2008) and around complex topography (Raible et al. 2008; Neu et al. 2013), which means these algorithms are better suited for studying large, strong, long-lived cyclones. However, the basic features of cyclone activity show strong agreement amongst the current assemblage of algorithms (Ulbrich et al. 2009; Neu et al. 2013).

4.5. An Alternative: Eulerian Approaches to Synoptic Studies

The last two sections detailed the process of detecting and tracking individual cyclones for creating climatologies of cyclone activity. This process is Lagrangian, focusing on the development of features as they move through space. However, it is not the only approach to understanding synoptic scale activity.

In Eulerian methods, the researcher focuses not on moving features, but on fixed locations. A common practice is to examine a series of GPH or SLP fields and apply a band-pass filter to the temporal frequency domain for each grid cell (Blackmon et al. 1984; Trenberth 1991; Rogers 1997; Norris 2000). A band pass filter of 2-6.5 days (Blackmon et al. 1984; Walter and Graf 2005; Jaiser et al. 2012) or 2-8 days (Trenberth 1991; Rogers 1997; Wu et al. 2010) will highlight variability in the GPH or SLP field associated with synoptic scale cyclones. These methods have been used to characterize the main mid-latitude storm tracks (Blackmon et al. 1984; Trenberth 1991), explore the dynamical relationships between baroclinic waves, eddy covariance fluxes, cloudiness, and SSTs (Trenberth 1991; Norris 2000), relate storm track variations to large-scale oscillation patterns (Rogers 1997; Walter and Graf 2005), and assess the impacts of global warming on synoptic-scale systems (Wu et al. 2010; Jaiser et al. 2012; Nishii et al. 2015).

Eulerian methods have some benefits over cyclone detection and tracking algorithms. First, they are simpler and faster to implement, so a three-dimensional view of cyclone activity can be easily obtained by applying band-pass filters to multiple levels of the atmosphere (Chang et al. 2002). Second, band-pass methods require fewer choices on the part of the researcher. Cyclone detection and tracking relies on a suite of parameters, starting with the detection variable, but also how to define a cyclone, what thresholds to impose, and how to
connect centers in consecutive observation times into tracks. Interaction amongst these variables makes it difficult to attribute differences between algorithm to any one parameter (Neu et al. 2013; Rudeva et al. 2014). By contrast, the Eulerian approach requires only one major decision: the limits of the band-pass filter. For this reason, Eulerian methods are less subjective and more consistent. Third, Eulerian methods might be preferred because they more closely relate to the experience of people living in fixed locations (Shippee and Atkinson 2014).

However, Eulerian methods are also limited in some ways. Variation with a frequency of 2-6 days will include features unrelated to synoptic weather systems (Anderson et al. 2003). At the same time, some cyclones will exhibit variation on timescales outside of that narrow range (Mesquita et al. 2008). Gulev et al. (2001) describe several additional limitations, such as how Eulerian methods provide no information about individual cyclones, where they form, or how they develop. They cannot provide details about cyclone deepening rates, propagation speeds, lifespans, or track lengths. All of these characteristics can be described using Lagrangian methods.

Moreover, cyclone detection and tracking output can still provide the same Eulerian perspective presented by band-pass filtering by, for instance, calculating cyclone track density or cyclone center frequency for each grid cell. So despite the computational drawbacks, cyclone detection and tracking algorithms have the potential to offer richer information about synoptic activity.
 CHAPTER 5. DATA & METHODS

5.1. MERRA, CFSR, and ERA Reanalyses

Atmospheric reanalyses provide gridded representations of atmospheric states spanning multiple
decades, generated by a consistent atmospheric model and assimilation system (Saha et al. 2010). Precipitation
(total and large-scale), SLP, GPH, u and v wind components, and temperature variables are derived using data from
three modern atmospheric reanalysis systems: 1) NASA’s Modern Era Retrospective-Analysis for Research and
Applications (MERRA; Rienecker et al. 2011; http://gmao.gsfc.nasa.gov/merra/); 2) NOAA’s Climate Forecast
System Reanalysis (CFSR; Saha et al. 2010; http://cfs.ncep.noaa.gov/cfsr/); and 3) the European Center for Medium
Range Weather Forecasts’ (ECMWF) interim European Reanalysis (ERA; Dee et al. 2011; http://apps.ecmwf.int/datasets/).
GPH is obtained at 400, 500, and 600 hPa, whereas the wind and temperature
data are obtained at more than 20 levels from 1000 to 100 hPa, the exact levels being different for each reanalysis.
Lastly, the 2 m temperature is also obtained for each reanalysis.11

MERRA (1979-2015) was developed by NASA for the satellite era and uses version 5 of the Goddard Earth
Observing System Data Assimilation System (GEOS-5). Analysis is performed at a horizontal resolution of 2/3
degree longitude by 1/2 degree latitude at 72 levels. Surface data, near surface meteorology, selected upper-air
fields and vertically integrated fluxes and budgets are produced at one-hour intervals. For the purposes of cyclone
detection, SLP and precipitation fields were downloaded both at a 6-hr intervals at the native resolution and at 3-
hr intervals and a reduced 1.125° latitude by 1.125° longitude resolution.

CFSR (Saha et al. 2010) extends from 1979-2010. The global atmosphere resolution is ~38 km (T382) with
64 levels extending from the surface to 0.26 hPa. The atmospheric model includes observed variations in carbon
dioxide along with changes in aerosols and other trace gases and solar variations. Atmospheric variables were
obtained at a 0.5° latitude by 0.5° longitude resolution at 6-hr time intervals. Note that the second generation of

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11 The 2 m level is 2 m above the displacement height. The displacement height is the height at which the wind
would fall to 0 m s⁻¹ if the wind profile was logarithmic (usually between 0.6 and 0.8 the canopy height) (Arya
2001).
CFSR (CFSv2) extends beyond 2010, but several differences to the operational model make inhomogeneities between 2010 and 2011 a concern (Saha et al. 2014).

ERA is an effort by the ECMWF to prepare for their next general reanalysis that will replace their older 40-yr reanalysis system (Dee et al. 2011). ERA data are available from 1979 onwards. Compared to past ECMWF reanalyses, ERA has increased the number of pressure levels from 23 to 27, uses improved model physics, new humidity analysis, more extensive use of radiances, an improved fast radiative transfer model, and includes additional cloud parameters. Atmospheric variables were obtained at 0.75° latitude by 0.75° longitude resolution and 6-hr intervals. Note that all of the reanalyses use observed satellite-derived snow cover and sea ice extent as boundary conditions.

Comparison of reanalysis data to tethersonde soundings from the Tara drifting ice station in 2007 showed that ERA has the closest agreement to temperature, humidity, and wind speed observations in the lowermost 890 m of the atmosphere in the Arctic (Jakobson et al. 2012). However, CFSR exhibits the most accurate surface conditions (2 m air temperature and 10 m wind speed), likely because of its more complicated treatment of sea ice. All three reanalyses exhibit warm biases at the surface and underestimate wind speeds throughout most of the lower atmosphere. Temperature accuracy suffers under high winds and when inversions are present. Warm surface air temperature bias in ERA (up to 2°C) may be related to reanalysis tendencies toward the melting point of sea ice, whereas the most commonly observed temperature is the freezing point of seawater (Lüpkes et al. 2010).

All three reanalyses have positive biases in monthly precipitation amounts relative to the station data; this is expected when station data have not been corrected for gauge undercatch of solid precipitation. Linear correlations between station time series (1979-2010) and reanalysis precipitation at the closest grid cell range from less than 0.5 to more than 0.9. No one reanalysis consistently outperforms another, although MERRA precipitation in spring is comparatively large. Consistent with earlier studies, focusing on earlier generation reanalyses (Serreze et al. 2005), the mix of high and low correlations with station data in part reflects both sampling issues and precipitation magnitude. In areas of complex topography, a single station is more likely to poorly represent the grid cell value represented in the reanalysis than in areas of simple topography. In turn, in areas of very low precipitation, the correlations obtained are sensitive to even small errors in the observations and
reanalyses. Past efforts have also shown that reanalyses consistently provide better depictions of Arctic precipitation than do datasets based on satellite retrievals.

The strength and variability of the AFZ is assessed using all three atmospheric reanalysis, but only MERRA and ERA are used for cyclone detection and tracking because ERA and CFSR have been shown to yield exceptionally similar cyclone results (see Section 4.4.3 or Raible et al. 2006; Allen et al. 2010; Hodges et al. 2011; Tilinina et al. 2013). Because 3-hr resolution data offers several benefits for cyclone tracking (see Section 5.6.4), the results in Chapter 6 focus on the application of the cyclone detection and tracking algorithm to MERRA data for the period 1979-2014. However, CESM-LE data are only available at a 6-hr resolution for the period 1990-2005. Therefore, to control for the effects of temporal resolution, reanalysis differences, and external forcing, results from CESM-LE cyclone analysis are compared to 6-hr MERRA and ERA data for the period 1990-2005 (see Appendix to Chapter 8).

5.2. The CESM Large Ensemble

5.2.1. Uncertainty in Climate Change Projections

Any projection of future climate change using a climate model is impeded by three main sources of uncertainty related to 1) external forcing, 2) model bias, and 3) internal variability of the climate system. External forcings include human emissions of greenhouse gases and aerosols, land use change, and natural forcings like solar output and volcanic emissions. Since human activities in particular are difficult to predict, uncertainty about future external forcings is typically constrained by a standard set of scenarios that represent possible pathways for human emissions and land use change. A new set of scenarios called Representative Concentration Pathways (RCPs) were incorporated into the Fifth Assessment Report (AR5) of the Intergovernmental Panel on Climate Change (IPCC 2013).

Model bias encompasses any error in model output resulting from imperfect representation of the Earth’s systems. These errors may result from simplifications of the Earth system, such as when a coarse spatial resolution is unable to resolve small-scale features. For example, mesoscale cyclones like polar lows are too small to be detected with a 2° latitude/longitude resolution but can be detected with a 0.44° latitude/longitude resolution (Zahn et al. 2008). Greater complexity does not always lead to more accurate results, however, as exemplified by
the depiction of permafrost area by CMIP5 models (Slater and Lawrence 2013). Errors may also result from shortcomings in our understanding of natural processes. For instance, much uncertainty still exists regarding the roles of clouds and aerosols in the climate system (Boucher et al. 2013).

One way to constrain model bias is to run simulations for the same period and external forcings in a suite of climate models and consider the mean and spread of the ensemble. This is a critical function of the Program for Climate Model Diagnosis and Intercomparison (PCMDI) at the Lawrence Livermore National Laboratories. However, such ensembles are problematic in several ways. For one thing, many of the modeling centers participating in CMIP5 submit multiple models that share many components. Different modeling centers will also share or mimic techniques used in other places. This means that the individual members of CMIP5 cannot be treated as independent, which complicates statistical quantities like the ensemble mean (Flato et al. 2013).

Additionally, the ensemble spread in CMIP5 and other intercomparisons encompasses not only model bias, but also unforced internal variability in the climate system, which can be the dominant source of uncertainty in climate model projections on decadal time scales, especially for regional projections (Kirtman et al. 2013). This is of particular concern with regard to the summer AFZ, which exhibits substantial interannual and regional variability (Crawford and Serreze 2015).

5.2.2. Details for the Large Ensemble

To understand how a particular climate model expresses the internal variability of a regional feature like the AFZ, it is preferable to run the same model for the same time period and the same external forcings multiple times, with only small round-off differences to initial conditions differentiating each run (Kay et al. 2015). Such an approach has been undertaken using the Community Earth System Model version 1 (CESM), resulting in a “large ensemble” (CESM-LE) of model runs (ibid.). Variation amongst the ensemble members represents the stochastic processes of internal climate variability, as described by Lorenz (1963) for the impact of initial conditions on numerical weather prediction models.

A complete description of the CESM-LE is provided by Kay et al. (2015). All simulations in the CESM-LE use the 1° latitude/longitude version of CESM (with Community Atmospheric Model version 5; 0.9x1.25_gx1v6; Hurrell...
et al. 2013). Following the standard of current earth system models (Flato et al. 2013), CESM includes coupled atmosphere, ocean, land, and sea ice components. It also incorporates a land carbon cycle and ocean biogeochemistry. Historical radiative forcing is used for the period 1920-2005, and RCP8.5 is used thereafter.

RCP8.5 is a high-emissions scenario with an anthropogenic radiative forcing of about 8 W m^{-2} at 2100 (Collins et al. 2013). The 95% confidence interval for the global annual mean surface air temperature difference between 1986-2005 and 2081-2100 in this scenario is 2.6 to 4.8°C (based on CMIP5; ibid.). Although more ensemble members have been added since the inception of CESM-LE, the research presented here is limited to members 1 through 30.

CESM-LE data is obtained from Earth System Grid (https://www.earthsystemgrid.org/home.html). Monthly temperature, u wind, and surface pressure fields are available for the periods 1920-2005 (1850-2005 for Member 1) and 2006-2080. The vertical grid for CESM uses a hybrid pressure-sigma system, so for direct comparison to the reanalysis data, these data are first interpolated to a pure pressure grid with 50 hPa intervals from 100 hPa to 800 hPa and 25 hPa intervals from 800 hPa to 1000 hPa using the interp1d function in scipy (https://www.scipy.org). Interpolation was conducted in only the vertical dimension to expedite computation time. Three-dimensional interpolation did not yield noticeably different results in a sample dataset. Sea ice and snow cover fraction and reference height temperature were also obtained at monthly time scales. Reference height temperature roughly corresponds to the 2 m level used in the atmospheric reanalyses.

Because the present study involves not only detection of the summer AFZ, but also cyclone characteristics, sub-daily data were also desired. CESM-LE data are available from Earth System Grid at a 6-hr temporal resolution for both 1990-2005 and 2071-2080. Temperature, surface pressure, SLP, and total precipitation were obtained for each period. Large-scale precipitation is not available at a 6-hr resolution, so daily large-scale precipitation was downloaded for the same periods.

One problem with using the CESM-LE instead of CMIP5 is the loss of information about model bias. To compensate for this loss, all results derived from CESM-LE through 2005 are compared to the output from the

12 This scenario is called RCP8.5 because 8.5 W m^{-2} was the “target” anthropogenic radiative forcing at 2100. However, the RCP process was more complicated than setting a number. For instance, each climate model in CMIP5 responded to the established time series of emissions in a slightly different way, so the mean for the ensemble of models was a radiative forcing just below 8 W m^{-2} at 2100. For a longer explanation, see Collins et al. (2013) or http://alexcrowford0927.wix.com/ipccar5ch12final#changes-in-forcing-agents/cvk1.
atmospheric reanalyses described above. Care must be taken when comparing a model simulation to observations. First, no ensemble member should be expected to perfectly replicate observations because both observations and model simulations contain substantial internal variability. Second, the ensemble mean in CESM-LE mutes internal variability, providing a better depiction of the long-term climate state than any individual ensemble member or even observations (assuming model bias is accounted for). Therefore, for any measured parameter in the observed period 1990-2005, CESM-LE is considered to provide an accurate result if the observations from atmospheric reanalyses fall within the 95% range of the histogram of that parameter for all 30 CESM-LE members. Assuming a normal distribution, the 95% confidence interval is calculated as ± 1.96σ, where σ is the standard deviation of the parameter in question.

5.3. Ancillary Data

Elevation in several maps is defined using NOAA’s ETOPO1 1 Arc-Minute Global Relief Model (Amante and Eakins 2009; https://www.ngdc.noaa.gov/mgg/global/global.html). The Circumpolar Arctic vegetation map (CAVM; Walker et al. 2005); http://www.arcticatlas.org/maps/catalog/index) is used to identify the southern limit of the tundra, and MODIS land cover (Friedl et al. 2010; Channan et al. 2014; http://glcf.umd.edu/data/lc/) is used to identify the northern limit of continuous boreal forest. Observed sea ice concentration and extent are taken from the combined Nimbus Scanning Multichannel Microwave Radiometer (SMMR, Jan 1979 – Jul 1987), the Defense Meteorological Satellite Program (DMSP) Special Sensor Microwave Imager (SSM/I, Jul 1987 – Feb 2007) and the Special Sensor Microwave Imager/Sounder (SSMIS, Mar 2007 – Dec 2014) 25-km gridded sea ice concentration data product from the NASA Team sea ice algorithm (Cavalieri et al. 1996; http://nsidc.org/data/NSIDC-0051), which is distributed by the National Snow and Ice Data Center (NSIDC).

13 Even if we had a perfect model, we should not expect observations to perfectly match the mean climate state represented by an ensemble mean of model outputs because observations result from the mean climate state plus chaotic internal variability.
14 The data are close enough to a normal distribution for this to be a fair assumption, but a non-parametric 95% confidence interval was also investigated. It did not change results in any substantial way.
5.4. Defining AFZ Parameters

5.4.1. Eady Growth Rate

The maximum Eady growth rate (EGR) is a common value used in synoptic research (e.g., Walter and Graf 2005; Serreze and Barrett 2008; Simmonds and Rudeva 2012) that relates baroclinic instability to the vertical wind shear and static (in)stability (Vallis 2006). In other words, it measures how conducive the environment is to synoptic development. Following Serreze and Barrett (2008) and Simmonds and Rudeva (2012), EGR was calculated using Equation 3.3. For calculation of EGR at 500 hPa, vertical profiles of wind and potential temperature were calculated as finite differences between 400 hPa and 600 hPa. Following Simmonds and Lim (2009), seasonal values of EGR were obtained by first calculating these values at 6-hr time intervals and then taking an average.

5.4.2. Temperature Gradient Magnitude

As in Crawford and Serreze (2015), temperature gradient magnitude fields, as well as meridional and zonal gradients, were calculated using a Sobel operator. A Sobel operator calculates a zonal and meridional gradient using a 3 by 3 kernel. The four cells that share an edge with the center cell are weighted twice as heavily as the four cells sharing a corner. The Haversine formula was used to convert from latitude and longitude to meters.

5.4.3. Arctic Frontal Zone Strength

As discussed by Lynch et al. (2001), the AFZ, whether on the basis of observations or model output, can be defined using any one of several thermal variables. While Serreze et al. (2001) focused on a thermal front parameter as given by Hewson (1998), Crawford and Serreze (2015) found it more straightforward to define AFZ strength simply in terms of the horizontal temperature gradient. Regardless of the exact parameter, the AFZ is traditionally defined by the climatological position of high frontal frequency or temperature gradient magnitude. Based on the climatological position of the strongest July or summer (JJA) temperature gradients north of 60°N as seen in longitudinal cross sections, the AFZ is a narrow area aligned with the Arctic Ocean coastline from about 42°E eastward to 234°E (126°W). (See Section 3.3.)
With this definition in mind, summer AFZ strength at any longitude and vertical level is defined as the mean horizontal temperature gradient magnitude (|∂T/∂X|) for the two grid cells closest to the Arctic Ocean coastline. The one exception to this definition is that the meridional temperature gradient (∂T/∂y) is used in latitudinal cross sections and to differentiate between summer and winter gradients. To better capture the vertical extent of the AFZ, this calculation was performed at 2 m (reference height for CESM-LE), and about 25 pressure levels$^{15}$. As demonstrated by the thermal wind equation (Equation 3.1), stronger horizontal temperature gradients lead to greater vertical wind shear and (as seen in Equation 3.3) greater baroclinic instability.

At monthly time scales, this framework works well for measuring AFZ strength near the surface, but it may not be ideal in the mid troposphere. Near the surface, the AFZ exhibits no monthly variability in location; the strongest horizontal temperature gradients are also fixed to the coastline. The only variation is the intensity of the temperature gradient magnitude (or AFZ strength). At higher levels of the atmosphere (e.g., 700 hPa or 500 hPa), the zone of strongest monthly temperature gradient magnitude is not always well-aligned with the coast.

For example, compare the mean July temperature gradient magnitude at 700 hPa in 1980 to 1994 (Figure 5.1). Based on the definition of the AFZ described above, 1994 shows a strong AFZ in eastern Siberia and Chukotka, whereas 1980 shows a weak AFZ. However, a zone of strong temperature gradient magnitude does exist in 1980; it is simply located to the south of the coast between 150°E and 180°E. Similarly, strong temperature gradients lie over the Beaufort Sea coast at 700 hPa in 1980 and north of the coast in 1994. Such monthly variation in the location of the strongest horizontal temperature gradients reflects not only the ability of the Arctic front and its attendant jet-like feature to shift position, but also their tendency to form ridges and troughs. Although land-sea contrasts facilitate the development of these features and influence their mean position, short-term variations suggest that a better measure of AFZ strength at middle levels of the troposphere might be one that is independent of location.

$^{15}$ The number varies: MERRA has 25, ERA has 27, and CESM-LE has 23.
5.4.4. The Migrating Arctic Front: An Alternative Metric

Therefore, an alternative measure of summer AFZ strength is introduced that defines the AFZ for any given month as all grid cells exceeding the 90th percentile of horizontal temperature gradient magnitude within the region bounded by 42°E to the west, 234°E to the east, 60°N to the south, and 88°N to the north. The northern boundary eliminates potentially spurious values near the pole which suffer from edge effects and converging meridians. The framework, hereafter called the “migrating Arctic front” (MAF) has the benefit of measuring variation in both AFZ strength and location.

However, this metric does have several limitations of its own. For instance, using a percentile ensures that the MAF will always have the same number of grid cells, which is an artificial limit. It might be more realistic to identify the MAF as any grid cell that exceeds some horizontal temperature gradient magnitude threshold; however, that threshold would have to be determined separately not only for each vertical level, but also for each data set since higher resolution data tends to yield stronger temperature gradients (see Section 7.2). This could add bias to any comparison of datasets. Additionally, a standard threshold measure would only capture spatial variability of the MAF, omitting variability in intensity. Variability in intensity is desirable to make the MAF more comparable to other measures of the AFZ.

Another downside of using the MAF is that because it has no consistent spatial location, it is difficult to break the MAF into regions to consider finer scale variation. Regionalization of the AFZ into sectors has proved a useful framework for describing spatial variation in its climatological and interannual strength (Crawford and
Serreze 2015). Therefore, the year-to-year variability in MAF location and strength is only considered for the MAF as a whole. Hereafter, “MAF strength” refers to the mean horizontal temperature gradient magnitude for grid cells exceeding the 90th percentile of the bounding box described above for any given level and month/season.

Going forward, AFZ strength (horizontal temperature gradient magnitude at a fixed position along the coast) is the primary metric. MAF strength and mean latitude is also considered when examining the relationship between the AFZ in the mid troposphere and cyclone activity.

5.5. Clustering Analysis and AFZ Sectors

With varying strength, the AFZ, as seen in the time-averaged horizontal temperature gradient at the surface, stretches across over 200° of longitude from the Kola Peninsula to the Bering Strait to the Canadian Arctic Archipelago (Figure 3.2). Results from a series of cluster analyses by Crawford and Serreze (2015) showed that the AFZ can be divided into sectors based on the annual variability of summer horizontal temperature gradient strength. To assess whether these sectors also apply to the AFZ as depicted in CESM-LE, the same cluster analysis is applied to each of its members for the period 1979-2005.

Since variability in AFZ strength is used to make these sectors, a fixed location for the AFZ is first determined. Following Crawford and Serreze (2015), the location of the AFZ at a particular longitude is defined as the two adjacent grid cells (by latitude) north of 60°N with the highest mean July reference height temperature gradient magnitude for the period 1979-2005. AFZ strength is defined for each longitudinal band of the CESM-LE data (about every 1°) from 41°E eastward to 126°W, excluding the Bering Strait (the longitudes at which the summer AFZ is present). Data from all CESM-LE members are combined for this calculation so that the AFZ has a standard location, but clustering analysis is then performed separately for each member following the methods reported by Crawford and Serreze (2015). As appropriate, these sectors are used to make regional generalizations of the following results.
5.6. Cyclone Detection and Tracking

Cyclone characteristics are obtained using an updated detection and tracking algorithm based on prior work by Serreze et al. (1993), Serreze (1995), and Serreze and Barrett (2008). Several features of more recent algorithms, especially by Wernli and Schwierz (2006) and Hanley and Caballero (2012) have been adopted. The main innovation is the explicit identification of multi-center cyclones and splitting and merger events, which have only previously been considered in a few algorithms (Inatsu 2009; Hanley and Caballero 2012). The algorithm was written using Python 2.7.11 (https://www.python.org). Required modules include GDAL 2.0.0 (GDAL Development Team 2015; http://www.gdal.org), NumPy 1.10.4 (van der Walt et al. 2011; http://www.numpy.org), SciPy 0.17 (Oliphant 2007; https://www.scipy.org), and Pandas 0.14 (McKinney 2010; http://pandas.pydata.org)16. A full list of input parameters can be found in the Appendix to Chapter 5 (hereafter 5A).

5.6.1. Cyclone Center Detection

The present study seeks to resolve relationships between the AFZ and cyclone characteristics over the Arctic Ocean, so SLP was chosen as the most appropriate variable for cyclone detection. This choice makes results more directly comparable to several prior Arctic studies (Serreze et al. 1993; Zhang et al. 2004; Serreze and Barrett 2008). Additionally, since synoptic-scale systems that cross into the Arctic Ocean are of chief interest, shortcomings of SLP methods related to small, weak systems (Hodges et al. 2003; Sinclair 1994) are not especially problematic. The one concern would be if erroneous conclusions about genesis areas resulted from late detection of cyclones (Neu et al. 2013; Rudeva et al. 2014).

The input SLP data are provided on longitude-latitude grids, and this unequal spatial resolution leads to biases in the density of cyclone centers detected at high latitudes (Sinclair 1997). Therefore, the SLP data were first re-projected using cubic convolution onto the northern hemisphere Equal-Area Special Sensor Microwave Imager (SSM/I) Grid 2.0 (EASE Grid 2.0; Brodzik et al. 2012) with a grid cell size of 100 km by 100 km. The EASE Grid 2.0 is

16 The algorithm also runs well using pandas 0.12 or 0.13, but beginning with version 0.15, changes to Pandas data frame indexing make the core functions for the algorithm run 25-50 times slower than with previous versions. The algorithm has been confirmed as compatible with NumPy 1.8 and SciPy 0.14, as well.
based on a Lambert’s equal-area azimuthal projection centered on 90° latitude. A problem with cyclone detection using SLP is that extrapolation of SLP from high elevation areas often leads to artificial pressure minima (König et al. 1993; Serreze 1995). Following Neu et al. (2013) and Tilinina et al. (2013), all grid cells with an elevation exceeding 1500 m were masked before implementing the detection stage.

**Figure 5.2.** SLP isobars (interval at 2 hPa) and cyclone detection at 0300Z 25 September 1989. Red dots indicate SLP minima that are accepted as cyclone centers. Black dots indicate minima that are rejected. Solid white shading masks elevations greater than 1500 m.

Cyclone detection involves three steps. First, local minima in SLP are identified as grid cells for which the pressure is lower than the eight adjacent grid cells. Second, following previous studies (e.g., Blender et al. 1997; Trigo et al. 1999; Hanley and Caballero 2012), a strength parameter is applied. A circle with a 1000 km radius is drawn around each minimum, and the average SLP difference between the minimum and each grid cell intersecting that circle is calculated. If the average SLP difference is less than 7.5 hPa for a minimum, it is
considered too weak to be a cyclone of interest and is discarded. The exact value of the parameter, which matches that used by Hanley and Caballero (2012), was determined from a series of sensitivity studies.

Figure 5.2 demonstrates the impact of the strength parameter. Each minimum detected in the SLP field is marked with a dot, but only the red minima have a strong enough SLP gradient to be accepted as cyclone centers. Despite the inclusion of a 1500 m elevation threshold, many rejected minima are shallow features in areas of complex topography, such as the Pacific Coast Ranges in Canada, the Alps in Europe, and coastal Greenland.

5.6.2. Area and MCC Detection

The final detection step simultaneously calculates cyclone area and determines the presence of multi-center cyclones (MCCs). This step is notable because it recognizes that a) cyclones are not point features but areas and b) a single system may contain two or more distinct but closely related minima in SLP. MCC detection is based on the method of Hanley and Caballero (2012), but the shared to unshared ratio is calculated using cyclone area enclosed by the last closed isobar instead of the number of isobars. Because of the novelty of MCC detection, the following account is written with special detail.

For each center, a series of isobars is drawn around the center with increasing pressure until a) more than one center is enveloped by the highest isobar or b) the highest isobar also envelopes a maximum in SLP. The isobar interval is a changeable input parameter, but 2 hPa was chosen to match past studies. All isobars are measured relative to the center being considered. If a maximum is enveloped by the highest isobar, then the process is immediately stopped and the previous isobar is taken to be the spatial extent of the cyclone area. For example, if a maximum is enveloped by the fifth isobar measured from center 13 in the SLP field, then the area for center 13 is defined by the fourth isobar and the depth is 8 hPa.

If multiple centers (but no maxima) are enveloped by the highest isobar, then a multi-center cyclone (MCC) may exist. Three additional criteria are considered to assess whether a MCC is present. First, the new center(s) cannot already be assigned to an area. Second, all centers must be close to each other, so a maximum MCC distance parameter was created. The parameter was set to 1000 km from the initial center, which roughly matches the scale of synoptic systems (but could be refined through a sensitivity study). Third, following Hanley
and Caballero (2012), MCCs must satisfy a shared/unshared criterion. Assuming the distance criterion is satisfied, all enveloped centers are tentatively called a MCC, and the isobar series is continued until another center or maximum is enveloped. At this point, the area contained by the highest isobar enveloping only the initial center (i.e., the highest “unshared” isobar) is divided by the area contained by the highest isobar enveloping all centers (i.e., the highest “shared” isobar). If the ratio of unshared area to shared area is less than the ratio threshold, the centers are considered part of a MCC and the shared area is assigned to the cyclone. Otherwise, the possibility of an MCC is eliminated, and only the unshared area is assigned. For this study, the tolerated ratio was set to 0.50 after testing several values on a subset of data and comparing them to a visual analysis. This means that for a group of centers to be considered part of a MCC, the cyclone area of the MCC must be at least double the cyclone area of one center alone.

One problem with this approach is that MCC assignment will depend on the order in which centers are examined. If two nearby centers have very different unshared areas, then the unshared/shared ratio may be higher than the ratio threshold for one center and lower than the ratio threshold for the other. To standardize the process, cyclone area calculations are always conducted beginning with the lowest pressure center in the SLP field and ending with the highest pressure center. Thus the initial center in a potential MCC is always the lowest pressure center, and is always the center used to measure the unshared/shared ratio.

The results of the MCC decision process for one cyclone center field can be observed in Figure 5.3. Two different MCCs were detected: one over northwest North America and one over the UK. For the former, center 13 has the lowest pressure, so it was the first center considered of the three. The yellow isobar roughly marks the area for center 13 if it were a single-center cyclone. This is the unshared area. However, by encompassing centers 14 and 15, the area for this cyclone system more than doubles, and both additional centers are within 1000 km of center 13. The group passes the tests and is reclassified as a MCC.

Centers 8 and 9 make an even more obvious MCC. They are about 630 km apart, and their shared area is about 1.5 million km$^2$. Alone, center 8 has an unshared area of 0 km$^2$, meaning that even an isobar of 2 hPa above its central pressure encircles center 9.

Two other cases were rejected during the MCC detection phase. One is centers 10 and 11. These two centers are slightly beyond the distance threshold, lying 1280 km apart, so a ratio test was never even performed.
However, if it a ratio test were calculated, the green isobar roughly marks out the shared area for these two
cyclone centers. This area is not more than double the unshared area of center 11, so these two centers would fail
the ratio test.

Centers 4 and 5 lie only about 500 km apart, but they were also rejected during the MCC test. As indicated
by the pink isobar, in order to include centers 4 and 5 within the same closed isobar, several other clearly distinct
cyclone centers would also have to be enveloped. Since not all the centers can be part of a MCC, centers 4 and 5
remain individual cyclones.

Although this treatment of MCCs is based on Hanley and Caballero (2012), it has some differences. First,
Hanley and Caballero (2012) do not include SLP maxima as a limit on area calculations. Especially at lower
latitudes, some cyclone areas are detected that have unrealistic shapes (e.g., including most of the Intertropical
Convergence Zone) unless maxima are included as an area limit. Second, Hanley and Caballero (2012) do not
mention any maximum distance between centers in a MCC. Third, instead of measuring the ratio between
unshared and shared areas, they consider the ratio between the number of shared and unshared isobars. The two
methods should be somewhat related, but compared to manual inspection of SLP fields, the use of area seemed to
lead to more accurate results. Lastly, Hanley and Caballero (2012) always combine centers into MCCs pair-wise and
allow a maximum of three centers in an MCC. This algorithm takes a different approach. It allows unlimited centers
in a MCC, but it will only perform one MCC test. This difference leads to different false negatives during MCC
detection. The algorithm of Hanley and Caballero (2012) will fail to capture a MCC that has four or more centers all
with very similar pressures, whereas this algorithm will fail to capture an MCC with two or three centers that have
very different pressures. Based on comparison to manual detection, the method used here seems preferable for
the input data being used. However, manual inspection will not identify all possible situations, and sensitivity
studies would be required to truly determine the ideal parameters. Spatial resolution of the input data and other
algorithm parameters may influence which method is better for a given application.
5.6.3. Cyclone-Associated Precipitation (CAP)

After cyclone detection for each SLP field, the precipitation associated with each cyclone center (cyclone-associated precipitation or CAP) is calculated. The method used here is based on Finnis et al. (2007) and Stroeve et al. (2011) but also incorporates cyclone areas calculated by the detection algorithm. The basic principal behind the CAP calculation is proximity. More specifically, the theory is that any precipitation, especially non-convective precipitation, that falls within a cyclone area or near a cyclone center is likely linked to that cyclone.
CAP is calculated in several steps. First, for any given observation period, all areas of contiguous grid cells with a large-scale precipitation rate exceeding 1.5 mm day$^{-1}$ are identified. Second, all precipitation areas are assigned to a cyclone if a) the precipitation area intersects the cyclone area or b) the precipitation area lies within 250 km of the cyclone center. The cyclone area calculations are sensitive to cyclone interference, leading to smaller cyclone areas when two separate systems lie close together. Using a minimum radius to determine intersections with precipitation areas alleviates this bias. However, the inclusion of cyclone area in addition to the minimum radius makes it more likely that larger cyclones will be associated with more precipitation. Third, the total precipitation for each precipitation area is assigned to its respective cyclone. Using total precipitation here recognizes that extra-tropical cyclones can still include convective lifting. Lastly, if one precipitation area is associated with multiple cyclones, the precipitation is partitioned amongst the cyclones so that each grid cell of precipitation is assigned to the nearest cyclone center.

For MERRA, both large-scale and total precipitation fields are provided at 3-hr time intervals, matching the SLP fields. ERA precipitation fields have a 12-hr interval, whereas the SLP fields are available at a 6-hr interval. For CESM-LE, although both SLP and total precipitation are provided at a 6-hr interval, the large-scale precipitation fields are only provided at daily time intervals. In order to match the SLP time interval, 12-hr precipitation fields are divided into two 6-hr fields with precipitation distributed evenly between the two halves. The daily fields are likewise divided into four equal parts. This necessity reduces both the precision and accuracy of CAP calculations derived from ERA and CESM-LE compared to MERRA.

5.6.4. Cyclone Tracking

Consider two fields of cyclone centers at times $t_n$ and $t_{n+1}$. To determine whether a cyclone center at $t_n$ also exists at $t_{n+1}$, continuing its track, a search radius of 450 km is used, corresponding to a maximum allowed propagation speed of 150 km hr$^{-1}$. The effective maximum propagation speed in previous algorithms has a broad range (compare the 80 km hr$^{-1}$ used by Blender et al. (1997) to the 167 km hr$^{-1}$ used by Wernli and Schwierz (2006)), but Neu et al. (2013) comment that cyclones can move faster than 110 km hr$^{-1}$ in extreme cases. Therefore, 150 km hr$^{-1}$ was chosen as a reasonable round number.
A predicted location (x*) is also incorporated for cyclone centers that have existed in multiple observation times. Following Wernli and Schwierz (2006) and Hanley and Caballero (2012), the predicted location x*(t_{n+1}) is a reduced linear continuation of the cyclone’s past propagation and calculated by the equation x*(t_{n+1}) = 0.75 • (x(t_n) - x(t_{n-1})), where x(t_n) is the cyclone’s current location and x(t_{n-1}) was the cyclone’s location at time t_{n-1}. The factor of 0.75 is included because the propagation speed of cyclones tends to decline with age. To be considered a continuation of a cyclone center from t_n, a center at t_{n+1} must lie within 450 km of both x(t_n) and x*(t_{n+1}). If multiple potential matches exist at t_{n+1} for a single t_n center, the nearest neighbor to the predicted location is chosen as the continuation.

**Figure 5.4.** illustrates a hypothetical tracking situation. Panel A shows the field of cyclone centers at t_n (red) overlain on the field of cyclone centers at t_{n+1} (green). The two centers at t_{n+1} are exactly the same distance from x(t_n). In Panel B, the past motion of the center at t_n (blue to red) is used to predict x*(t_{n+1}), the location of the same center at t_{n+1} (orange). Lastly, Panel C shows how the search area is defined by the intersection of two circles: one around the t_n location, and one around the t_{n+1} prediction. Although the two t_{n+1} centers are the equidistant from the t_n center, only one of them falls within the search area. In this way, predicting the location at t_{n+1} allows the algorithm to make the more likely match between two observation times.

One benefit of using the MERRA 3-hr fields is that, with average propagation speed lying between 10 and 60 km hr^{-1} for most cyclones (Rudeva et al. 2014), cyclone propagation between each time interval is usually only one or two grid cells. This results in very few ambiguous tracking situations and permits a simple nearest neighbor approach as the basis for tracking. ERA and CESM-LE data, which have a 6-hr time interval, yield more complicated tracking situations and rely more heavily on the inclusion of a predicted location.
5.6.5. Cyclone Tracking Events

After all matches have been made between the centers from \( t_n \) and \( t_{n+1} \), the occurrence of cyclone events is assessed, including cyclogenesis and cyclolysis. Cyclogenesis occurs if any center at \( t_{n+1} \) had no corresponding center at \( t_n \). Cyclolysis occurs if any center at \( t_n \) has no corresponding center at \( t_{n+1} \). These events and others are illustrated in Figure 5.5.

Centers can also split apart and merge together. When a cyclone center splits, two centers at \( t_{n+1} \) correspond to the same center at \( t_n \). Conversely, a cyclone merge occurs when two cyclone centers from \( t_n \) correspond to the same center at \( t_{n+1} \). Splitting and merging always involve genesis and lysis, respectively, of a cyclone track. When multiple \( t_{n+1} \) centers are the result of a split, one of them is a continuation of the original track. The remaining centers are new (center genesis). Conversely, when multiple \( t_n \) centers merge at \( t_{n+1} \), the merged center is the continuation of just one track. The remaining tracks are ended (center lysis).

Because this algorithm includes MCCs, cyclones are tracked simultaneously in two manners. The manner described above is the more traditional one, in which tracking is conducted on centers. This works well for single-center cyclones, but for MCCs, tracking is also performed on the entire system, which requires additional considerations. First, each MCC is assigned a primary track for the purposes of comparing system characteristics or counting system frequency. Second, MCCs can experience splitting and merging of cyclone areas even when centers experience no event. An area split occurs when a MCC at \( t_n \) divides into two separate systems at \( t_{n+1} \). An area merge occurs when two separate systems at \( t_n \) become a MCC at \( t_{n+1} \). These events result in system genesis and lysis, respectively.

The top two rows of Figure 5.5 represent simplistic examples of each event, but tracking can become more complicated. For instance, merges and splits may involve more than two cyclone tracks. Additionally, center and area events are not mutually exclusive; neither are merges and splits. It is possible (although rare) for a single cyclone to experience several of these events during a single time interval. An example of one complicated situation called “regenesis” is shown in Figure 5.5. In this case, three times are considered. At \( t_{n-1} \), a single-center cyclone exists with some areal extent. At \( t_n \), the system develops a secondary center. This is an example of a center split. At \( t_{n+1} \), the primary track experiences lysis; however, the secondary center persists. In this case, the secondary center from \( t_{n+1} \) has become a primary center of a new cyclone at \( t_{n+2} \), so area genesis has occurred.
However, the cyclone is clearly related to the system that existed at $t_0$. Therefore, this represents a special type of genesis. Since it involves the redevelopment of an aging cyclone as a new system, it is marked as regenesis.

MCC detection, and detection of regenesis in particular, helps address a documented problem with simple SLP-based methods such as that used by Serreze (1995). These algorithms are prone to unrealistic splitting of a single cyclone track into multiple parts (Mesquita et al. 2009). Often this problem can be fixed by considering tracks that begin in regenesis events as the continuation of the former system instead of new systems. (See example in Section 5A.4.4.)

5.6.6. Cyclone Characteristics

The algorithm returns a large number of cyclone tracks, many of which consist of only a few observation times. Despite the elevation mask and strength parameter of 7.5 hPa (1000 km)$^1$, some detected centers are not synoptic-scale cyclones but heat lows or artifacts related to high elevation. To eliminate some of these unwanted features, only cyclones that last at least 24 hours, travel at least 100 km (1 grid cell), and spend at least one observation time over elevations lower than 500 m were retained. Additionally, only one track was used for each MCC when measuring average cyclone frequency and intensity for the study area. (However, this last choice makes only minor differences to results; see Section 5A.4.3.)

Several cyclone characteristics are assessed, including frequency, intensity, and deepening rate (see full list in Section 5A.2). Following past research (e.g., Hodges et al. 2011; Rudeva et al. 2014; Zolina and Gulev 2002), all maps of cyclone characteristics have been smoothed. The value at any particular grid cell represents the average for a 500 km by 500 km area centered on that grid cell. Frequency is measured using track density, defined as the number of cyclone tracks passing through a 500 km by 500 km area centered on the grid cell for a given period. Similarly, the event density (genesis or lysis) is the number of events that occur in a 500 km by 500 km area for a given period.

Intensity is measured using both the cyclone central pressure and the Laplacian of central pressure. The latter is often preferred because it is independent of latitude and proportional to geostrophic relative vorticity (e.g., Murray and Simmonds 1991; Serreze et al. 1997). Following Roebber (1989), Serreze et al. (1997) and others,
the deepening rate (Equation 4.2) is scaled by latitude to account for the change in the relationship between a unit pressure gradient and the geostrophic wind by latitude. The reference latitude is arbitrary, so 60°N was chosen to match Serreze et al. (1997).

![Diagram of cyclone tracking events]

**Figure 5.5.** Schematics of seven cyclone tracking events. Dark shading indicates cyclone centers and pale shading represents cyclone areas. Blue, red, and green indicate cyclones at \( t_{n-1} \), \( t_n \), and \( t_{n+1} \), respectively. Thick solid lines indicate track continuation (no event occurs); thin dashed lines indicate center-related events; and thin solid lines indicate area-related events.
5.7. Seasonally Variable Northern Annular Mode (SVNam) Index

The AFZ-Arctic cyclone relationship cannot be properly assessed without controlling for large-scale atmospheric patterns that might also impact cyclone activity. Several patterns, including the Northern Annular Mode (NAM), North Atlantic Oscillation (NAO), Pacific Decadal Oscillation (PDO), and Pacific-North America Pattern (PNA), have been linked to Arctic cyclone activity in the past (e.g., Gulev et al. 2001; Zhang et al. 2004; Serreze and Barrett 2008). However, most of these studies focused on the winter season, and only the NAM has been consistently shown to have a significant relationship with summer Arctic cyclones (Ogi et al. 2004; Serreze and Barrett 2008; Simmonds et al. 2008). For the dataset used here, only the seasonally variable version of NAM (SVNam; Ogi et al. 2004) has a significant ($p < 0.05$) relationship with summer CAO+BCEL cyclone frequency for both ERA and MERRA, so the other oscillations are not considered further.

The SVNAM is computed separately for each calendar month and is defined as the leading empirical orthogonal function (EOF) of monthly SLP anomalies north of 40°N. The same inputs are used from EOF analysis as for cyclone detection. The use of an equal area grid removes the need for area-weighting in the EOF analysis. The SVNAM indices derived from MERRA and ERA inputs are compared to the values reported by Ogi et al. (2004), who used GPH from 1000 hPa to 200 hPa from the NCEP/NCAR Reanalysis as inputs instead of SLP. The time periods for the inputs also differ, with Ogi et al. (2004) using 1958-2002 (later extended to 2011; http://www.jamstec.go.jp/ress/ress/masayo.ogi/SVNAM.txt) and MERRA and ERA data being available for 1979-2014. First, for each calendar month, a paired t-test is administered with the null hypothesis being that no difference exists between the SVNAM indices derived from each reanalysis time series. Next, Pearson’s $r$ coefficient is used to assess how the indices co-vary, with the null hypothesis being that no correlation exists between the SVNAM indices from different reanalyses. Parametric tests are used because visual inspection of histograms and Q-Q plots showed no substantial deviation from a Gaussian (normal) distribution.

To assess how similar the SVNAM expressed by CESM-LE is to that expressed by MERRA and ERA, the SVNAM index for each CESM-LE member-month is calculated using a standard EOF solver (Dawson 2016). SVNAM

\[ r_{Spearman} = 0.37 \text{ for ERA and } 0.36 \text{ for MERRA.} \]
is calculated independently from the full period of reanalysis data (1979-2014) and from all 30 CESM-LE members for the period 1990-2005. Since each member of CESM-LE uses the same model and boundary conditions, the latter method can be considered analogous to having 480 years of data for a stable 1990-2005 climate. This is preferable to conducting a separate EOF analysis for each member because 16 years is insufficient to capture the full range of variability in the SVNAM.

After the EOF solvers are constructed, each monthly CESM-LE SLP anomaly field is mapped to the leading principal component of each solver, providing three independent SVNAM index series for each month (one derived from the MERRA solver, one from the ERA solver, and one from the CESM-LE solver). These resulting indices are then compared using the same two statistical tests used to compare amongst the reanalyses. For each calendar month, a paired t-test is used to assess the presence of bias in CESM-LE, with the null hypothesis being that no difference exists between the SVNAM indices derived from different solvers. Pearson’s r coefficient is used to assess how the indices co-vary, with the null hypothesis being that no correlation exists between the SVNAM indices from different solvers.18 If the expression of the SVNAM were identical in both MERRA and CESM-LE, for instance, then the tests would yield a mean difference of 0 and a correlation of 1.

5.8. Statistical Comparison of AFZ and Cyclone Characteristics

5.8.1. Track-by-Track Analysis

For the full MERRA cyclone dataset (1979-2014), several statistical analyses were performed. A track-level analysis was performed for all cyclone centers that crossed through the AFZ and into the combined CAO and BCEL Seas study area (dark blue shading in Figure 1.1; hereafter CAO+BCEL) for the summer season (JJA). The Barents and Kara Seas were omitted because they are primarily influenced by the North Atlantic storm track. For each track, the deepening rate (DR) of a cyclone crossing the AFZ was described as

18 This is a slightly different null hypothesis than the one used when comparing the reanalyses. The reanalysis comparisons test differences in both the construction of the solvers and the time series of data to which the solvers are applied. When comparing CESM-LE to the reanalyses, only construction of the solvers is being tested.
\[ DR = \alpha + \beta_1 \frac{Age}{Lifespan} + \beta_2 \frac{MPAge}{Lifespan} + \beta_3 (AFZ) + \beta_4 \left( \frac{\pi}{12} \sin(t_{local} + \sigma) \right) \]

(Equation 5.1)

where \( AFZ \) is the average AFZ strength \( (|\partial T/\partial X|) \), described in Section 5.4.3) at the time of AFZ crossing for the group of five contiguous coastal grid cells centered on the point of AFZ crossing. Following the baroclinic instability model, strong horizontal temperature gradients at middle and upper levels of the troposphere are favorable for cyclone deepening \( (DR < 0) \) because they lead to greater vertical shear \( (\text{Equation 3.1}) \) and therefore greater baroclinic instability \( (\text{Equation 3.3}) \).

However, other variables may also impact the deepening rate and must be controlled for to ensure that any resulting coefficients are physically meaningful. The first two variables describe a typical cyclone life cycle. \( Age \) is the age of the cyclone at the time of AFZ crossing, \( MPAge \) is the age of the cyclone when it experiences its minimum pressure, and \( Lifespan \) is the age of the cyclone when it experiences lysis. Cyclone age is included because cyclone deepening rate tends to be strongest (most negative) when a cyclone is just forming \( (\text{Rudeva et al. 2014}) \). The older a cyclone is at the point of AFZ crossing, the more likely it is to be filling instead of deepening. Since a cyclone begins filling \( (DR > 0) \) after it hits its minimum pressure, stronger deepening might be expected during AFZ crossing if the cyclone has yet to reach its minimum pressure. Implicitly, these variables are describing the influence of upper-level circulation on surface cyclone development.

The final term represents the diurnal thermal atmospheric tide, a SLP oscillation in the summer Arctic driven by the sensible heat flux from the ground \( (\text{Dai and Wang 1999}) \). In Equation 5.1, \( t_{local} \) is the local time of day, and \( \sigma \) is the phase \( (9 \text{ hr}, \text{determined empirically}) \). The null hypothesis is that, after controlling for the relative age of the cyclone at AFZ crossing, relative age at its point of minimum pressure, and the diurnal cycle of SLP enforced by the thermal tide, the temperature gradient magnitude during AFZ crossing has no impact on the deepening rate \( (\text{i.e., } \beta_4 = 0) \).

In addition to considering the significance of coefficients, models were assessed using the variance inflation factor for multicollinearity amongst right-hand variables, Cook’s distance for outliers, histograms and Q-Q plots for normality of residuals, and the Breuch-Pagan test for heteroskedasticity of residuals.
5.8.2. Seasonal Analysis

For the full MERRA cyclone dataset (1979-2014), Spearman’s rho coefficients\(^\text{19}\) are calculated to estimate the correlation between overall AFZ strength and EGR and cyclogenesis along the Arctic coastline (red outline in Figure 1.1), as well as between AFZ strength and cyclone intensity, frequency, and CAP in the CAO+BCEL for the summer season. A non-parametric (rank-based) test of correlation is used because, based on visual inspection of histograms, not all variables are normally distributed for each time period and dataset assessed. Correlations are calculated both before and after subtracting the linear relationship between SVNAM and the other variables. All statistical analyses are performed separately using AFZ strength at both 700 hPa and 500 hPa. They are then repeated using MAF strength.

For comparing to CESM-LE, these same correlations (except for EGR) are replicated for the period 1990-2005 for MERRA (6-hr inputs), ERA, each CESM-LE member, and all CESM-LE members combined. The SVNAM adjustment is applied to CESM-LE data using the indices from CESM-LE solvers.\(^\text{20}\)

5.9. Climate Change Assessment for 2071-2080

Comparison of the AFZ and summer Arctic cyclone activity in CESM-LE for the period 1990-2005 to MERRA and ERA helps to identify any notable bias in CESM, but it also establishes a baseline for examining climate change projections. To assess how the AFZ will respond to climate change under RCP8.5, the same variables are examined for the period 2071-2080 as 1990-2005 and partitioned into the same regional sectors. Student’s t-tests are applied to mean monthly AFZ strength separately for each calendar month. Each ensemble member is treated as an independent observation, so \(n = 30\) for each population in the t-tests.

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\(^{19}\) For the sake of comparing, Pearson’s correlations were also calculated, but as might be expected from data without a normal distribution, the correlation estimates were much more sensitive to minor adjustments to study area or omission of outliers. This was especially true for the shorter 1990-2005 period.

\(^{20}\) Using the MERRA-based or ERA-based solver makes little difference, though, since the CESM-LE SVNAM based on the CESM-LE solver and the CESM-LE SVNAM based on the reanalysis solvers were highly correlated (\(r > 0.95\) and \(p < 0.001\) for all months).
Similarly, to assess how the AFZ-cyclone relationship will respond, the same cyclone detection and tracking algorithm is applied to each member individually for the period 2071-2080. T-tests are used to test for the presence of differences in cyclone characteristics for any given calendar month or season. The SVNAM is also calculated using the methods described above. Finally, the same correlations are calculated as for the 1990-2005 period for each member individually as well as for all members taken together.
CHAPTER 6. THE AFZ AND SUMMER CYCLONE DEVELOPMENT

Figure 6.1. Latitudinal cross sections along 151°E longitude for (a,b) average meridional temperature gradient (K (100 km)$^{-1}$) and (c,d) vertical velocity ($10^{-2}$ Pa s$^{-1}$) for (a,c) January and (b,d) July 1979-2014, and (e,f) maps of average Eady growth rate (EGR) at 500 hPa (day$^{-1}$), for (e) winter (DJF) and (f) summer (JJA) 1979-2014.

Many of the results in this chapter are also summarized in Crawford and Serreze (2016).
6.1. The AFZ and Baroclinic Instability

Prior studies (e.g., Reed and Kunkel 1960; Serreze et al. 2001; Serreze and Barrett 2008) argue that baroclinicity in the troposphere induced by differential heating of adjacent land and ocean surfaces in the Arctic may help to generate cyclones that impact the Arctic Ocean. **Figure 6.1** shows latitudinal cross sections of the (a,b) meridional temperature gradient and (c, d) vertical velocity in (a,c) January and (b,d) July along the 151°E meridian, a longitude typical of the AFZ. In winter, the only substantial meridional temperature gradients above the surface level are around 30°N. However, in summer, two distinct frontal zones are apparent extending from the surface to the tropopause. The polar front sits at about 40°N, but north of and distinct from that zone lies the AFZ, which is anchored to the coastline. Contribution by surface energy fluxes to the coastal temperature gradients aloft is supported by the presence of upward (negative) vertical velocity (omega) just south of the coastline.

From the thermal wind equation (Equation 3.1), strong horizontal temperature gradients throughout the troposphere create vertical wind shear, which is a major component in the calculation of maximum EGR (Equation 3.3). In winter, when the polar front is strongest and the AFZ is absent, EGRs in excess of 1.0 day\(^{-1}\) are common around the Icelandic Low and the Canadian Rockies (Figure 6.1e). While all the relative maxima in EGR lie in the mid latitudes, the coastline of Siberia is a relative minimum.

In summer, the strength of the polar front is greatly reduced, as are EGRs in the Icelandic Low region (Figure 6.1f). The Eurasian coastline, on the other hand, experiences higher EGRs in summer and is a relative maximum during that season. In other words, the presence of the summer AFZ supports a region of baroclinic instability. Following the baroclinic instability model of cyclone development, the AFZ environment seems to favor cyclone development along the Eurasian Arctic coastline.
Figure 6.2. Frequency of (a,b) cyclone tracks, (c,d) cyclogenesis events, and (e,f) cyclolysis events for (a,c,e) winter (DJF) and (b,d,f) summer (JJA), averaged for the period 1979-2014. Units are number of tracks or events per 500 km by 500 km area per season.
6.2. Results from Cyclone Analysis

6.2.1. Cyclone Characteristics

Figure 6.1 demonstrates that the AFZ is favorable for cyclone development, but is it actually a preferred region of summer cyclogenesis? Figure 6.2 shows the average (a,b) track density, (c,d) cyclogenesis frequency, and (e,f) cyclolysis frequency for (a,c,e) winter and (b,d,f) summer seasons 1979-2014. Cyclone frequency is greater in winter (Figure 6.2a) than summer (Figure 6.2b) on the Atlantic side of the Arctic and Gulf of Alaska. By contrast, activity over the CAO and eastern Siberia is notably higher in the summer season.

Cyclogenesis shows similar seasonal differences. In winter, a broad area of cyclogenesis extends from the Icelandic Low to the Norwegian and Barents Sea. Cyclogenesis is also common in the lee of the Rockies and Greenland, and in northern Baffin Bay. Lastly, cyclogenesis is common in southern Kamchatka and in the Gulf of Alaska. The prominence of the Icelandic Low and Norwegian and Barents Seas as cyclogenesis regions is reduced in summer, while the relative importance of continental cyclogenesis is increased. Eastern Siberia is particularly notable in this regard. It contains a strong maximum in summer cyclogenesis in the Kolyma Lowland (KL in Figure 1.1). The AFZ, however, is not a preferred area for summer cyclogenesis.

For both winter and summer, many of the regions experiencing frequent cyclogenesis also experience frequent cyclolysis, such as the Icelandic Low and the Gulf of Alaska. Other regions exhibit more cyclogenesis than cyclolysis. In winter, these include the Norwegian and Barents Seas and the lee of the Rockies in the USA and Alberta (see also Figure 6.3). In summer, the strongest net genesis regions are the Kolyma Lowland and the lee of the Rockies. The CAO+BCEL is distinct because it does not exhibit substantial cyclogenesis in either season, but it is consistently an area for cyclolysis. This quality is especially apparent in summer when cyclone frequency in the CAO+BCEL increases.
6.2.2. Source Regions for Summer Arctic Ocean Cyclones

One notable finding from Figure 6.2 is that although the AFZ may be a favorable environment for cyclone development, it is not a preferred region of summer cyclogenesis. Where, then, do cyclones affecting the CAO+BCEL originate? The genesis region for every cyclone that spends any part of its lifespan in the CAO+BCEL is recorded in Figure 6.4. In both winter and summer, 35-40% of all cyclones that enter the CAO+BCEL originate locally (including the “arctic cyclones” described by Tanaka et al. (2012)). About the same number of cyclones migrate into the CAO+BCEL from elsewhere in each season, but the source regions change substantially. The North American, Pacific, and Atlantic regions all contribute more cyclones in winter. By contrast, the Eurasian region contributes many more cyclones in summer and accounts for more than half of all externally sourced summer cyclones in the CAO+BCEL.

These results complement Figure 6.2, which shows substantially more cyclogenesis over the Eurasian continent in summer than winter, especially around the Kolyma Lowland and the Putorana Plateau (PP in Figure 1.1). Focusing on the former, Figure 6.5 shows the (a,b) track density and (c,d) individual tracks for all cyclones originating in region B of Figure 6.4 for (a,c) winter and (b,d) summer. Few cyclones form here in winter, and on average only one or two cyclones migrate into the CAO+BCEL each winter. By contrast, the average summer
includes five to seven storms that form in this region and migrate into the CAO+BCEL and another one to three that track eastward into the Pacific Ocean. The greatest track density in summer is in the Kolyma Lowland, which lies to the north and east of the Siberian mountains.

Figure 6.4. Distribution of cyclones entering the CAO+BCEL for winter (DJF) and summer (JJA) 1979-2014. (a) Map showing 14 source regions (A through N) for CAO+BCEL cyclones. (b) Table showing the quantity and percentage of cyclones originating in each region.
The algorithm points to central Siberia (Figure 6.6a, or Region C in Figure 6.4), especially around the Putorana Plateau, as another summer source region for the CAO+BCEL. Cyclogenesis in this region is less pronounced and more diffuse than in eastern Siberia. Cyclones track from central Siberia in several directions, although they again most often move north and/or eastward. One particularly populous track is across the Lena River Delta and over the Laptev Sea.

While Siberia is a significant source of summer cyclones for the CAO+BCEL, the prominent genesis region east of the Mackenzie Range in Canada is only a minor source. Whereas about a third of the detected cyclones forming over the Siberian regions eventually travel into the CAO+BCEL, only 13% of those forming in the northwest
corner of North America (Region L in Figure 6.4) ever do. This region also lacks the seasonal contrasts observed in Siberia; it contributed exactly 80 cyclones to the CAO+BCEL during the study period in both winter and summer. As seen in Figure 6.6b, this is largely because a majority of Mackenzie cyclones track southeastward across the Canadian Shield.

One final observation from Figures 6.5 and 6.6a is that all of the cyclones that migrate from Siberia to the Arctic Ocean pass over the AFZ. Indeed, to reach the CAO+BCEL, nearly all of the 522 systems generated over Asia, and many of those generated over the Pacific Ocean, western North America, and Europe, pass through the AFZ. In total, 764 cyclones over 36 years make the passage, which accounts for the majority of cyclones that migrate into

Figure 6.6. (a,b) Contour plots and (c,d) spaghetti plots of cyclone tracks originating (a,c) in central Siberia (Region C in Figure 6.4) and (b,d) northwest North America (Region L) during summer (JJA) 1979-2014.
the CAO+BCEL. This presents a new possibility: because of its strong baroclinicity, the AFZ may be an area where pre-existing cyclones experience additional deepening. This idea was explored using multiple statistical analyses whose results are presented below.

6.2.3. Track-by-Track Analysis of Arctic Cyclone Deepening/Filling Rate

Of the 764 summer cyclones that crossed the Arctic coastline on their way to the CAO+BCEL from 1979-2014, 540 of them spent at least one observation time in a coastal grid cell. For winter, the counts are 539 and 315, respectively. The AFZ is only present in summer, so if it acts as a cyclone intensifier, then cyclones crossing the coastline in summer should experience more deepening (or less filling) than those crossing in winter. On average, summer cyclones experience a weak deepening rate of -0.4 hPa day$^{-1}$ when crossing the coastline, whereas winter cyclones experience a filling rate of +6.7 hPa day$^{-1}$. The difference is significantly different from zero with a 99% confidence interval using a two-sample Wilcoxon rank test. Because of different source regions, winter cyclones in this comparison tend to be older than summer cyclones. However, the difference remains significant at the 99% confidence level after controlling for the relative age at crossing. Therefore, this test is preliminary evidence that cyclones are likely to fill less rapidly (or even deepen) during coastline passage when the AFZ is present.

If the baroclinicity of the AFZ is truly the cause of this relationship, then day-to-day variations in AFZ strength should also influence the deepening rate of cyclones passing through it. The subset of 540 summer cyclones was examined more closely using the regression analysis described in Section 5.8.1. The null hypothesis is that, after controlling for other cyclone life cycle characteristics and time of day, AFZ strength at the given level will have no impact on the deepening rate of cyclones as they cross the Arctic Ocean coastline. This hypothesis was tested examining the AFZ at both 700 hPa and 500 hPa.

The estimated coefficients for each regression model are presented in Table 6.1. Note that since deepening occurs when the central pressure drops, a negative coefficient means that an increase in the right-hand

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22 The number for winter may seem high based on Figure 6.5. Some cyclones originating in Europe and the North Atlantic travel eastward over Russia first and then cross the AFZ on their way to the CAO+BCEL region, and these are much more common in winter.
variable is associated with greater deepening (or less filling). Conversely, a positive coefficient indicates that an increase in the right-hand variable is associated with less deepening (or greater filling). As expected, a positive relationship exists between the relative age of a cyclone and its deepening rate, meaning that cyclones have a tendency to deepen less vigorously as they age. After reaching its minimum pressure, a cyclone will begin to fill, so if the minimum is experienced before the cyclone crosses the AFZ, the cyclone is more likely to experience filling during the crossing. On the other hand, if the minimum is experienced after the cyclone crosses the AFZ, the cyclone is more likely to experience deepening during the crossing. This yields a negative relationship: The later the occurrence of minimum pressure, the stronger the deepening rate should be during AFZ crossing. These two relationships are significant regardless which vertical level is used to measure AFZ strength.

Also significant is the local time of day. The positive coefficient indicates that a cyclone’s central pressure is expected to fall from about 2100 to 0900 and rise from 0900 to 2100, similar to findings by Dai and Wang (1999) that the diurnal tide over Siberia in summer peaks in the mid morning.

Table 6.1. Coefficients (± 1.96 standard errors) for multiple linear regression models (Equation 5.1) describing the deepening rate of cyclones as they pass through the AFZ during June, July, and August. AFZ strength was measured at both 700 hPa and 500 hPa, so two models were constructed, one for each measure of AFZ strength. Significance at p < 0.05 is denoted using bold text, and at p < 0.01 using bold italics.

<table>
<thead>
<tr>
<th>Coefficient</th>
<th>Description</th>
<th>Units</th>
<th>700 hPa</th>
<th>500 hPa</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\beta_1$</td>
<td>Relative Age at AFZ Crossing</td>
<td>--</td>
<td>$+17.50 \pm 2.91$</td>
<td>$+17.63 \pm 2.89$</td>
</tr>
<tr>
<td>$\beta_2$</td>
<td>Relative Age at Minimum Pressure</td>
<td>--</td>
<td>$-11.19 \pm 2.83$</td>
<td>$-11.49 \pm 2.81$</td>
</tr>
<tr>
<td>$\beta_3$</td>
<td>AFZ Strength at AFZ Crossing</td>
<td>K (100 km)$^{-1}$</td>
<td>$-2.10 \pm 1.61$</td>
<td>$-2.30 \pm 1.70$</td>
</tr>
<tr>
<td>$\beta_4$</td>
<td>Diurnal Oscillation of Thermal</td>
<td>hours</td>
<td>$+1.16 \pm 0.99$</td>
<td>$+1.10 \pm 1.00$</td>
</tr>
<tr>
<td>$\alpha$</td>
<td>Intercept</td>
<td>hPa day$^{-1}$</td>
<td>$+1.14 \pm 2.76$</td>
<td>$+1.27 \pm 2.77$</td>
</tr>
</tbody>
</table>

Finally, in both models, the null hypothesis that the AFZ strength has no impact on cyclone deepening rates after controlling for cyclone life stage and time of day can be rejected. The variable for AFZ strength has a significant negative coefficient, meaning that stronger temperature gradients at mid-tropospheric levels are associated with less filling (or greater deepening) of cyclones crossing the coastline. The AFZ strength variable also remains significant for both 700 hPa and 500 hPa if any one, any two, or all other right-hand variables are removed from the model.
These models explain a small but significant 29.4% (for 700 hPa) or 29.6% (for 500 hPa) of the variance in cyclone deepening rate. Multicollinearity amongst right-hand variables is minimal, with the highest variance inflation factor under 1.04 for each model. The residuals in both models follow closely with a normal distribution, and no single observation was a major outlier. However, in the 700 hPa model, there may be some concern for heteroskedasticity in residuals (see Figure 6A.2).

6.2.4. Seasonal Analysis of Arctic Cyclone Intensity

If, as the previous results suggest, the AFZ at mid-tropospheric levels is a cyclone intensifier, then it might also be expected to exhibit some relationship with aggregate cyclone characteristics in the Arctic Ocean for each summer season. Table 6.2 shows the correlation between AFZ strength (at either 700 hPa or 500 hPa) and average cyclone intensity (central SLP, its Laplacian, and CAP) and frequency in the CAO+BCEL. It also shows the correlation between AFZ strength and EGR at 500 hPa, cyclone frequency, and cyclogenesis throughout the AFZ. The two right-hand columns show the same correlations, only using MAF strength instead of AFZ strength.

Consistent with Equations 3.1 and 3.3, a strong positive relationship is observed between AFZ strength at 500 hPa and EGR at 500 hPa. The correlation is weaker but also positive for the AFZ at 700 hPa. These results complement Figure 6.1, which shows the co-location of the AFZ and high EGR along the Siberian coast. Very similar results are obtained using MAF strength instead of AFZ strength.

Cyclone intensity is measured directly using central SLP and its Laplacian and indirectly using CAP. Central SLP only yields a significant correlation at 700 hPa for AFZ strength and shows no relationship with MAF strength. The Laplacian of central pressure and CAP show significant positive correlations with both AFZ strength and MAF strength, especially at 700 hPa. All significant correlations indicate that when the AFZ is stronger at mid-tropospheric levels, cyclones in the Arctic Ocean are more intense than normal. This is consistent with the idea presented in the previous section: When the AFZ is stronger at mid-tropospheric levels, cyclones passing through it experience more strengthening (or less weakening). This relationship is more strongly expressed at 700 hPa than 500 hPa.
Table 6.2. Spearman’s correlations between AFZ strength and average cyclone central pressure, Laplacian of cyclone central pressure, CAP for cyclone centers, and cyclone frequency in the CAO+BCEL, and EGR at 500 hPa, cyclone frequency, and cyclogenesis frequency along the AFZ for all summers (JJA) 1979-2014. Correlations were conducted using AFZ strength and MAF strength at 700 hPa and 500 hPa. Significance at p < 0.05 is denoted using **bold text** and at p < 0.01 using **bold italics**.

<table>
<thead>
<tr>
<th>Vertical Level</th>
<th>AFZ Strength</th>
<th>MAF Strength</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>700 hPa</td>
<td>500 hPa</td>
</tr>
<tr>
<td>EGR at 500 hPa</td>
<td>+0.46</td>
<td>+0.63</td>
</tr>
<tr>
<td>Laplacian of Central SLP</td>
<td>+0.53</td>
<td>+0.31</td>
</tr>
<tr>
<td>CAP</td>
<td>+0.56</td>
<td>+0.42</td>
</tr>
<tr>
<td>Central SLP</td>
<td>-0.37</td>
<td>-0.32</td>
</tr>
<tr>
<td>AFZ Track Frequency</td>
<td>+0.14</td>
<td>+0.16</td>
</tr>
<tr>
<td>CAO+BCEL Track Frequency</td>
<td>+0.26</td>
<td>+0.18</td>
</tr>
<tr>
<td>AFZ Cyclogenesis</td>
<td>+0.05</td>
<td>+0.15</td>
</tr>
</tbody>
</table>

Cyclone frequency is also measured in three ways, including cyclogenesis within the AFZ (along the coastline) and cyclone frequency (the number of unique tracks) both crossing the AFZ and within the CAO+BCEL. All twelve correlation coefficients are positive, and a few are significant at the 90% confidence level, which provides a whisper of evidence that a weak relationship may be confirmed if more data were included, but no significant relationship exists at the 95% confidence level for any measure of cyclone frequency with any measure of the AFZ.

Table 6.3. Same as Table 6.2 but with the effect of SVNAM removed from each variable using a linear model. Significance at p < 0.05 is denoted using **bold text** and at p < 0.01 using **bold italics**.

<table>
<thead>
<tr>
<th></th>
<th>AFZ Strength</th>
<th>MAF Strength</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>700 hPa</td>
<td>500 hPa</td>
</tr>
<tr>
<td>EGR at 500 hPa</td>
<td>+0.38</td>
<td>+0.60</td>
</tr>
<tr>
<td>Laplacian of Central SLP</td>
<td>+0.40</td>
<td>+0.18</td>
</tr>
<tr>
<td>CAP</td>
<td>+0.42</td>
<td>+0.34</td>
</tr>
<tr>
<td>Central SLP</td>
<td>-0.26</td>
<td>-0.28</td>
</tr>
<tr>
<td>AFZ Track Frequency</td>
<td>+0.08</td>
<td>+0.14</td>
</tr>
<tr>
<td>CAO+BCEL Track Frequency</td>
<td>+0.21</td>
<td>+0.16</td>
</tr>
<tr>
<td>AFZ Cyclogenesis</td>
<td>+0.08</td>
<td>+0.15</td>
</tr>
</tbody>
</table>

A caveat of these correlations is that they do not take into account potential confounding factors like large-scale circulation patterns. Particular concern lies with the seasonally varying Northern Annular Mode (SVNAM). Ogi et al. (2004) note a positive correlation between the SVNAM index and AFZ strength, and Serreze and Barrett (2008) describe a positive correlation between the SVNAM index and cyclone frequency in the CAO. Since large-scale circulation may influence both the AFZ and cyclone activity, correlations were also computed.
after subtracting the linear relationship with the SVNAM index for each variable. Results (Table 6.3) show that the correlations between AFZ strength and MAF strength with EGR, CAP, and the Laplacian of central pressure are all robust to this modification. Results for cyclone frequency are also the same, showing no significant correlations. The only substantial difference between Table 6.2 and 6.3 is that central SLP no longer shows a significant correlation for any test after adjusting for SVNAM.

6.3. Discussion & Limitations

6.3.1. General Cyclone Characteristics

Conclusions from this study rest in part on the validity of the cyclone detection and tracking algorithm. The general results from this updated algorithm (Figure 6.2) compare well to other published algorithms (see Section 5A.4). Consistent with prior studies (e.g., Serreze 1995; Wernli and Schwierz 2006; Simmonds et al. 2008), cyclone frequency is greater in winter (Figure 6.2a) than summer (Figure 6.2b) on the Atlantic side of the Arctic and Gulf of Alaska. By contrast, activity over the CAO and eastern Siberia is notably higher in the summer season.

Patterns of cyclogenesis (Figure 6.2c-d) and cyclolysis (Figure 6.2e-f) are also similar to those identified in prior studies (e.g., Trigo 2005; Wernli and Schwierz 2006; Simmonds et al. 2008; Hanley and Caballero 2012). Decreased cyclogenesis in summer in the Icelandic Low region, the Norwegian Sea, and the Barents Sea was also noted by Whittaker and Horn (1984), Serreze (1995) and Simmonds et al. (2008). While the polar front weakens and its attendant cyclone activity slackens in summer, cyclogenesis over the continents increases. In part, this may be linked to radiative and turbulent heat fluxes from the warmer continent heating the overlying atmosphere, reducing stability (Serreze et al. 1992), but the areas of greatest continental cyclogenesis in Figure 6.2 appear in the lee of mountain ranges. Lastly, the role of the CAO+BCEL as a cyclone collection area, experiencing much lysis but little genesis, has also been noted in several studies (e.g., Reed and Kunkel 1960; Whittaker and Horn 1984; Serreze and Barrett 2008).

A feature in Figure 6.2 not shared by all algorithms is frequent winter cyclogenesis in the Gulf of Alaska. The Gulf of Alaska lies at the end of the North Pacific storm track (Sinclair 1997; Gulev et al. 2001; Hoskins and Hodges 2002), so it might be expected to be dominated by cyclolysis. Indeed, even though cyclogenesis is also
common according to the present algorithm, cyclolysis occurs more frequently in the Gulf of Alaska than
cyclogenesis (Figure 6.3). Studies using SLP or the Laplacian of SLP (e.g., Gulev et al. 2001; Pinto et al. 2005; Wernli
and Schwierz 2006) similarly show some cyclogenesis but more cyclolysis in this area, but some studies using low-
level vorticity (e.g., Sinclair 1997; Hoskins and Hodges 2002) show only cyclolysis. Gulev et al. (2001) commented
on the difficulty of tracking systems that stagnate when entering this area. In the updated algorithm used here,
stagnation of mature systems is often accompanied by splitting events, when a cyclone, most often a MCC, divides
into two separate systems before its demise. These events account for more than half of the occurrences of winter
cyclogenesis in the Gulf of Alaska and may not be readily captured in algorithms without detection of MCCs.

Another issue is the high track density and frequent summer cyclogenesis in the Kolyma Lowland, to the
north and east of the Siberian mountains. This feature has been noted in many past studies using multiple
methods (Whittaker and Horn 1984; Serreze 1995; Serreze et al. 2001; Sorteberg and Kvingedal 2006; Wernli and
Schwierz 2006; Raible et al. 2008; Simmonds et al. 2008), although Mesquita et al. (2008) did not find such a
feature. Cyclones forming north and east of the Siberian mountains are likely the result of lee cyclogenesis. The
Verkhoyansk Range especially is often perpendicular to air flow on the downwind side of the Urals trough, placing
the Kolyma Lowland in an area of frequent divergence aloft. This was noted by Serreze and Barrett (2008) and is
also apparent in Figure 6.7, which shows the mean summer 500 hPa GPH field, as well as the mean field when
summer cyclogenesis is occurring north and east of the Siberian mountains. The trough in 500 hPa GPH over
central Siberia is normally broad and weak, but airflow is generally west to east across the Verkhoyansk Range.
When cyclogenesis occurs, the trough is tighter and stronger, shifted eastward, and includes a south-to-north
component in the contours. That Region B in Figure 6.4 generates the most summer cyclones in the Asian sector
despite having the smallest area may reflect the combined impacts of these mountain ranges lying across a
favorable area of mid-level flow.

As for cyclogenesis in the AFZ, the results presented here run counter to Serreze et al. (2001). Because they were using data with a coarser spatial resolution, Serreze et al. (2001) could not clearly distinguish between
cyclogenesis along the AFZ and cyclogenesis farther inland, such as in the Kolyma Lowland (in the lee of the
Siberian mountains). Past research has suggested that cyclone detection using vorticity may lead to earlier
detection of some cyclones (Mesquita et al. 2009; Rudeva et al. 2014); however, since the relevant summer
cyclogenesis maxima identified here lie to the south of the AFZ and their cyclones then track northeast, any shift in cyclogenesis regions resulting from the implementation of a vorticity-based method would likely move the cyclogenesis even farther from the AFZ.

Figure 6.7. (a) Mean summer GPH at 500 hPa (m) 1979-2014 and (b) mean GPH at 500 hPa (m) for all instances of summer cyclogenesis in the Kolyma Lowland.

6.3.2. The AFZ as Cyclone Intensifier

The statistical tests described in Section 6.1 indicate that the AFZ acts as a modest intensifier of storms crossing the Arctic Ocean coastline. A stronger AFZ is associated with stronger storms that contribute more precipitation to the Arctic Ocean. These results confirm that the AFZ is not only a notable factor for Arctic cyclones, but also for the Arctic hydrological system, which is consistent with the findings of Serreze et al. (2001) and Sorteberg and Walsh (2008). However, it is difficult to determine just how important the AFZ really is. The greatest source of ambiguity is arguably the relationship between the AFZ and large-scale circulation. Past research has reported positive correlations between the SVNAM index and both AFZ strength Ogi et al. (2004) and cyclone activity in the CAO (Serreze and Barrett 2008). Therefore, large-scale circulation may influence both the AFZ and cyclone activity. However, the cause and effect may also be reversed. For example, Ogi et al. (2004) concluded that a stronger AFZ could favor the positive mode of the SVNAM.

Additionally, results from Serreze et al. (2001), Ogi et al. (2004), and Crawford and Serreze (2015) all show that stronger zonal winds aloft (e.g., at 300 hPa) are related to a strong AFZ, so the formation of a strong
circumpolar vortex in positive SVNAM years may be encouraged by the presence of a strong AFZ. Based on Serreze and Barrett (2008), this likely means that the AFZ also has an indirect connection to cyclone development through its influence on large-scale circulation. If the AFZ can be a considered a forcing on large-scale circulation, then controlling for the SVNAM index in correlations may not be appropriate when assessing the link between the AFZ and cyclone activity.

The murkiness in this cause and effect may also make declarations that the AFZ has no role in cyclogenesis, only cyclone intensification, premature. Consider, for instance, that the lee cyclogenesis in the Kolyma Lowland occurs in concert with troughs in 500 hPa GPH. These troughs are part of the Arctic jet-like feature associated with the summer AFZ. If land-sea contrasts are responsible for the development of the summer AFZ, which in turn inspires the development of an Arctic jet-like feature, then those same land-sea contrasts indirectly impact cyclogenesis in the lee of the Kolyma Lowland. In other words, although the data presented here show no sign that the AFZ is a direct cyclone generator, it may still have some indirect influence on cyclogenesis.

Similarly, some ambiguity exists in the regression models built to compare AFZ strength and the deepening rate of individual systems. The AFZ is not responsible for generating the systems that it affects, and any cyclone that passes through the AFZ will have a cycle of deepening and filling independent of the AFZ. For instance, typical of extratropical cyclones, the storms generated in the lee of the Siberian mountains are associated with divergence aloft downstream of the trough axis in the 500 hPa GPH contours. Therefore, the AFZ is a secondary forcing operating in addition to the general life cycle of these storms. Accordingly, the most important predictors for cyclone deepening rate are cyclone age and the age at which minimum pressure is reached. These variables implicitly link back to the long wave troughs that drive cyclone development. Additionally, the variance in deepening rate explained by the regression models is modest (around 30%), as is the correlation between AFZ strength and cyclone intensity ($r = 0.51$ for the Laplacian of central pressure and the AFZ at 700 hPa, and $r = 0.41$ after controlling for the SVNAM index).

Examination of the AFZ-cyclone relationships could be improved in several ways, such as explicitly including upper-level divergence rather than implicitly including it by using cyclone life cycle measures. Using a broader coastal width to define the AFZ might allow a larger population of storms to be examined. The relationship may also be explored using the regional or cyclone scale Lorenz energy box paradigm (Pezza et al. 2010).
6.3.3. Use of Arctic Frontal Zone versus Migrating Arctic Front

Another caution with these results is that although historically the AFZ has been most often discussed in terms of its surface expression, that surface expression is only relevant to cyclone development insofar as thermal contrasts between land and ocean surfaces are translated to higher levels of the troposphere. Although evidence like Figure 6.2 demonstrates that such mechanisms exist and that the AFZ stretches vertically from the surface to the tropopause, the AFZ is less confined by surface processes with increasing height in the atmosphere. Therefore, in addition to measuring AFZ strength as temperature gradient magnitude at fixed locations along the Arctic Ocean coastline, an alternative measure, MAF strength, is also considered in seasonal correlations.

As might be expected, AFZ strength and MAF strength are positively correlated with each other (Table 6.4). However, considering that each metric is attempting to measure the same thing (the intensity of the AFZ), the correlations of +0.59 at 700 hPa and +0.49 at 500 hPa are modest. The two metrics are far from identical, then, which justifies examining both in the context of cyclone development.

Table 6.4. Spearman’s correlation coefficients between AFZ strength and MAF strength for all summers (JJA) 1979-2014 calculated at 700 hPa and 500 hPa both with and without an adjustment for the SVNAM. Significance at p < 0.05 is denoted using bold text and at p < 0.01 using bold italics.

<table>
<thead>
<tr>
<th></th>
<th>700 hPa</th>
<th>500 hPa</th>
</tr>
</thead>
<tbody>
<tr>
<td>Basic Correlation</td>
<td>+0.59</td>
<td>+0.49</td>
</tr>
<tr>
<td>SVNAM-Adjusted Correlation</td>
<td>+0.51</td>
<td>+0.51</td>
</tr>
</tbody>
</table>

The correlations examined here are largely robust to the choice of AFZ metric. At 700 hPa, the only difference is that, before adjusting for the SVNAM, AFZ strength yields a significant correlation with central SLP whereas MAF does not. At 500 hPa, the only difference is that the Laplacian of central pressure is only significant for MAF. In no case is there a difference in the sign of a coefficient (significant or not). The fact that the correlations described here are robust to changes in measuring the AFZ is further evidence they reflect a meaningful physical relationship between the variability in the AFZ and variability in Arctic cyclone development.

6.4. Conclusions

Past studies have proposed a relationship between development of the summer AFZ along the Arctic coastline and the summer maximum in cyclone activity over the central Arctic Ocean (Dzerdzevskii 1945; Reed
and Kunkel 1960; Serreze et al. 2001), but the present study is the first to rigorously address this issue. High EGRs along the coast indicate that the AFZ has the potential to affect cyclone development, but contrary to previous arguments, the summer AFZ is actually not a direct generator of cyclones. Instead, it acts as an intensifier. Many of the cyclones generated over Eurasia travel northeastward towards the Arctic Ocean and cross the AFZ along the way. When the AFZ is strong at mid-tropospheric levels, cyclones passing through it experience less filling (or more deepening). At the seasonal scale, a stronger AFZ at 700 hPa is also associated with higher EGR, greater cyclone intensity, and more cyclone-associated precipitation in the CAO+BCEL.

The connection between the AFZ and cyclone development is hence more nuanced than envisioned by previous studies. This is particularly true of eastern Siberia, where several geological coincidences converge to promote cyclone development, including an L-shaped set of mountain ranges, apparently ideally oriented to intersect the frequent troughs east of the Urals, and a zonally oriented coastline lying just to the north that fosters baroclinicity. However, the confluence of topographic features with circulation patterns (as represented by the SVNAM index), along with the capacity for cyclone activity to in turn influence the surface expression of the AFZ, makes observational data alone insufficient to separate cause and effect. (Reed and Kunkel came to a similar conclusion back in 1960.) A planned next step is to manipulate boundary conditions, such as topography and surface cover, in CESM and then conduct several model runs under each perturbed state. These sensitivity studies would help to better isolate the various environmental factors leading to AFZ development and its relationship to summer cyclone activity.
CHAPTER 7. THE AFZ IN A WARMING WORLD

7.1. Introduction

To date, most work regarding the AFZ has used observational data to describe its present state. However, the Arctic is currently undergoing rapid changes, including rising temperatures (Serreze and Barry 2011), shrinking sea ice cover (Serreze and Stroeve 2008; Stroeve et al. 2012), earlier snow melt (Armstrong and Brodzik 2001; Brown and Robinson 2011; Derksen and Brown 2012), degradation of permafrost (Jorgenson et al. 2006; Lawrence et al. 2008; Romanovsky et al. 2010), and accelerated vegetation growth (Euskirchen and McGuire 2006; Kimball et al. 2007; Zhang et al. 2013a). Moreover, these observed trends are projected to continue throughout the 21st century as greenhouse gas emissions persist in driving more global warming (Stroeve et al. 2012; Collins et al. 2013; Zhang et al. 2013b). Therefore, especially since the timing of snow and sea ice retreat each summer seems to impact the development of the AFZ (Crawford and Serreze 2015), the basic characteristics of the AFZ, such as its seasonal cycle, its peak intensity, and its spatial variability, may also change.

This chapter evaluates how the summer AFZ responds to climate change in RCP8.5, a strong global warming scenario, using CESM-LE. First, the depiction of the AFZ in CESM-LE is compared to that of atmospheric reanalyses, validating CESM-LE as a study tool for the AFZ (Section 7.2). Next, the global context of RCP8.5 is presented in Section 7.3. The response of the AFZ to this warming is assessed by comparing its characteristics during 2071-2080 in RCP8.5 to 1990-2005 (Section 7.4). Finally, the response of the AFZ to RCP8.5 is compared to observed trends (1979-2014) in those same AFZ parameters from atmospheric reanalyses (Section 7.5).

7.2. The AFZ in CESM-LE

7.2.1. Justification for Using CESM-LE

A common source of projections for future climate scenarios is Phase 5 of the Climate Model Intercomparison (CMIP5). CMIP5 is a large collection of model simulations from dozens of modeling centers throughout the world. Each modeling center used a standard set of external forcing parameters, time periods, and output variables to better evaluate a) the veracity of model hindcasts, b) projections for the near term (out to
about 2035) and long term (out to 2100 and beyond), and c) the factors responsible for model-to-model differences. CMIP5 was the basis for climate projections reported by the Intergovernmental Panel on Climate Change (IPCC) in its Fifth Assessment Report (AR5; IPCC 2013).

CMIP5, and climate models in general, are incredibly useful for studying the climate for periods beyond the observational record. However, one of the drawbacks of CMIP5 is that each modeling center only submitted one or a few model runs under each scenario. Considering only one (or a few) run(s) per model obscures unforced internal variability in the climate system, which can be the dominant source of uncertainty in climate model projections on decadal time scales, especially for regional projections (Kirtman et al. 2013). This is of particular concern for this study because past research has demonstrated that the intensity of the summer AFZ exhibits substantial interannual and regional variability (Crawford and Serreze 2015). To understand how a particular climate model expresses the internal variability of a regional feature like the AFZ, it is preferable to run the same model for the same time period with the same external forcings multiple times, with only small round-off differences to initial conditions differentiating each run (Kay et al. 2015). Such an approach has been undertaken using the Community Earth System Model (CESM), resulting in a “large ensemble” (CESM-LE) of model runs (ibid.). Variation amongst the ensemble members represents the stochastic processes of internal climate variability, as described by Lorenz (1963) for the impact of initial conditions on numerical weather prediction models.

When evaluating the potential impacts of climate change on the AFZ, the variability of chief interest is that due to external forcings. As described in Chapter 5, the RCP8.5 scenario (which involves about 8 W m⁻² of net radiative forcing by 2100) is used to simulate these forcings. To determine the forced change, the characteristics of the AFZ in the period 2071-2080 is compared to a baseline of 1990-2005 using CESM-LE. Internal variability in the climate system is accounted for by repeating all comparisons for each of 30 ensemble members. These results are presented in Section 7.4.

7.2.2. Validation of the AFZ in CESM-LE

Before that, however, another potential source of variability and uncertainty must be considered: bias in CESM compared to the atmospheric reanalyses. In order for CESM-LE to be a useful tool for projecting the response
of the AFZ to climate change, CESM must depict the AFZ realistically in the present. To that end, the Appendix to Chapter 7 compares in detail the horizontal extent (Section 7A.1), seasonality (Section 7A.2), vertical expression (Section 7A.3), spatial variability (Section 7A.4), and interannual variability (Section 7A.5) of the AFZ in CESM-LE to three atmospheric reanalyses (MERRA, ERA, and CFSR) for the period 1979-2005. The methods applied follow those of Crawford and Serreze (2015) except that they used a longer study period of 1979-2012. A shorter period is used here because starting in 2006, CESM-LE is forced by the RCP8.5 scenario. The results of that validation effort are summarized below.

In all aspects, CESM-LE shows a realistic AFZ that develops as strong horizontal cold-to-the-north temperature gradients each summer along the coastline of the Arctic Ocean, extends zonally across over 190° of longitude, and persists vertically up to the tropopause (Figure 7A.1-7). In cross section, CESM-LE accurately depicts the latitudinal and vertical extents of the AFZ, showing that it lies distinctly north of the polar front and, importantly, is associated with a jet-like feature aloft, in accordance with the thermal wind equation (Figure 7A.8-10). As in the reanalyses, the AFZ that develops in CESM-LE experiences substantial interannual variability in its strength, and that variability is manifested regionally (Figure 7A.17). Cluster analysis showed that the sectors developed by Crawford and Serreze (2015) based on reanalysis data are also applicable to the AFZ depicted by CESM-LE, and these sectors show similar seasonal and interannual variability in all datasets (Figure 7A.13-14).

However, a few examples of bias in CESM-LE compared to the reanalyses can be found. First, CESM-LE seems to have greater coupling between snow cover extent and atmospheric temperature gradients, showing a southerly bias in the position of both the strongest climatological summer temperature gradients at mid-tropospheric levels (Figure 7A.15) and the MAF (Figure 7A.16). This bias is notable for the AFZ-cyclone relationship because the best temperature-gradient-based predictor of cyclone intensification in the AFZ is at the 700 hPa level (Section 6.2.2).

Second, along the Taymyr Peninsula, lingering snow cover well into July retards AFZ development, leading to later emergence in CESM-LE than in the reanalyses (Figure 7A.3). This bias is most apparent at the surface, where AFZ development is most strongly determined by surface type. It has little impact on overall AFZ strength since the Taymyr Peninsula area has a weak and late-developing AFZ compared to other regions even in the
reanalyses. Additionally, in the sectors where the AFZ is strongest (eastern Siberia and western North American), CESM-LE exhibits strong agreement with the reanalyses.

Third, temperature gradients in CESM-LE are biased weaker in CESM-LE than in the reanalyses (Figure 7A.1-7). This likely results from a coarser spatial resolution. CFSR, which exhibits the strongest temperature gradients, has a resolution of 0.5° latitude and longitude. CESM-LE is about 1° of latitude and longitude. If two adjacent grid cells have the same temperature difference in both CESM-LE and CFSR, the temperature gradient calculated for CFSR will be greater (exactly how much greater being dependent on latitude and the aspect of the gradient). ERA and MERRA fall in between CFSR and CESM-LE for both spatial resolution and temperature gradient strength. ERA results are actually closer to those of CESM-LE than CFSR. In fact, CESM-LE often matches better with one of the reanalyses than the reanalyses do with each other for AFZ parameters. Therefore, for studying the AFZ, CESM-LE can be treated as a valid data set whose accuracy is on par with atmospheric reanalyses so long as the biases related to the surface cover are monitored. This justifies the use of CESM-LE to project the response of the AFZ to specific climate change scenarios.

### 7.3. RCP8.5: The Global Warming Context

Based on the full ensemble of CMIP5 models, including CESM, global mean annual near-surface air temperature is between 2.6°C and 4.8°C higher for the period 2081-2100 compared to 1986-2005 under RCP8.5 (Flato et al. 2013). For near-term projections (2013-2046), the range of temperature trends observed within CESM-LE is very similar to the range for CMIP5 (Kay et al. 2015), indicating that much of the near-term variability observed in CMIP5 can be attributed to internal variability of the climate system.

The current study compares 2071-2080 to 1990-2005. The relative importance of internal variability as a source of uncertainty is lesser for such long-term projections than in projections for 10 or 30 years into the future. Whereas internal variability is fairly similar in magnitude through time, model spread and scenario uncertainty both increase with time (see Figure 11.8 in Kirtman et al. (2013)).

However, since 2071-2080 represents only a single decade, natural climate oscillations like the Arctic Oscillation (AO) may exert a strong influence on the AFZ or cyclone characteristics, potentially skewing results. For
example, if the AO were coincidentally predominantly in its positive phase for this period for a particular ensemble member, then an increase in Arctic cyclone activity might be observed in comparison to 1990-2005 even if the forcing of global warming exerts no change. By examining 30 separate ensemble members, such anomalous decades are averaged out, leaving only the forced climate change signal in the ensemble mean.

RCP8.5 at 2071-2080 represents a transient response to increased greenhouse gas concentrations that exert a radiative forcing of about 7 W m$^{-2}$ by 2075. In CESM-LE, this results in an increase of about 3.27 ± 0.02 °C (Table 7.1) for global near-surface air temperatures during northern summer (JJA) between 1990-2005 and 2071-2080. Continuing RCP8.5 past 2100 would result in equilibrium radiative forcing of about 12 W m$^{-2}$ after 2200. Major emissions reductions, such as those pledged in relation to the Paris Agreement (COP21, 2015), would substantially reduce the radiative forcing experienced and thereby decrease the magnitude of the changes by the period 2071-2080. All projected changes to the AFZ must be considered in this context: they are a projection of one possible scenario, not a prediction of the future.

Table 7.1. Mean June-August temperature differences between 2071-2080 and 1990-2005 and 95% confidence interval based on 30 CESM-LE members. Differences are calculated at the near-surface (reference height) and 700 hPa levels for the entire world and the Arctic (latitude > 60°N).

<table>
<thead>
<tr>
<th>Period</th>
<th>Near-Surface</th>
<th>700 hPa</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>World</td>
<td>Arctic</td>
</tr>
<tr>
<td>June</td>
<td>3.24 ± 0.02</td>
<td>4.59 ± 0.08</td>
</tr>
<tr>
<td>July</td>
<td>3.27 ± 0.02</td>
<td>4.73 ± 0.07</td>
</tr>
<tr>
<td>August</td>
<td>3.30 ± 0.02</td>
<td>5.50 ± 0.08</td>
</tr>
<tr>
<td>Summer (JJA)</td>
<td>3.27 ± 0.02</td>
<td>4.94 ± 0.07</td>
</tr>
</tbody>
</table>

Table 7.1 also shows the temperature change for just the Arctic region (north of 60°N). Mean summer near-surface air temperature increases by 4.94 ± 0.07 °C, an increase more than 1.5°C greater than for the world as a whole. In June, this Arctic amplification of the warming signal is due primarily to enhanced warming over the Arctic land (Figure 7.1d). In July, the continental enhancement is restricted to the Taymyr Peninsula, the northern coast of Greenland, and the CAA (Figure 7.1e). Meanwhile, Hudson Bay and the Barents Sea also begin to show stronger-than-average warming. By August, the most notable areas of amplified warming are Greenland, the Greenland Sea, and the Arctic Ocean (Figure 7.1f). In other words, Arctic amplification transitions from being a primarily continental phenomenon to being a primarily oceanic phenomenon. Another notable observation is that, consistent with current trends (Screen and Simmonds 2010; Serreze and Barry 2011, Pithan and Mauritsen 2014),
Arctic amplification of warming is strongest near the surface, as shown by a) more similar warming trends between regions at 700 hPa in Table 7.1 and b) lesser spatial variation indicated for the 700 hPa level in Figure 7.1.

Arctic amplification of global warming has several causes, many of which are often overlooked. For instance, the simple fact that outgoing longwave radiation is proportional to temperature taken to the fourth power means that for the increase of 7 W m\(^{-2}\) of emitted radiation by 2075 found in RCP8.5, a blackbody starting at 30°C would need to warm by 0.40°C, less than a blackbody at 0°C.\(^{23}\) Even more important is the negative lapse rate feedback in the tropics, whereby a warmer atmosphere transports more latent heat aloft when warmed

\[^{23}\] Using the Stefan-Boltzmann Law: \(R = \varepsilon \sigma T^4\), where \(R\) is irradiance (W m\(^{-2}\)), \(\varepsilon\) is emissivity (1), \(\sigma\) is \(5.67 \times 10^{-8}\) W m\(^{-2}\) K\(^{-4}\), and \(T\) is temperature (in K).
(Pithan and Mauritsen 2014). When that latent heat is released in the upper troposphere, it compounds the sensible heat transport, making the upper troposphere warm more vigorously than the lower troposphere. This effectively reduces the moist adiabatic lapse rate, decreasing the amount of surface warming in the tropics necessary to balance the increase in downwelling longwave radiation. The Arctic is much drier than the tropics and rarely experiences deep convection. More common in the Arctic, especially in winter, are surface inversion layers that maintain high static stability and inhibit vertical mixing (Serreze et al. 1992; Desvathale et al. 2010). Therefore, warming at the surface is less easily transferred aloft in the Arctic, meaning that even more surface warming is required to strike the same top of the atmosphere radiative balance as in the tropics (Pithan and Mauristen 2014).

In summer, though, the most important amplifying feedback for Arctic warming is the albedo feedback (Pithan and Mauritsen 2014), whereby warming leads to a decrease in albedo, which in turn allows the surface to absorb more incoming shortwave radiation and warm even further. Most often, the albedo feedback is discussed in relation to sea ice loss (e.g., Screen and Simmonds 2010; Screen et al. 2013; Yim et al. 2016). However, loss of snow cover in late spring and early summer can also lead to a positive albedo feedback over land (Serreze and Barry 2011). The snow cover and sea ice loss under RCP8.5 is depicted in Figure 7.2. The top row shows that even in 1990-2005 for CESM-LE, snow cover is already absent in European Russia and most of Alaska in June but lingers across much of Siberia and along the Arctic Ocean coastline. By July, it remains only in the CAA, the Taymyr Peninsula, and Greenland. In 2071-2080, CESM-LE shows exceptional June snow cover declines over the Taymyr Peninsula and Chukotka, and the CAA. These are the same areas with especially high June temperature increases near the surface (Figure 7.1d). Snow cover declines are less in July and August because the initial snow cover is already so sparse in these months.

The sea ice changes are drastic, with June 2071-2080 looking most similar to August 1990-2005. Sea ice concentrations exceeding an average of 15% are virtually absent from the Arctic Ocean in August 2071-2080. Of any summer month, August experiences both the greatest decline in sea ice concentration and the greatest amplification of warming over the ocean. However, the lack of ocean-based amplification in June and July also stems from the ocean’s tendency to stores much of the energy it absorbs throughout summer. Although raising the ocean’s skin temperature can still increase the upward longwave flux in summer, large quantities of energy are released to the atmosphere only when the atmosphere cools down and the turbulent heat fluxes are directed
upward (Serreze and Barry 2011). This situation is most typical in autumn and winter, and Arctic amplification is even stronger in those seasons.

**Figure 7.2.** Mean June-August snow cover and sea ice concentration based on CESM-LE ensemble mean for (left) 1990-2005 and (center) 2071-2080 and (right) their difference. Stippling indicates a significant change based on the 95% confidence interval of a student’s t-test in which each member is considered an independent observation (30 observations per period).

In conclusion, Arctic summer warming is amplified compared to the global trend, especially at the surface. This amplification is greatest over the continents in June, reflecting the decline of spring snow cover. August
exhibits greater warming over the ocean following the transition from a predominantly sea ice surface to a predominantly open water surface. The next section examines how these asymmetrical warming patterns affect the seasonal development of the future AFZ.

7.4. The AFZ Under RCP8.5 (1990-2005 v. 2071-2080)

7.4.1. Surface Expression

The near-surface response of the AFZ to Arctic warming is closely tied to the difference in warming over the continents and ocean. In June, which exhibits major declines in snow cover, temperature gradient magnitude weakens across the continental interiors and strengthens along the coast (Figure 7.3, top). With the snow cover greatly diminished, coastal areas experience the greatest warming in the Northern Hemisphere. Despite some decline in sea ice concentration, air over the Arctic Ocean experiences a quarter of the warming seen over Siberia and the North Slope of Alaska. Although sea ice concentration is decreased, sea ice is still being melted, holding skin temperature at the melting point. In response, the coastline becomes an even greater focus of thermal contrast and the AFZ strengthens substantially (in some sectors more than doubling in strength).

Whereas snow cover decline clearly leads to AFZ strengthening in June, it has a lesser impact in July (Figure 7.3, middle). In the month of peak AFZ strength, the Taymyr Peninsula (Sector 3) shows strengthening, but the rest of the Eurasian coast appears robust to change. Along the North American Arctic coast, the AFZ actually experiences some weakening. As shown in Figure 7.2, sea ice decline is much greater in July and August, and the Arctic Ocean experiences much more open water in these months. Even though the sea ice-albedo feedback is most strongly expressed later in the year, some amplification of warming is still apparent over the Arctic Ocean in summer.

Warming over land and ocean have counteracting effects on the AFZ in July, but changes to the ocean dominate in August (Figure 7.1, right). With no more snow to remove, the main surface change is the loss of sea ice (Figure 7.2, bottom). This loss is expressed as amplified warming over the ocean compared to over land and weakening of the surface AFZ (Figure 7.3, bottom).
Figure 7.3. Mean June-August 2 m horizontal temperature gradient magnitude (K [100 km$^{-1}$]) based on CESM-LE ensemble mean for (left) 1990-2005 and (center) 2071-2080 and (right) their difference. Stippling indicates a significant change based on the 95% confidence interval of a student’s t-test in which each member is considered an independent observation (30 observations per period).

Although not focused on the AFZ specifically, Nishii et al. (2015) also note strengthening of near-surface meridional temperature gradients along the Eurasian coastline. Taking a multi-model ensemble mean of 17 CMIP5 models$^{24}$, they found an Arctic-wide increase of 0.15 to 0.20 K (100 km$^{-1}$) for summer (JJA) in RCP4.5 from 1981-

$^{24}$ CESM was not included.
1998 to 2081-2098. This increase is much weaker than observed in Figure 7.3 for June (top), but similar to the increase observed for July (middle). They also attributed this strengthening to greater warming occurring over the continents.

Figure 7.4. Seasonality of near-surface AFZ strength based on monthly means of nine AFZ sectors for both 1990-2005 (grey) and 2071-2080 (white) from 30 CESM-LE members. Observations total 480 for 1990-2005 and 300 for 2071-2080. Each box contains the median and first and third quartiles; the whiskers extend to the most extreme observation within 1.5 times the interquartile range (outliers beyond that point being omitted).

7.4.2. Seasonality at the Surface

As shown in Crawford and Serreze (2015) and in Section 7A.4, spatial and seasonal variability in near-surface AFZ development can be summarized by dividing the AFZ into nine sectors. Figure 7.4 compares boxplots of the monthly near-surface meridional temperature gradient for each AFZ sector between the periods 1990-2005 (480 observations each) and 2071-2080 (300 observations each). Broadly speaking, the seasonal cycle of AFZ development in spring and AFZ diminishment in autumn remains strong in all sectors. The transition months of
April/May and September continue to show neutral gradients on average. However, differences in the early summer and late autumn are apparent in all sectors.

More specifically, June temperature gradients become more negative for every sector, indicating earlier development of the AFZ accompanying earlier snow melt. This is especially notable in Sectors 2 and 3, where June temperature gradients from 1990-2005 are nearly neutral but become clearly negative for 2071-2080. Rising June AFZ strength is perhaps more impressive in the four sectors with the strongest AFZ (Sectors 5, 6, 8, and 9). For these areas, the peak strength of the AFZ also shifts earlier, with mean June temperature gradients matching or exceeding those of July for 2071-2080 whereas July is clearly the month of peak AFZ strength for 1990-2005. At the surface at least, the AFZ experiences both changes to intensity and timing.

Temperature gradients may become more negative in June, but most months showing change exhibit temperature gradients becoming more positive (or less negative). The greatest difference between the two periods is observed in November – February and likely represents the greater frequency of open coastal waters warming the oceanic atmosphere in autumn and winter to an even greater degree. Additionally, this is the time when more of the extra energy stored in the ocean column is released to the atmosphere. In other words, although they have a lesser impact on the summer AFZ, positive feedbacks to Arctic warming associated with sea ice loss are strongly manifested in CESM-LE in autumn and winter.

7.4.3. AFZ Response in Cross Section

As noted in Section 7.3., Arctic amplification of warming is experienced most strongly near the surface. At higher levels of the atmosphere, warming is more similar over land and ocean surfaces. Therefore, although Figures 7.3 and 7.4 provide evidence that the AFZ will experience more strengthening in June at the surface, the AFZ must also be examined at higher levels of the atmosphere.

**Figure 7.5** shows cross sections along four longitudes that depict the change in June meridional temperature gradients (top row) and zonal wind velocity (bottom row) from 1990-2005 to 2071-2080 under RCP8.5. Each cross section extends from 40°N to 90°N and has the Arctic Ocean coastline marked with “C”. Recall from Figure 7.2 that June exhibits the greatest decline in snow cover of any summer month and the smallest
decline in sea ice concentration for the BCEL seas. Along 205°E (154°W), 168°E, and 152°E, this leads to strengthening (red shading) of the AFZ, most prominently near the surface, but extending up to at least 600 hPa. These cross sections are representative of sectors 8-9, 7, and 5-6, respectively. They show that for a substantial part of the AFZ, June strengthening is experienced throughout the troposphere.

Figure 7.5. Latitudinal cross sections of the difference in June meridional temperature gradient (a-d) and zonal wind velocity (e-h) between 1990-2005 and 2071-2080 under RCP8.5 in CESM-LE. Stippling indicates a significant change based on the 95% confidence interval of a student’s t-test in which each member is considered an independent observation (30 observations per period).
The final cross section, along 50°E, crosses the Barents Sea and sector 1 of the AFZ. Even in 1990-2005, snow cover is usually melted away in sector 1 by June 1, although a little sea ice remains in the Barents Sea (Figure 7.2). Since no snow cover exists to lose, the only major surface change comes from the loss of June sea ice. Since the surface change is much different from the other longitudes, this cross section serves as a good test for whether the strengthening of the AFZ observed elsewhere is truly related to a change in surface type. Consistent with some sea ice loss and no change to the land surface, temperatures increase more over the Barents Sea than over European Russia, and the AFZ appears to weaken (blue shading) somewhat at 50°E.

These changes to the AFZ are translated to zonal velocity to varying degrees. Significant strengthening (blue shading) of the Arctic jet-like feature is observed at 205°E (154°W) and 168°E. No significant change is found along 152°E, where the AFZ strengthening was less consistent with increasing altitude, and at 50°E, the jet-like feature weakens (red shading).

Two other notes about these cross sections are worth mentioning. First, all four show weakening of the polar front and polar jet around 45°N, which agrees with the results of past studies that examine the impact of global warming on large-scale circulation features (e.g., Bengtsson et al. 2009; Collins et al. 2013; Ulbrich et al. 2013). Second, all show apparent strengthening of meridional temperature gradient and zonal wind velocity centered at about 200 hPa. These shifts are related to an increase in tropopause height as the atmosphere warms (Lorenz and DeWeaver 2009) and represent atmospheric changes on a larger scale than the focus of this study.

Looking along these same cross sections for July (Figure 7.6) shows that although some strengthening of temperature gradients is apparent near the surface, the AFZ experiences less change overall. The cross section through Alaska (206°E/154°W) is the only one to show strengthening of the AFZ or the westerlies throughout the troposphere. The absence of change in July follows logically from the surface response (Section 7.4.1). Snow cover loss is no longer a major factor in July, and ice-albedo feedback largely operates through ocean heat uptake in summer that is mostly released in fall and winter.

With no clear surface-based signal, the strengthening along the Alaskan coast is surprising. Such strengthening also exists in August (Figure 7.7), when all other cross sections show no significant changes to either the AFZ or its attendant jet-like feature even as significant changes are visible to both the north and south. This behavior may not be limited to CESM. Over 80% of the models considered by Nishii et al. (2015) showed
strengthening of summer westerlies at 850 hPa over the northwest North America coastline, and the multi-model ensemble mean change was $+0.4 \text{ m s}^{-1}$. Similar changes were not observed over the Eurasia coastline despite similar strengthening of the near-surface AFZ.

Figure 7.6. Same as Figure 7.5 except for July.
Surface changes in August are even less likely to encourage AFZ strengthening, so another factor is likely affecting the Alaskan coastline despite local surface changes in the Arctic. As noted above, these cross sections also depict the polar front and polar jet stream. Weakening of the westerlies is observed from 40°N to over 50°N in every cross section. However, in addition to weakening, the polar jet is also expected to shift poleward as the tropical Hadley Cell expands (Kang and Lu 2012; Collins et al. 2013). This shift is apparent, for instance, along 168°E
and 152°E in July (Figure 7.6b-c), where weakening of the polar front at 45°N is partnered with strengthening at 60°N. Over the Arctic Ocean coastline, around 70°N, little change is observed to meridional temperature gradients.

In addition to this proposed trend, observations from 1990-2005 show that the polar jet typically sits farther north as the summer progresses, making the distinction between Arctic and polar front clearest in June and least clear in August. Therefore, it is possible that the strengthening of westerlies and meridional temperature gradients observed along the coast of Alaska for August, and perhaps July, is actually caused by more frequent ridging of the polar jet and front over the North American cordillera, and more specifically the Brooks Range. In other words, regardless of what occurs locally in the Arctic, changes occurring at lower latitudes may also influence the behavior and distinctness of the AFZ. In late summer Alaska, the forcing by the surface energy balance may be overwhelmed by migration of the polar front.

7.4.4 Change to the Migrating Arctic Front

As discussed in Section 5.4.4, an alternative way to measure AFZ strength is to average the temperature gradient magnitude of the grid cells in the 90th percentile of the area bounded by 60°N to the south, 88°N to the north, 42°E to the west and 234°E (126°W) to the east. Doing this for the 700 hPa level and comparing 2071-2080 to 1990-2005 (Figure 7.8), CESM-LE shows a story that complements the findings in Section 7.4.3. In June (Figure 7.8a), the mean latitude of the MAF grid cells sits farther to the north, consistent with the loss of June snow cover and the strengthening of temperature gradients along the Arctic Ocean coastline.

In July and August, the opposite shift is observed; the mean latitude of the MAF actually decreases. As noted in Section 7A.2.4.2, CESM-LE is already biased toward a more southerly MAF, whereas the reanalyses more often show it shifting north of the coastline. In fact, the most common average latitude for the MAF in August is 69°N, a little south of the coastline (which has an average latitude of 71°N).

However, the cross sections in the previous section, along with monthly maps of temperature gradient magnitude, suggest that the perceived southerly shift is spurious. Rather, this result likely reflects the northward shift of the polar front, which more often pushes north of the 60°N boundary in the future, confounding efforts to measure the AFZ in this way. In other words, the MAF metric might only be appropriate for the current climate.
Figure 7.8. Mean latitude of the migrating Arctic Front (defined as grid cells within the area bounded by 42°E to the west, 234°E (126°W) to the east, 60°N to the south, and 88°N to the north that exceed the 90th percentile of horizontal temperature gradient strength at 700 hPa) for (a-c) the summer months and (d) the combined summer season for 1990-2005 (solid orange) and 2071-2080 (red outline) based on 30 CESM-LE members.

7.5. Observed Trends in the AFZ

7.5.1. Near-Surface Level Trends

Based on a 5-year moving average, Northern Hemisphere summer (JJA) temperatures rose by about 0.6-0.8°C (Hansen et al. 2010; GISTEMP Team 2010; NOAA 2017) between 1979 and 2014 (a trend of about 0.17-0.22°C per decade). Considering the small degree of warming and the presence of substantial interannual variability in the AFZ, finding a climate change signal in the AFZ may seem difficult even if it is changing.

However, as shown in Figure 7.9, the AFZ does exhibit some substantial changes near the surface. The graphs in Figure 7.9 show linear trends in AFZ strength at 2 m for June, July, and August of the period 1979-2014 for each AFZ sector using data from (a) MERRA and (b) ERA. Near the surface, AFZ strength has increased in several
sectors over this time period, especially in the month of June. Two sectors (7 and 8) show a decline in AFZ strength in July, but only in MERRA. Strengthening near the surface in June with less change in July is similar to the AFZ response under RCP8.5. Unlike RCP8.5 which showed weakening of the AFZ at the surface in August, no trends are found for 1979-2014. As in RCP8.5, these trends can be linked to the seasonally offset trends in snow and sea ice loss and the albedo feedbacks to warming induced by those trends. All sectors of the AFZ exhibit a decline in coastal (within 250 km of the coast) July snow cover, and all but sector 3 show a decline in June (Figure 7.10a). For many sectors, snow cover is often absent in August even in the early part of the record, so the few significant snow loss trends in August are small. Arctic amplification of warming over the continents is therefore strongest in June, as in RCP8.5. Coastal sea ice cover is on the decline in most sector-months as well (Figure 7.10b), but its albedo feedback is largely delayed by ocean heat uptake, as stated above.
Figure 7.11. Latitudinal cross sections of the trend in June (a-d) meridional temperature gradient and (e-h) zonal wind velocity over the period 1979-2014 in ERA. Stippling indicates a significant trend based on the 95% confidence interval (after removing 1-lag autoregression). Also marked is the Arctic Ocean coastline (C).

Figures 7.11, 7.12, and 7.13 show four longitudinal cross sections of trends in ERA for meridional temperature gradient and zonal wind velocity for June, July, and August, respectively. These longitudes are

25 Except where noted, MERRA cross sections show consistent results.
representative of four sectors in which the AFZ is especially strong: sector 4 (120°E), sector 5 (152°E), sector 6 (168°E), and sector 8 (206°E/154°W). In general, the AFZ has experienced less change over 1979-2014 than it does from 1990-2005 to 2071-2080 in RCP8.5. Most observed trends, especially in June, are restricted to below 925 hPa. In most cases, then, the trends at the surface are not yet strong enough to effect significant changes aloft.

**Figure 7.12.** Same as Figure 7.11, except for July.
Two exceptions exist to this. First, Alaska in August (Figure 7.13a,e) shows a prominent weakening trend for the AFZ and Arctic jet-like feature. (Trends are weaker but present in MERRA). The loss of sea ice in the Beaufort Sea may be implicated here, but given the discussion in Section 7.4.3, some influence of the polar front cannot be ruled out. (Note the strengthening of the polar front at 45°N in the same cross section.)

Second, eastern Siberia in July (Figure 7.12) shows some modest strengthening of the AFZ at mid-levels of the troposphere (strongest in MERRA) despite little change at the surface. The reasons for this are uncertain, but
because the strengthening aloft is decoupled from the surface, it is likely that some other influence is at work. Crawford and Serreze (2015) found that in addition to sea ice and snow cover retreat, the seasonal development of the AFZ is also strongly related to advection from onshore and offshore winds. Although they focused on the lower troposphere, advection is also likely to be important aloft. Additionally, unlike a large ensemble of model runs, ERA and MERRA express only one realization of the climate system for 1979-2014, so these trends are more subject to internal variability than the t-tests conducted with CESM-LE in Section 7.4.

In addition to examining average temperature gradient strength along the coastline, the AFZ was also measured using the intensity of the MAF at 700 hPa (Table 7.2). Consistent with the lack of widespread changes in Figures 7.11-13, no month has seen a significant trend in MAF strength in either reanalysis. No change is observed in mean latitude for any month in ERA, either, although MERRA does show a southerly shift for July. The sign for each of these trends matches the changes observed for RCP8.5. Overall, trends for the period 1979-2014 show less change than found in CESM-LE for RCP8.5 between 1990-2005 and 2071-2080.

Table 7.2. Linear trends in monthly and summer MAF strength (K [100 km]$^{-1}$ decade$^{-1}$) and mean latitude (°N decade$^{-1}$) for the period 1979-2014 according to MERRA and ERA. Bold values are significant at p < 0.05.

<table>
<thead>
<tr>
<th></th>
<th>MAF Strength</th>
<th></th>
<th>MAF Latitude</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>MERRA</td>
<td>ERA</td>
<td>MERRA</td>
</tr>
<tr>
<td>June</td>
<td>+0.01</td>
<td>+0.00</td>
<td>+0.24</td>
</tr>
<tr>
<td>July</td>
<td>+0.01</td>
<td>+0.00</td>
<td>-0.89</td>
</tr>
<tr>
<td>August</td>
<td>-0.02</td>
<td>-0.02</td>
<td>-0.24</td>
</tr>
<tr>
<td>JJA</td>
<td>+0.01</td>
<td>+0.00</td>
<td>-0.72</td>
</tr>
</tbody>
</table>

7.6. Conclusions

The Arctic is currently experiencing uneven warming at the surface. The loss of snow cover enhances continental warming in June, whereas enhanced oceanic warming begins to appear in August. This is leading to strengthening of coastal temperature gradients near the surface, but so far no substantial and widespread changes are visible aloft.

In the RCP8.5, warming continues throughout the 21st century, and snow cover and sea ice retreat from the coastal regions of the Arctic earlier and earlier each spring and summer. In this scenario, the AFZ experiences some strengthening in June, especially at the surface, but also aloft. Warming occurs throughout the Arctic, but
the loss of continental snow cover in June helps generate a robust AFZ a little earlier in the year. Little change is observed in July and August, although the AFZ may actually weaken in some places in August as the loss of sea ice reduces the thermal contrast between ocean and land. Poleward shifts in mid latitude features may, however, obscure somewhat the AFZ in Alaska later in summer.

Since RCP8.5 represents an aggressive emissions scenario, these projections may overestimate the amount of warming experienced in the Arctic. If that is the case, then even less change to the timing of AFZ development can be expected. Overall, though, the AFZ is fairly robust to warming temperatures, with significant changes being restricted both temporally and spatially.
CHAPTER 8. SUMMER ARCTIC CYCLONES IN A WARMING WORLD

8.1. Introduction

The previous chapter detailed how although a substantial fraction of the AFZ may experience some strengthening in June, changes to the AFZ under RCP8.5 are largely restricted to the surface in July and August. Therefore, the AFZ is likely to be a stabilizing force on summer Arctic cyclone activity in the future. However, the AFZ is not the only factor affecting Arctic cyclones. Most of the storms it interacts with are generated over the Eurasian continent, and a large fraction of storms that affect the CAO+BCEL domain are generated locally or over the North Atlantic, never interacting with the AFZ. So despite the constancy of the AFZ, summer cyclone activity in the Arctic may yet experience some changes in the future.

This chapter examines the response of summer Arctic cyclones to a warming world. First, Section 8.2 is a validation study that highlights a few biases in the treatment of cyclones by CESM despite general agreement in seasonal and spatial patterns of cyclone activity in CESM-LE, MERRA, and ERA. Next, the response of summer Arctic cyclone activity and its relationship with the AFZ under RCP8.5 are detailed in Section 8.3. Section 8.4 contains a brief description of the near-absence of trends in two atmospheric reanalyses (MERRA and ERA) for the period 1979-2014. Lastly, several limitations of CESM-LE, RCP8.5, and the methods used to analyze them are discussed in Section 8.5.

8.2. Validation of Arctic Cyclones in CESM-LE

Just as the Appendix to Chapter 7 was devoted to validating the AFZ in CESM-LE, so too does the Appendix to Chapter 8 serve as a validation study of cyclone activity in CESM-LE, with particular attention paid to summer Arctic cyclones and their relationship to the AFZ. A few important findings from that validation study are summarized here.

The cyclone database constructed from CESM-LE data produces the same overall patterns of cyclone frequency, genesis, and lysis as the databases based on ERA and MERRA (Figures 8A.3-5). It also has similar seasonal patterns. Track density is especially high along the North Atlantic and North Pacific storm tracks,
especially in winter. By contrast, track density over the continents and within the CAO is highest in summer. The source regions for cyclones entering the CAO+BCEL also share the same seasonal patterns. Local cyclogenesis is similar between winter and summer. The number of cyclones migrating into the CAO+BCEL from the Atlantic, Pacific, and North America is higher in winter than summer, but the declines in summer are countered by increased migration of cyclones from Eurasia.

However, CESM-LE does exhibit several biases. It contains noisier SLP fields that produce more SLP minima (Figure 8A.1-2), and it seems to overestimate summer track density in areas of complex topography (Figure 8A.3). On the other hand, summer track density is underestimated over many ocean basins, including the Arctic Ocean, which is a common bias in climate models (Zappa et al. 2013; Nishii et al. 2015). This bias is borne out in cyclogenesis patterns, too, as CESM-LE underestimates local cyclogenesis in the CAO+BCEL while overestimating migration of cyclones from Siberia and elsewhere (Figures 8A.6,9,10). Most notable for this study, CESM-LE identifies part of the AFZ as a cyclogenesis maximum for storms that track through the CAO+BCEL, whereas the reanalyses do not.

Some of these biases can be better understood by recalling the biases in the MAF noted in Section 7A.4.2 and Figures 8A.15 and 8A.16. The MAF is observed north of its mean position (along the coastline) less frequently in CESM-LE than in the reanalyses. Therefore, classic top-down cyclone development downwind of a trough may be less likely to occur over the Arctic Ocean in CESM-LE than in the reanalyses and more likely to occur over eastern Siberia. More specifically, though, cyclogenesis is biased along the AFZ. Looking more closely at the 500 hPa GPH contours during east Siberian cyclogenesis events (Figure 8A.7), the preferred alignment of troughing in the 500 hPa flow appears biased toward favoring cyclogenesis along the coastline in CESM-LE.

Despite these biases in its depiction of both Arctic cyclones, CESM-LE still shows a realistic relationship between the AFZ and cyclone intensity. When the AFZ is stronger, CAO+BCEL cyclones are more intense and produce more precipitation (Figures 8A.11 and 8A.14). Unlike the reanalyses, it also shows some evidence for AFZ strength influencing cyclogenesis, which likely follows from the bias in cyclogenesis along the east Siberian coastline. Therefore, CESM may assign excessive importance to the role of the AFZ in cyclone development, which must be remembered when examining the fate of Arctic cyclones under RCP8.5 in the next section. More
confidence, then, is assigned to results regarding the AFZ’s influence on cyclone intensity, and less confidence is assigned to results regarding its influence on cyclone frequency.

**Figure 8.1.** Cyclone track density for summer (JJA, top) and (bottom) winter (DJF, bottom) averaged from CESM-LE for the periods 1990-2005 (left) and 2071-2080 (center) and their difference (right). Stippling indicates a significant change based on the 95% confidence interval of a student’s t-test in which each member is considered an independent observation (30 observations per period).

### 8.3. Arctic Cyclones in RCP8.5 (1979-2005 v. 2071-2080)

#### 8.3.1. Overall Cyclone Activity

**Figure 8.1** shows the difference in average cyclone track density between 2071-2080 (using the RCP8.5 scenario) and the reference period of 1990-2005 for (top) summer (JJA) and (bottom) winter (DJF). With few exceptions, track density is unchanged or declines throughout the area 45-60°N in both seasons. In winter, the Icelandic Low region experiences a decline of over 4.8 storms per winter (between about 10 and 20%). The North Pacific storm track experiences less dramatic change, with declines in track density for winter restricted to the
western end around Kamchatka and Japan. The places that experience an increase in track density are generally at higher latitudes and include north of Hudson Bay, north of Ellesmere Island, the Sea of Okhotsk, and along the coasts of the Chukchi and East Siberian Seas.

Although winter is not the focus of the current study, it provides more opportunities to compare CESM-LE results to previous work. A decline in cyclone frequency throughout the mid-latitudes is a consistent finding in modeling studies that compare cyclone activity at the end of the 21st century to current conditions (Finnis et al. 2007; Pinto et al. 2007; Bengtsson et al. 2009; Ulbrich et al. 2013). The Arctic, meanwhile, exhibits either no change to cyclone frequency or a slight increase (Ulbrich et al. 2013; Akperov et al. 2015). These findings are often linked to weakening and/or shifting of the polar front, which is observed in several climate model projections (Bengtsson et al. 2009; Schuenemann et al. 2009; Collins et al. 2013). In other words, the results for winter reported here are consistent with other studies that use different climate models, cyclone detection and tracking algorithms, and/or global warming scenarios.

Studies focusing on summer are less common. Using the Bergen climate model, Orsolini and Sorteberg (2009) found an overall reduction in summer cyclone frequency in the Northern Hemisphere, especially south of 48°N. The model indicated increases in cyclone frequency between 48°N and about 70°N, once again suggesting a shift in the major storm tracks. Using the HIRHAM regional climate model, Akperov et al. (2015) found no change to summer Arctic cyclone frequency by the end of the 21st century. Both sets of experiments considered the SRES A1B scenario (700 ppm atmospheric carbon dioxide concentration by 2100), and Orsolini and Sorteberg (2009) also used the A2 scenario (800 ppm concentration). The RCP8.5 used in CESM-LE is more similar to the A2 scenario (see Figure 12.3 in Orsolini and Sorteberg (2009)).

Many similarities can be found between CESM-LE results for summer (Figure 8.1, top) and the Bergen model results reported by Orsolini and Sorteberg (2009). Since the mid-latitude storm tracks typically sit farther south in summer than winter, a northward shift in summer means fewer storms striking the UK and Scandinavia and more tracking along the east coast of Greenland and into the Icelandic Low region, which is precisely what Figure 8.1 shows. Besides the east side of Greenland, areas with increasing track density in summer include the CAA, the Beaufort Sea, and, notably, along parts of the summer AFZ, from 85°E to 185°E (175°W). This is consistent with Orsolini and Sorteberg (2009). However, CESM-LE and the Bergen model differ for the Pacific Ocean. CESM-LE
shows a decline in cyclone frequency throughout the North Pacific Ocean, indicating a weakening of the storm track. Orsolini and Sorteberg (2009) show a clear shift in the storm track instead, especially in the A2 scenario.

The results for CESM-LE shown in Figure 8.1 have fewer similarities to those of Akperov et al. (2015), who find significant change in cyclone frequency almost nowhere for April-September. The inclusion of three additional months is likely an important factor for the different results in Akperov et al. (2015). The use of different models may also play a role. Other factors, including the scenario and the detection and tracking algorithm, are less likely to be important. For instance, the A1B results in Orsolini and Sorteberg 2009) are similar to the CESM-LE results. The choice of cyclone detection and tracking algorithm has been shown to have only a modest impact on projections of cyclone activity throughout the 21st century (Ulbrich et al. 2013).

Figure 8.2. Cyclogenesis frequency for summer (JJA, top) and (bottom) winter (DJF, bottom) averaged from CESM-LE for the periods 1990-2005 (left) and 2071-2080 (center) and their difference (right). Stippling indicates a significant change based on the 95% confidence interval of a student’s t-test in which each member is considered an independent observation (30 observations per period).
Unlike the studies referenced above, Nishii et al. (2015) focus specifically on the future of summer cyclone activity over the Arctic Ocean. Using a Eulerian method of cyclone detection, they identify enhanced cyclone activity in the CAO at the end of the 21st century under RCP4.5 in the ensemble mean of seventeen CMIP5 models. These results are similar to Figure 8.1, although the focus of increased summer Arctic track density observed in CESM-LE under RCP8.5 is offset somewhat into the Beaufort Sea. Nishii et al. (2015) attribute the increased cyclone activity in the CAO in part to enhanced thermal contrast between Eurasia and the Arctic Ocean, although they also note that large-scale dynamic forcings represented by the SVNAM are likely involved in the projected changes, too.

Turning to changes in cyclogenesis, Figure 8.2 shows that many trends in track density can be attributed to changes in storm formation. In winter, for instance, many fewer cyclones develop in Baffin Bay, the Icelandic Low region, Alaska, or Kamchatka in 2071-2080 than 1990-2005. Meanwhile, cyclogenesis increases in the Sea of Okhotsk, the Chukchi Sea, the Kara Sea, northern Hudson Bay, and north of Greenland. Except for that final area, all of these regions experience a sharp transition in December from consistently over 75% sea ice concentration to essentially ice-free (average concentration < 15%) in December (Figure 8.3, top). Sea ice declines are also prominent in January and February for these regions, although December shows the most drastic changes. These results support the theory that increased open water in the Arctic Ocean will provide greater energy fluxes to the atmosphere in winter that can drive instability and cyclone development (Bengtsson et al. 2011; Jaiser et al. 2012; Vihma 2014).

In summer, the increased track density within the Arctic Ocean is not accompanied by increases in local cyclogenesis. This is despite major declines in summer sea ice concentrations (Figure 7.2) and increases to summer open water periods. Unlike in autumn and winter, the Arctic Ocean is generally cooler than the atmosphere in summer even in ice-free areas, so increasing the open water percentage does not have the same impact (Simmonds and Rudeva 2012). However, the AFZ does see increased cyclogenesis in some areas, even as cyclogenesis in areas of high topography (e.g., the Siberian mountains and Mackenzie Range) declines.
Figure 8.3. Mean December-February snow cover and sea ice concentration based on CESM-LE ensemble mean for (left) 1990-2005 and (center) 2071-2080 and (right) their difference. Stippling indicates a significant change based on the 95% confidence interval of a student’s t-test in which each member is considered an independent observation (30 observations per period).

8.3.2. Arctic Cyclone Activity

Figure 8.4 displays cyclogenesis locations for only those cyclones that contribute to cyclone frequency in the CAO+BCEL domain in summer. The only month showing any widespread significant changes is June (top). In
this month, cyclogenesis increases along the the coasts of eastern Siberia and Chukotka (AFZ sectors 5-7).

Meanwhile, cyclogenesis farther inland in these regions experiences a marked decline.

Figure 8.4. Mean June-August cyclogenesis frequency for cyclones that spend any time over the CAO+BCEL domain based on the CESM-LE ensemble mean for the periods 1990-2005 (left) and 2071-2080 (center) and their difference (right). Stippling indicates a significant change based on the 95% confidence interval of a student’s t-test in which each member is considered an independent observation (30 observations per period).

This pattern represents a shift in where CAO+BCEL cyclones are generated. Under the current climate, many cyclones in CESM-LE that enter the CAO+BCEL in summer form over the Siberian mountains. However, under
RCP8.5, cyclogenesis over the mountains is exchanged for cyclogenesis along the coast (Figure 8.5). The number of June cyclones that the eastern Siberia/Chukotka region contributes to the CAO+BCEL domain rises from about 20.0 to 22.5 per decade under RCP8.5 ($p = 0.02$). This may represent increasing importance of the AFZ in late spring and early summer. The swath of the AFZ in question, from about 130°E to 180°E, does show some strengthening in June (Figures 7.3, 7.5). The Arctic jet-like feature does, too, although to a lesser degree.

**Figure 8.5.** (a,b) Contour plots and (d,e) spaghetti plots of cyclone tracks originating in eastern Siberia (outlined in black or blue) during summer (JJA) 1990-2005 (right) and 2071-2080 (center). Also shown are the differences in (c) track density and (f) cyclogenesis frequency for cyclones originating in eastern Siberia. Stippling indicates a significant change based on the 95% confidence interval of a student’s t-test in which each member is considered an independent observation (30 observations per period).

However, two caveats should be noted here. First, recall from Section 7.4.3 that at mid-troposphere levels, the greatest changes to the AFZ were actually along the coast of the Beaufort Sea (sectors 8 and 9), and no change in cyclogenesis is observed there. Another caution is that Chapter 6 showed the AFZ as a cyclone intensifier but not a direct cyclone generator. CESM-LE is biased toward showing more cyclogenesis along the coast already.
(Section 8A.1.4), so the observed coastal cyclogenesis increase may only occur because of that bias. On the other hand, the Eady growth rates in Chapter 6 showed that the AFZ has the capacity to encourage cyclogenesis. If climate change alters other parts of the Arctic system that influence cyclone development, perhaps the AFZ could become a direct cyclone generator in the future (at least in June).

Despite greater AFZ cyclogenesis in June, the CAO+BCEL sees no significant change to track frequency in any summer month under RCP8.5, as shown in Table 8.1. Northeast Asia only accounts for about 30% of all summer cyclones entering the CAO+BCEL in CESM-LE from 1990-2005, and increases from this region are offset by decreases elsewhere. For example, a decline from 1.02 to 0.85 cyclones per year ($p = 0.03$) occurs from the combined area of the Kara, Barents, Greenland, and Norwegian Seas, at the northern end of the weakened North Atlantic storm track.

Table 8.1. Changes in cyclone variables from 1990-2005 to 2071-2080 in CESM-LE (using RCP8.5). **Bold** and **bold italics** indicate significance at the 95% and 99% levels, respectively, determined by a two-tailed student’s t-test. Intensity measures are averaged values for the CAO+BCEL domain. Cyclone frequency is measured as the number of cyclones whose tracks intersect the area of interest. AFZ cyclogenesis is calculated as the average value for 500 km by 500 km areas centered on each grid cell lying within the AFZ.

<table>
<thead>
<tr>
<th>Variable</th>
<th>Units</th>
<th>June</th>
<th>July</th>
<th>August</th>
<th>Summer</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Intensity Measures (Averaged within CAO+BCEL)</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Central Pressure</td>
<td>hPa</td>
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<td>-0.57</td>
<td>-1.21</td>
<td>-0.61</td>
</tr>
<tr>
<td>Cyclone Depth</td>
<td>hPa</td>
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<td>-0.17</td>
<td>+0.01</td>
<td>-0.18</td>
</tr>
<tr>
<td>Laplacian of Central Pressure</td>
<td>hPa (100 km)$^2$</td>
<td>-0.05</td>
<td>+0.01</td>
<td>+0.01</td>
<td>-0.01</td>
</tr>
<tr>
<td>Cyclone-Associated Precipitation</td>
<td>mm</td>
<td>+1.03</td>
<td>+2.32</td>
<td>+4.24</td>
<td>+7.55</td>
</tr>
<tr>
<td><strong>Frequency Measures</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>AFZ Cyclone Frequency</td>
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<td>+0.22</td>
<td>-0.05</td>
<td>+0.34</td>
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<tr>
<td>AFZ Cyclogenesis</td>
<td># area$^{-1}$ yr$^{-1}$</td>
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<td>-0.01</td>
<td>-0.18</td>
<td>+0.78</td>
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<tr>
<td>CAO+BCEL Cyclone Frequency</td>
<td># yr$^{-1}$</td>
<td>-0.05</td>
<td>+0.23</td>
<td>-0.04</td>
<td>+0.13</td>
</tr>
</tbody>
</table>

Turning to cyclone intensity, whether the CAO+BCEL domain observes any change in RCP8.5 depends on the metric used. The depth and Laplacian of central pressure indicate weakening of June storms, whereas central pressure indicates strengthening of storms in July and August. The lack of agreement amongst intensity measures makes the July and August results dubious. Especially since the Laplacian and depth are more robust intensity measures, the pressure difference is likely just a background change and not a change to cyclone intensity.

The June weakening, on the other hand, shows agreement between the Laplacian and depth, so this is likely a true reflection of change. Surprisingly, though, cyclone weakening is the opposite change expected if the AFZ strengthens. The average intensification rate for cyclones while crossing the AFZ in June does not change...
under RCP 8.5 regardless of the metric used. Looking regionally, the only sectors with a significant change are Sectors 8 and 9, which see an increase in cyclone intensification during AFZ crossing of +0.08 hPa (100 km)\(^{-2}\) day\(^{-1}\) (p = 0.037). (These are also the sectors with the greatest strengthening of the AFZ.) In short, the CAO+BCEL’s June cyclones are weaker in the future despite intensification within the AFZ.

Why, then, are they weaker? Table 8.2 shows the change in intensity between 1990-2005 and 2071-2080 for June cyclones that cross the CAO+BCEL by source region. Based on the Laplacian of central pressure, cyclones are weaker for all source regions except the Pacific Ocean (which contributes the fewest storms each June). Moreover, cyclones migrating from Eurasia, North America, or the Atlantic Ocean are weaker in the future before they even enter the CAO+BCEL. Especially for the Atlantic and North American cyclones, this likely results from the weakening of the polar front. Central pressure and depth show fewer significant differences, but in no region or period do they indicate significantly stronger cyclones in June.

Table 8.2. Difference in intensity of cyclones whose tracks intersect the CAO+BCEL in June between 1990-2005 and 2071-2080 under RCP8.5. Bold and bold italics indicate significance at the 95% and 99% levels, respectively, determined by a two-tailed student’s t-test. Intensity is averaged both before entry into the CAO+BCEL and while each cyclone is crossing the CAO+BCEL. Cyclones are divided based on their source region (see Figure 6.4).

<table>
<thead>
<tr>
<th>Variable</th>
<th>Units</th>
<th>Period</th>
<th>Local</th>
<th>Eurasia</th>
<th>Atlantic</th>
<th>North America</th>
<th>Pacific Ocean</th>
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<tbody>
<tr>
<td>Laplacian of</td>
<td>hPa</td>
<td>Before CAO+BCEL</td>
<td>--</td>
<td>-0.08</td>
<td>-0.07</td>
<td>-0.05</td>
<td>+0.04</td>
</tr>
<tr>
<td>Central Pressure</td>
<td>(100 km)(^{-2})</td>
<td>In CAO+BCEL</td>
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<td>-0.04</td>
<td>-0.11</td>
<td>-0.07</td>
<td>-0.01</td>
</tr>
<tr>
<td>Central Pressure</td>
<td>hPa</td>
<td>Before CAO+BCEL</td>
<td>--</td>
<td>+0.42</td>
<td>+1.41</td>
<td>-0.26</td>
<td>+0.26</td>
</tr>
<tr>
<td></td>
<td></td>
<td>In CAO+BCEL</td>
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<td>+0.42</td>
<td>+0.85</td>
<td>-0.39</td>
<td>+0.00</td>
</tr>
<tr>
<td>Depth</td>
<td>hPa</td>
<td>Before CAO+BCEL</td>
<td>--</td>
<td>-0.44</td>
<td>-0.26</td>
<td>-0.04</td>
<td>-0.39</td>
</tr>
<tr>
<td></td>
<td></td>
<td>In CAO+BCEL</td>
<td>-0.08</td>
<td>-0.79</td>
<td>-0.80</td>
<td>-0.25</td>
<td>-0.26</td>
</tr>
</tbody>
</table>

For the Eurasian cyclones, the reasons the change may be more nuanced. Because cyclogenesis over Asia shows a poleward shift, the cyclones from this region are 12 hours younger on average in the future at the time they cross the AFZ and move into the CAO+BCEL (p = 0.000). In both periods, the maximum intensity for such cyclones typically occurs about 12 hours after crossing into the CAO+BCEL. This means the cyclones in the 2071-2080 have less time to develop and deepen. Once they reach the Arctic Ocean, these cyclones begin to stagnate and dissipate on average regardless of age or intensity. The result is that although the AFZ is still acting as an intensifier, the intensification period for these cyclones is being shortened, so they never attain as great an intensity.
The final variable depicted in Table 8.1 that merits discussion is CAP, which has noticeable increases in all months. Increases in CAP can arise from 1) an increase in cyclone frequency, 2) an increase in cyclone intensity, or 3) an increase in atmospheric moisture (Finnis et al. 2007; Stroeve et al. 2011). Increased atmospheric moisture helps transfer energy throughout the Earth’s system and balance radiation inputs (O’Gorman et al. 2012), so increasing CAP because of greater moisture availability is thermodynamically driven. Changes because of cyclone activity, on the other hand, are dynamically driven.

Intensity changes are dubious for all months, and July and August see no change in frequency, so the increase in CAP is mostly thermodynamically driven. In addition to the basic temperature dependence of water vapor, increased atmospheric moisture from the substantial loss of sea ice cover is still relevant to CAP even if it does not affect cyclone intensity (Vihma 2014). Enhanced moisture fluxes can also be expected from lower latitudes (O’Gorman et al. 2012), so the extra moisture necessary for increased CAP may be both internally and externally sourced. CAP also increases in June, but to a lesser degree. Sea ice decline in the Arctic Ocean is less drastic in June than in July or August (Figure 7.2), so the local thermodynamic forcing is likely less. Some of the June increase in CAP may also be attributable to an increase in cyclone frequency.

8.3.3. The AFZ – Cyclone Relationship

Chapter 6 established the AFZ as a cyclone intensifier; significant correlations exist between the strength of the AFZ and the intensity of cyclones that enter the CAO+BCEL domain. This relationship remains consistent under RCP8.5. The histograms in Figure 8.6 show the distribution of Spearman’s correlation coefficients between AFZ strength and six cyclone characteristics for each CESM-LE member for 1990–2005 (black outlines) and 2071–2080 (green columns). The latter period has a greater range of coefficient estimates for most variables. But rather than a more uncertain or erratic climate, this likely simply reflects the number of observations in each period. Each member contains 16 years for 1990-2005 and 10 years 2071-2080. For a better understanding of the climate

26 See also any discussion of the Clausis-Clapyeron equation.
represented by each period, the same correlation coefficients were also calculated combining all members, with 480 observations for 1990-2005 (black vertical line) and 300 observations for 2071-2080 (green vertical line).

Figure 8.6. Frequency plots of Spearman’s correlation coefficients between AFZ strength at 700 hPa and average (a) Laplacian of central SLP, (b) CAP, and (c) central pressure for cyclones in the CAO+BCEL region, as well as AFZ Strength at 700 hPa and the average number of tracks that (d) cross the AFZ (red outline in inset map), (e) cross the CAO+BCEL domain (light blue in inset map), or (f) form in the AFZ for all summer months (JJA). Green solid bars show frequency for 2071-2080, and black outlined bars show frequency for 1990-2005. The height of each bar indicates how many members for which the correlation coefficient falls within the bar’s width. The green line marks the correlation coefficient that results when combining observations from all members for the 2071-2080 correlation. The black line marks the same, only for 1990-2005.
For all six variables, the climatological correlation coefficients are very similar between periods. All three intensity measures have a modest correlation with AFZ strength between 0.30 and 0.60. A stronger AFZ means stronger summer cyclones. Additionally, at least 28 of the 30 members have a correlation with the same sign for each intensity measure. The frequency measures all have positive but weak correlations (below 0.30) and a wider spread in the coefficients from individual members compared to the intensity measures. In short, the impact of the
AFZ on cyclone activity remains robust under RCP8.5. It is clearly still an intensifier, and evidence for any effect on cyclone frequency is tenuous.

Controlling for the SVNAM does not change any of these conclusions. Figure 8.7 shows the same correlation results, only after removing the linear relationship of the SVNAM from both AFZ strength and the cyclone characteristics. As with the basic correlations, more variability amongst individual members is apparent in 2071-2080 than in 1990-2005, but the climatological correlations are consistent. Removing the relationship with SVNAM weakens the correlations slightly, but for 2071-2080, the three intensity measures still yield a climatological correlation coefficient exceeding 0.30 magnitude. The frequency measures show no clear relationship with AFZ strength after accounting for SVNAM. Therefore, although results from CESM-LE show that the summer AFZ may develop earlier in the future, and although AFZ cyclogenesis in June under RCP8.5 may increase, the basic role of the AFZ as a cyclone intensification mechanism that acts independently of large-scale circulation will persist.

8.4. Observed Trends in Summer Arctic Cyclone Activity (1979-2014)

Despite dramatic changes to the Arctic climate system under RCP8.5, the main narrative for summer Arctic cyclones was one of resilience to change, especially in July. The warming experienced during 1979-2014 has been less than what occurs under RCP8.5. The AFZ also exhibits less change. Accordingly, even fewer signs of change to summer Arctic cyclone activity are present over the recent period of 1979-2014 than under RCP8.5. For instance, Figure 8.8 shows linear trends (1979-2014) in cyclogenesis for (top) MERRA and (bottom) ERA for each summer month. A few patches with significant trends exist in both reanalyses, but none covers more than a few grid cells. The most intriguing patch of increased cyclogenesis is in eastern Siberia. Although few grid cells show a significant change, the swath of increasing trends is relatively large and aligns with the one region that showed a significant strengthening in the AFZ. However, that AFZ strengthening was observed in July, not June.
Figure 8.8. Trend in cyclogenesis frequency ([events per 250,000 km² per month] per decade) over the period 1979-2014 for each summer month using data from (top) MERRA and (bottom) ERA. Stippling indicates a significant trend based on the 95% confidence interval (after removing 1-lag autoregression).

Linear trends (1979-2014) in overall cyclogenesis and cyclone frequency in the AFZ, as well as cyclone frequency and intensity measures for the CAO+BCEL domain are presented in Table 8.4. Using MERRA data, no measure of intensity or frequency has a significant trend in any summer month. ERA does show significant positive trends in central pressure for August and summer as a whole, which represents a weakening trend. However, since no trend is present for the Laplacian of central pressure or either intensity measure in MERRA, this result is unreliable. Although ERA shows a slight downward trend in the number of cyclones actually forming within the AFZ for summer overall, it shows no trend for the number of cyclones that cross the AFZ or spend time over the CAO+BCEL. Regardless, since MERRA shows no trends and the two reanalyses produce otherwise comparable cyclone regimes, these results indicate no consistent trends for either cyclone frequency or intensity. Given that,
as shown in Chapter 6, the strength of the AFZ at 700 hPa has a significant impact on summer cyclone development, this robustness in the face of Arctic change is unsurprising.

Table 8.4. Linear trends (per decade) of cyclone variables for 1979-2014 based on MERRA (top of each cell) and ERA (bottom of each cell). Bold and bold italics indicate significance at the 95% and 99% levels, respectively. Intensity measures are averaged values for the CAO+BCEL domain. Cyclone frequency is measured as the number of cyclones whose tracks intersect the area of interest. AFZ cyclogenesis is calculated as the average value for 500 km by 500 km areas centered on each grid cell lying within the AFZ.

<table>
<thead>
<tr>
<th>Variable</th>
<th>Units</th>
<th>June</th>
<th>July</th>
<th>August</th>
<th>Summer</th>
</tr>
</thead>
<tbody>
<tr>
<td>Average Laplacian of Central Pressure in CAO+BCEL</td>
<td>hPa (100 km)^{-2}</td>
<td>-0.02</td>
<td>+0.01</td>
<td>-0.01</td>
<td>-0.01</td>
</tr>
<tr>
<td>Average Central Pressure in CAO+BCEL</td>
<td>hPa</td>
<td>+0.71</td>
<td>+0.08</td>
<td>+0.27</td>
<td>+0.36</td>
</tr>
<tr>
<td>Average Cyclone-Associated Precipitation in CAO+BCEL</td>
<td>mm</td>
<td>-0.77</td>
<td>-0.01</td>
<td>+0.12</td>
<td>-0.72</td>
</tr>
<tr>
<td>AFZ Cyclone Frequency</td>
<td># yr^{-1}</td>
<td>+0.04</td>
<td>+0.11</td>
<td>+0.16</td>
<td>+0.00</td>
</tr>
<tr>
<td>AFZ Cyclogenesis</td>
<td># area^{-1} yr^{-1}</td>
<td>+0.02</td>
<td>-0.01</td>
<td>-0.01</td>
<td>+0.00</td>
</tr>
<tr>
<td>CAO+BCEL Cyclone Frequency</td>
<td># yr^{-1}</td>
<td>-0.45</td>
<td>-0.61</td>
<td>-0.16</td>
<td>-1.17</td>
</tr>
</tbody>
</table>

As noted in Section 2.5, several other studies have considered whether any trends exist in summer Arctic cyclone activity, with mixed results. Using the previous generation of reanalyses, Serreze and Barrett (2008) and Simmonds et al. (2008) found no trends in total cyclone frequency north of 70°N in summer over the periods 1957-2006 or 1979-2006. Simmonds et al. (2008) did find some evidence for strengthening of summer cyclones, but only for the former period, which includes large inhomogeneities in assimilated data. They considered those trends, therefore, to be specious. By contrast, Sepp and Jaagus (2010) found an increase in cyclone frequency and a decrease in mean cyclone central pressure north of 68°N from 1948 to 2002 in the NCEP/NCAR Reanalysis. Although each study incorporated the NCEP/NCAR Reanalysis data, each used a different tracking algorithm and study area.

Much of the reasoning behind projections for increased Arctic cyclone activity rests with the projected poleward shift in the North Atlantic storm track (König et al. 1993; McCabe et al. 2001; Bengtsson et al. 2009; Schuenemann and Cassano 2010; Collins et al. 2013). This study purposefully omits the Kara and Barents Seas from analysis because of their strong relationship with this storm track. Therefore, the trends in Table 8.4 are most
influenced by cyclone activity related to locally sourced Arctic Ocean cyclones and those forming over Eurasia. The study area of Serreze and Barrett (2008) was even more restrictive, focusing on a small area of the CAO. The two studies that identified significant trends had broader study areas of 70°N and 68°N that likely captured more cyclones from the North Atlantic storm track. These broader study areas, then, may be more likely to exhibit some trend in cyclone frequency.

8.5. Discussion of Limitations

The last two chapters have described the response of the AFZ and Arctic cyclone activity to global warming. However, before considering whether the results shown here are a likely future, several limitations to the analysis should be understood. Most significantly, this analysis uses a single climate model and a single warming scenario, both of which contribute to the uncertainty of results.

Based on the comparisons of Sections 7.2 and 8.2, CESM-LE presents a very similar depiction of both the AFZ and cyclone activity when compared to reanalyses. Three key differences for this study are: 1) CESM-LE shows the Arctic front at middle levels of the troposphere shifting south of the coastline more often and north of the coastline less often compared to the reanalyses, 2) CESM-LE shows fewer locally sourced cyclones for the CAO+BCEL and more externally sourced cyclones, and 3) the AFZ has a more direct impact on cyclogenesis in CESM-LE. Because of these differences (and others), the CESM world and the real world may not react to the same radiative forcing in the same way. Any projected change involving these aspects must be considered less certain than aspects of the Arctic climate that CESM-LE reproduces well.

For example, CESM-LE shows an extension of the summer season to earlier in the year. The AFZ develops sooner in concert with earlier snow melt. This result has higher confidence because the seasonality of the AFZ and its ties to the timing of snow melt are consistent between CESM-LE and the reanalyses. As part of this extended summer, CESM-LE also shows cyclogenesis in June becoming less focused on the Siberian mountains and more focused near and along the Arctic Ocean coastline under RCP8.5. This result has less confidence because of the bias CESM-LE shows for coastal cyclogenesis in 1990-2005.
CESM-LE also shows some bias in sea ice cover and other variables that might influence how the AFZ and Arctic cyclones respond to the forcings in RCP8.5. For example, CESM has a small but significant positive bias for summer sea ice concentration throughout most of the CAO+BCEL when compared to the combined passive microwave record for 1990-2005 (Figure 7A.2). Additionally, the sea ice edge in the Atlantic Ocean and Barents Sea extends farther south in CESM-LE. Like many models, CESM also underestimates the rate at which September sea ice has declined over the past few decades (Stroeve et al. 2007), although based on the Large Ensemble, the difference in rates may be solely from internal variability (Swart et al. 2015). If systematic and not due to internal variability, these biases would only be relevant for June and early July, when substantial sea ice is still projected within the coastal seas.

Another limitation is that RCP8.5 is just one example scenario and involves the greatest amount of greenhouse gas emissions of any IPCC AR5 scenario. It may be tempting to call this pathway alarmist. However, the IPCC clearly states that the RCPs are not predictions of the future, and no probability is assigned to them (Collins et al. 2013). Examining RCP8.5 is still useful, though, for understanding the response of the Arctic climate system to warming. For many aspects of the climate system, other scenarios yield the same changes as RCP8.5, simply at lesser rates or magnitudes (IPCC 2013). With that said, global emissions have followed RCP8.5 more closely than any other scenario since 2005 (Sanford et al. 2014; Le Quéré et al. 2015), so RCP8.5 seems a reasonable scenario to consider. Although recent emissions reduction pledges, such as those associated with the Paris Agreement (COP21 2015)\(^\text{27}\), may alter the current course, RCP8.5 remains the “business-as-usual” scenario.

The time periods considered create additional limitations. Under RCP8.5, global air temperatures will not reach equilibrium until well after 2100, so the period 2071-2080 represents a transient response, not a new normal. By 2200, changes to the Arctic may become strong enough to see more of an impact on the AFZ or cyclone development. Additionally, 2071-2080 is limiting because each member only contains ten years of data. The Arctic climate system is complex enough that ten years is not enough time to isolate some relationships. Using CESM-LE,

\(^{27}\) Go to http://unfccc.int/paris_agreement/items/9485.php for real-time updates on progress toward achieving the Paris Agreement goals.
which provides 30 times the number of observations largely remedies this issue since three hundred observations is sufficient for statistical descriptions of the 2071-2080 climate state.

However, these results still by no means predict the exact sequence of events in 2071-2080. With so much internal variability, the 2071-2080 decade may manifest that climate in a myriad of ways. Even if CESM were a perfect model and RCP8.5 the perfect scenario, much uncertainty would remain regarding specifics like the number of cyclones in the Arctic Ocean during 2071-2080 or how intense the strongest ten cyclones in this period will become.

8.6. Conclusions

The Arctic is currently experiencing uneven warming at the surface, but no coherent trends are observed for the AFZ except at the lowest model levels of the reanalyses for the period 1979-2014. Similarly, neither cyclone frequency nor intensity have changed in the CAO+BCEL domain during this period.

RCP8.5 is a strong warming scenario involving roughly four times as much warming from 1990-2005 to 2071-2080 than observed in the reanalyses. Strengthening of the AFZ is prominent in June under RCP8.5, but July and August still exhibit little change above the near-surface level. Accordingly, June exhibits enhanced cyclogenesis along the east Siberian coast, but July and August exhibit no changes to cyclone frequency. Changes to cyclone intensity are less certain, which may reflect the sometimes conflicting influences of local changes in the Arctic and changes in the mid-latitudes. The one cyclone characteristic that most clearly changes is CAP, which rises in all months. This, however, is likely driven primarily by thermodynamic changes, not changes to cyclone development.

On an interannual scale, the AFZ still plays an important role in modifying cyclone intensity under RCP8.5. The overall correlations for CESM-LE in each period show nearly identical coefficients for all cyclone intensity and frequency measures. Therefore, regardless of any changes occurring in June, the fundamental role of the AFZ as a cyclone intensifier remains robust to climate change.
CHAPTER 9. CONCLUSION

This study set out to address three main questions:

1) In what ways, if any, does the AFZ influence summer Arctic cyclone activity?

2) How does the summer AFZ respond to a global warming scenario?

3) How does the AFZ-cyclone relationship respond to a global warming scenario?

To investigate the first and third question, a new cyclone detection and tracking algorithm was developed based on several past efforts (Serreze 1995; Wernli and Schwierz 2006; Serreze and Barrett 2008; Hanley and Caballero 2012). This algorithm includes more sophisticated treatment of MCCs and cyclone interactions like merging and splitting of storms. When applied to data from the MERRA and ERA reanalyses, this algorithm reproduces the main features of Northern Hemisphere cyclone activity in both winter and summer and re-affirms several basic aspects of summer Arctic cyclone activity.

However, output from this algorithm also led to several new findings. First, contrary to past theories (Serreze et al. 2001), the AFZ is not a region of preferred summer cyclogenesis. Rather, cyclones contributing to the summer maximum in CAO+BCEL activity are more often generated locally or over the Eurasian continent. Second, although not a direct cyclone generator, the AFZ does act as a cyclone intensifier. Even after accounting for large-scale circulation patterns, cyclone intensity increases in the CAO+BCEL when the AFZ is stronger.

The response to global warming was assessed using the 30-member CESM-LE. CESM-LE has several small biases in sea ice, snow cover, and AFZ characteristics, but overall its depiction of the AFZ matches very well with that of three atmospheric reanalyses. The main bias is that at mid levels of the troposphere, meanders in the MAF are more likely to push south of the coastline in CESM-LE than in the reanalyses. This bias may in turn influence the preference in CESM-LE for less cyclogenesis within the CAO+BCEL domain and more along the AFZ and Eurasian interior. Unlike the reanalyses, CESM-LE shows slight evidence for a direct AFZ influence on cyclone frequency, but the role of the AFZ as a cyclone intensifier is well reproduced.

Current changes to the AFZ in response to climate change have been largely limited to near-surface levels, but with additional warming, the more rapid loss of snow cover in spring leads to earlier AFZ development. June AFZ strength increases throughout the troposphere at some longitudes, and some evidence exists showing an
increase in June cyclogenesis along the coast of eastern Siberia. CAP increases for every month, but these increases occur despite little coherent change to cyclone frequency in July or August or cyclone intensity in any month.

Often in discussions of global warming, focus is directed to aspects of the climate system that are highly sensitive to change. However, it is important to also recognize aspects that are robust to external forcing. Under RCP8.5, little change occurs to the AFZ or Arctic cyclone activity in July, the month when the AFZ is strongest. Additionally, the AFZ maintains its role as a cyclone intensifier. Still, the importance of the AFZ likely to increase for two reasons. First, other cyclone influences (such as the polar front) are likely to diminish. Second, as humans become more active in the Arctic Ocean, they will have to contend ever more often with Arctic storms.
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APPENDICES

LIST OF ABBREVIATIONS

AFZ: Arctic Frontal Zone
BCEL: Beaufort, Chukchi, East Siberian, and Laptev Seas
CAO: Central Arctic Ocean
CAP: Cyclone-Associated Precipitation
CESM: Community Earth System Model
CESM-LE: Community Earth System Model Large Ensemble
CFSR: Coupled Forecast System Reanalysis
CMIP: Climate Model Intercomparison Project
DMSP: Defense Meteorological Satellite Program
EASE2-Grid: Equal Area SSM/I Earth Grid (version 2)
ECHAM: ECMWF Hamburg Model
ECMWF: European Center for Medium-range Weather Forecasting
ERA: ECMWF Reanalysis (Interim version)
EGR: Eady Growth Rate
ESM: Earth System Model
GCM: Global Climate Model
GPH: Geopotential Height
IMILAST: Intercomparison of Mid Latitude Storm Diagnostics
IPCC AR5: The Fifth Assessment Report of the Intergovernmental Panel on Climate Change
JMA: Japan Meteorological Agency
JRA-25: Japanese Reanalysis (25-yr version)
MAF: Migrating Arctic Front
MCC: Multi-Center Cyclone
MERRA: Modern-Era Retrospective Analysis for Research and Applications
NASA: National Aeronautics and Space Administration
NCAR: National Center for Atmospheric Research
NCEP: National Center for Environmental Prediction
NOAA: National Oceanic and Atmospheric Administration
NSIDC: National Snow and Ice Data Center
NWP: Numerical Weather Prediction
PCMDI: Program for Climate Model Diagnosis and Intercomparison
RCP: Representative Concentration Pathway
SLP: Sea Level Pressure
SSM/I: Special Sensor Microwave Imager
SSMIS: Special Sensor Microwave Imager/Sounder
SSMR: Special Senosr Microwave Radiometer
WGS84: World Geodetic System 1984
APPENDIX TO CHAPTER 5

5A.1. Input Parameters

The algorithm described in Chapter 5 is designed to work with any of several atmospheric reanalyses with spatial resolution finer than 2°×2° latitude/longitude and a temporal resolution of 6-hr or finer. In order to achieve the needed flexibility, it incorporates eleven different parameters that can be used to tune the algorithm for different data sources or different research questions (Table 5A.1).

Table 5A.1. Cyclone detection and tracking parameters that can be used to tune the algorithm and the value used for each in this study.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value Used</th>
<th>Explanation</th>
</tr>
</thead>
<tbody>
<tr>
<td>kSize</td>
<td>1 grid cell</td>
<td>Half-length of each side of the kernel used to identify whether a grid cell is a local minimum</td>
</tr>
<tr>
<td>nanThresh</td>
<td>0.5</td>
<td>Maximum fraction of neighboring grid cells with no data allowed for a grid cell to be considered a candidate minimum</td>
</tr>
<tr>
<td>d_slp / d_dist</td>
<td>7.5 hPa (1000 km)^{-1}</td>
<td>Minimum average pressure difference at a certain distance measured from a SLP minimum required for that SLP minimum to be considered a center</td>
</tr>
<tr>
<td>maxElev</td>
<td>1500 m</td>
<td>All elevations above this threshold are masked before analysis; no centers can be detected at elevations above this threshold</td>
</tr>
<tr>
<td>contint</td>
<td>2 hPa</td>
<td>Isobar (contour) interval used when searching for the last closed isobar around a cyclone center; used to define cyclone areas and MCCs</td>
</tr>
<tr>
<td>mcctol</td>
<td>0.5</td>
<td>For multiple centers to be grouped as a MCC, this is the maximum allowed ratio of unshared area (defined by closed isobars) around the lowest pressure center to shared area for all centers</td>
</tr>
<tr>
<td>mccdist</td>
<td>1000 km</td>
<td>Maximum distance allowed between the primary center of a MCC and any secondary center</td>
</tr>
<tr>
<td>pMin</td>
<td>1.5 mm day^{-1}</td>
<td>Minimum precipitation rate (scaled to time interval) used to determine contiguous precipitation areas</td>
</tr>
<tr>
<td>rPrecip</td>
<td>250 km</td>
<td>Minimum radius for cyclone area (additive with algorithm’s cyclone area detection)</td>
</tr>
<tr>
<td>maxSpeed</td>
<td>150 km hr^{-1}</td>
<td>Defines the search radius for extending cyclone tracks</td>
</tr>
<tr>
<td>red</td>
<td>0.75</td>
<td>Modifies the projection of a cyclone center’s propagation over a time interval, accounting for the tendency for cyclone propagation to slow with age</td>
</tr>
</tbody>
</table>

5A.2. Outputs

The output from this algorithm is two-fold. First, the synoptic information for each SLP field is stored in a customized cyclone field object, which is readable by Python. These objects contain information regarding cyclone location, area, intensity, and associated precipitation. Second, a list of cyclone track objects is saved for each
month. Cyclones that exist during two months are grouped with the month in which they experience lysis. Table 5A.2 lists the full set of characteristics that are recorded at each observation time for each cyclone.

<table>
<thead>
<tr>
<th>Variable</th>
<th>Units</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>x, y</td>
<td>grid cells</td>
<td>Column and row in EASE2 grid (from upper-left) of cyclone center</td>
</tr>
<tr>
<td>Dx, Dy</td>
<td>grid cells</td>
<td>Propagation of cyclone across the EASE2 grid since last observation</td>
</tr>
<tr>
<td>long, lat</td>
<td>-180 to +180°E, -90 to 90°N</td>
<td>Longitude and latitude of cyclone center</td>
</tr>
<tr>
<td>u, v</td>
<td>km hr⁻¹</td>
<td>Zonal and meridional propagation velocity since last observation</td>
</tr>
<tr>
<td>uv</td>
<td>km hr⁻¹</td>
<td>Propagation speed since last observation</td>
</tr>
<tr>
<td>id</td>
<td>--</td>
<td>Unique ID for the cyclone center in the instantaneous cyclone field</td>
</tr>
<tr>
<td>pid</td>
<td>--</td>
<td>Unique ID of the lowest pressure cyclone center in a MCC in the instantaneous cyclone field</td>
</tr>
<tr>
<td>tid</td>
<td>--</td>
<td>Unique ID of the cyclone center track for the given month</td>
</tr>
<tr>
<td>ftid</td>
<td>--</td>
<td>Former track ID of the cyclone center track, only relevant if it existed in the prior month</td>
</tr>
<tr>
<td>ptid</td>
<td>--</td>
<td>Track ID of the primary center in a MCC</td>
</tr>
<tr>
<td>sid</td>
<td>--</td>
<td>Unique ID for the track of a cyclone system for the given month</td>
</tr>
<tr>
<td>p_cent</td>
<td>Pa</td>
<td>SLP at cyclone center</td>
</tr>
<tr>
<td>p_edge</td>
<td>Pa</td>
<td>SLP at cyclone edge (last closed isobar)</td>
</tr>
<tr>
<td>area</td>
<td>10⁴ km²</td>
<td>Area enclosed by last closed isobar</td>
</tr>
<tr>
<td>radius</td>
<td>10² km</td>
<td>Radius of a circle with the same area as cyclone</td>
</tr>
<tr>
<td>depth</td>
<td>Pa</td>
<td>Edge pressure – central pressure</td>
</tr>
<tr>
<td>DpDr</td>
<td>Pa / 10⁴ km</td>
<td>Depth / radius</td>
</tr>
<tr>
<td>DpDt</td>
<td>Pa / day</td>
<td>Deepening rate (scaled by latitude)</td>
</tr>
<tr>
<td>DsqP</td>
<td>Pa / 10⁴ km²</td>
<td>Laplacian of central pressure (∇²p)</td>
</tr>
<tr>
<td>precip</td>
<td>mm</td>
<td>Cyclone-associated precipitation recorded since last observation</td>
</tr>
<tr>
<td>precipArea</td>
<td>10⁴ km²</td>
<td>Area over which cyclone-associated precipitation fell</td>
</tr>
<tr>
<td>type</td>
<td>0, 1, 2</td>
<td>1 = primary center, 2 = secondary center (in a MCC), 0 = this row is only present for calculating propagation (used during splits, merges, and lysis events)</td>
</tr>
<tr>
<td>centers</td>
<td>#</td>
<td>Number of centers in the cyclone system; if a secondary center of a MCC, set to 0</td>
</tr>
<tr>
<td>time</td>
<td>days</td>
<td>Days since 1 Jan 1900 0000 UTC</td>
</tr>
<tr>
<td>Ege, Ely,</td>
<td></td>
<td>Records whether the cyclone experienced genesis (ge), lysis (ly), merging (mg), splitting (sp), or regeneration (rg); 0 = no event, 1 = center-only, 2 = area-only, 3 = both center and area involved in event</td>
</tr>
</tbody>
</table>

5A.3. Limitations

As with any cyclone detection and tracking algorithm, this one has several limitations. For example, the use of closed isobars to define cyclone areas is imperfect because of interference that occurs between nearby systems (Neu et al. 2013; Wernli and Schwierz 2006; Hanley and Caballero 2012). The algorithm is biased toward
detecting larger areas for more isolated cyclones. This can be observed at 0900Z 25 September 1989 (Figure 5A.1), where the Greenland Sea is impacted by two strong cyclones, but only a small fraction of the region is detected as being within a cyclone area.

Problems with interference are mitigated by the inclusion of MCC detection. In the subsequent observation time (1200Z 25 September 1989; Figure 5A.2), the two systems merge together (and spawn two additional centers of low pressure). The cyclone area over the Greenland Sea is now broader, and likely more realistic. However, including MCCs involves more parameters, more computation time, and more opportunities for divergence between two detection and tracking algorithms. Whereas Hanley and Caballero (2012) report 32% of all winter cyclone tracks being a part of a MCC at some point in their lifespan, the output from this algorithm with the previously stated parameters yields 38% (with ERA data as the input). The exact reasons for this discrepancy are unclear, though, because the methods of detection and tracking differ in several ways. The inclusion of MCC detection is still very new, and more experimentation and sensitivity assessments would improve understanding of how best to incorporate them.

The tracking method for this algorithm is simpler than some, but especially for the 3-hr temporal resolution of MERRA, the distance travelled by cyclone centers for each observation is most often smaller than the distance between cyclones. This is demonstrated in Figures 5A.1 and 5A.2, where the green circles represent the positions of centers in the previous time, while the black and red squares represent the positions of centers in the current time. Between 0600Z and 0900Z on 25 September 1989, no center moved more than two grid cells in any direction, and nearest neighbor tracking was the only mechanism necessary. Between 0900Z and 1200Z, a more complicated situation developed as two new centers appeared within a MCC in the Greenland Sea and another two systems merged to make a MCC just south of Greenland, but the fastest movement was still only three grid cells, providing more confidence that the matching between two observation times is accurate.

With 6-hr data, the nearest neighbor approach is more likely to make tracking errors, but this likelihood is reduced by the inclusion of a predicted location based on past propagation. Using a predicted location further constrains the possible ways in which a track may be extended, as shown in Figure 5.4. Adding this feature increases computation time, but it also makes the algorithm more flexible than its predecessor (Serreze 1995).
Figure 5A.1. SLP isobars (2 hPa interval) and detected cyclone centers (black dots) and areas (grey shading) for 0900Z 25 September 1989. Green dots show cyclone centers detected in the prior step (0600Z 25 September 1989). Elevations over 1500 m are masked in white.

Figure 5A.2. SLP isobars (2 hPa interval) and detected cyclone centers (black dots for primary, red dots for secondary) and areas (grey shading) for 1200Z 25 September 1989. Green dots show cyclone centers detected in the prior step (0900Z 25 September 1989). Elevations over 1500 m are masked in white.

Other limitations to the output from this algorithm are determined by the user-defined parameters (Table 5A.1). For instance, including a minimum SLP gradient for centers will help eliminate heat lows and other spurious minima, but it also risks eliminating some authentic centers that are merely weak and perhaps in early or late
stages of development. Similarly, masking high elevation avoids the problem of SLP extrapolation, but it may also truncate the tracks of real synoptic systems that originate at high elevation (Rudeva et al. 2014).

5A.4. Comparisons to Other Algorithms

5A.4.1. Cyclone Frequency Comparison

Before using this new cyclone detection and tracking algorithm to answer research questions, its behavior was compared to previously published algorithms to ensure consistency in established features of northern hemisphere cyclone characteristics. The IMILAST project has compared cyclone characteristics from 15 different algorithms (Neu et al. 2013), providing a broad view of the range of possible results. The mean and extreme seasonal cyclone track frequency from that project for the period 1989-2008 is presented in Table 5A.3 alongside the results from the new algorithm. Results from Hanley and Caballero (2012), whose algorithm is the only other to incorporate both MCCs and merging and splitting events, is also shown.

In both seasons, the new algorithm performs well within the range of IMILAST algorithms in terms of cyclone track frequency for each input type. However, the new algorithm yields one uncommon result: it identifies about 130 more cyclones in winter than summer on average whereas a majority (2/3) of algorithms identify more cyclones in summer. All five algorithms reviewed in Neu et al. (2013) that identify more cyclones in winter use vorticity or the Laplacian of central pressure in the detection stage. Of these, only Murray and Simmonds (1991) and Sinclair (1997) identify at least 100 more cyclones per winter than per summer. Since initial detection of minima is more similar to Wernli and Schwierz (2006) and Serreze and Barrett (2008) in this new algorithm, detecting more cyclones in winter is unexpected.

One potential reason for the difference is the incorporation of MCCs. However, Hanley and Caballero (2012), who use a SLP method incorporating MCCs, find an average of 360 cyclones in winter and 410 in summer. The percentage of tracks in the new algorithm that are part of a MCC at some point in their lifespan is also consistent between winter (38% using ERA) and summer (39%) for this period. Additionally, running the new algorithm with MCC detection turned off still returns more winter cyclones.
A more likely reason for the lower summer values is that, compared to other algorithms, the new one detects fewer weak systems. **Figure 5A.3** compares the intensity distributions for winter cyclones in the Northern Hemisphere for each algorithm examined by Neu et al. (2013). Results from the new algorithm are indicated by red and blue dots for MERRA and ERA inputs, respectively. For most bins, the new algorithm lies at the upper end of the interquartile range. But for the highest-pressure bin (1000-1010 hPa), the new algorithm lies at the lower end. Although such comparisons are only possible for winter, summer cyclones tend to be both higher pressure and shallower (less distinct from background pressure) (Ulbrich et al. 2009; Neu et al. 2013). Therefore, the more stringent behavior of the new algorithm is likely to limit summer cyclone counts more so than winter counts.

**Table 5A.3.** Average seasonal cyclone track frequency for the area north of 30°N in winter (DJF) and summer (JJA) for the new algorithm compared to prior studies. To match the IMILAST project, analyses are for the period Jun 1989 to Feb 2009.

<table>
<thead>
<tr>
<th>Algorithm</th>
<th>DJF</th>
<th>JJA</th>
<th>JJA - DJF</th>
</tr>
</thead>
<tbody>
<tr>
<td>New Algorithm (MERRA 3-hr)</td>
<td>537</td>
<td>402</td>
<td>-135</td>
</tr>
<tr>
<td>New Algorithm (MERRA 6-hr)</td>
<td>534</td>
<td>398</td>
<td>-136</td>
</tr>
<tr>
<td>New Algorithm (ERA 6-hr)</td>
<td>458</td>
<td>380</td>
<td>-128</td>
</tr>
<tr>
<td>Hanley &amp; Caballero (2012)</td>
<td>360</td>
<td>410</td>
<td>+50</td>
</tr>
<tr>
<td>IMILAST Minimum</td>
<td>285</td>
<td>255</td>
<td>-155</td>
</tr>
<tr>
<td>IMILAST Average</td>
<td>620</td>
<td>759</td>
<td>+139</td>
</tr>
<tr>
<td>IMILAST Maximum</td>
<td>1070</td>
<td>1425</td>
<td>+790</td>
</tr>
</tbody>
</table>

However, differences in cyclone intensity may also relate to different spatial resolution in this study (100 km by 100 km) and Neu et al. (2013) (1.5° by 1.5°). Finer resolution data typically has deeper low pressure minima (Blender and Schubert 2000; Pinto et al. 2005; Jung et al. 2006), and below about 70°N, the inputs for this study are indeed finer.

Another potential source of variation in cyclone detection is the input data. Neu et al. (2013) and Hanley and Caballero (2012) used data from ERA, however, the difference between winter and summer is consistent regardless of input. As noted in section 4.3.3., studies using MERRA have a tendency to identify more cyclones than those using ERA (Tilinina et al. 2013), and discrepancies are greatest for high elevation areas in summer (Raible et al. 2008). This algorithm shows the same bias, especially in winter. Additionally, consistent with past work by Hodges et al. (2011), **Figure 5A.3** shows that MERRA has a slightly higher fraction (2.1%) of cyclones in the deepest bin (pressure < 950 hPa) than does ERA (1.6%).
**Figure 5A.3.** Annotated version of Figure 3a from (Neu et al. 2013), which shows the “normalized distribution and statistical spread between methods of different cyclone life cycle characteristics, in box–whisker format” for cyclone intensity (minimum central pressure) for the northern hemisphere in winter (DJF). “Each bar indicates the fraction of cyclones detected in the parameter range denoted by the x-axis labels on either side. The horizontal line indicates the mode of all methods and the box the standard deviation; the whiskers extend to the maximum and minimum values.” Results from the new algorithm are compared using red and blue dots for 6-hr MERRA and ERA inputs, respectively.

A final factor to consider in cyclone counts is the maximum allowed propagation speed. The new algorithm was run with a very liberal maximum of 150 km hr$^{-1}$, which is almost twice as fast as algorithms by Sinclair (1994) (83 km hr$^{-1}$) and Blender et al. (1997) (80 km hr$^{-1}$). A more generous maximum propagation speed should lead to longer tracks but also may lead to fewer cyclones. This is because at a lower maximum propagation speed, a track continuation might fail, leading to two separate tracks (Rudeva et al. 2014).

Overall cyclone counts vary widely by algorithm, but Neu et al. (2013) found that all algorithms show the same general spatial patterns of cyclone frequency except around high elevation areas, which some algorithms mask out. In comparison to Figure 1 in Neu et al. (2013), **Figure 5A.4** shows the percentage of cyclone center density for winter (DJF) and summer (JJA) in ERA$^{28}$. The same general patterns are discernable in Figure 5A.4a as for the 15 algorithms reviewed by Neu et al. (2013). Distinct maxima in cyclone frequency occur in the Gulf of Alaska and just off the southeast coast of Greenland, and the main North Pacific and North Atlantic storm tracks are strongly shown. Relative maxima are also noticeable near regions of high topography, including northwest Greenland and the leeward side of the North American Cordillera. The Arctic Ocean and adjacent seas exhibit a

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$^{28}$ Results from MERRA are very similar, but ERA is used here for more direct comparison to Neu et al. (2013), who also used ERA. See Chapter 8 and its appendix for side-by-side comparisons.
gradient of higher frequency on the Atlantic side and lower frequency on the Pacific side. Lastly, a distinct local maximum in the Mediterranean area is located over Italy. In short, this new algorithm has strong qualitative agreement with past algorithms with regard to the general distribution of cyclones in winter.

Fewer studies exist describing characteristics of summertime extratropical cyclones, and review studies such as Neu et al. (2013) and Ulbrich et al. (2009) focus on the winter season for each hemisphere. However, the maps of summer northern hemisphere cyclone frequency that have been published (e.g., Zhang et al. 2004; Wernli and Schwierz 2006) compare well to the output presented in Figure 5A.4b. For instance, the main storm track regions in the North Atlantic and North Pacific exhibit lower cyclone center density in summer than winter, but they remain areas of relatively high cyclone frequency. Similarly, many areas of leeside cyclogenesis, such as the US Rockies and around Greenland experience fewer cyclone centers in summer.

On the other hand, several areas exhibit more cyclone activity in summer than winter. The most notable is the Arctic Ocean, which in summer is a distinct area of high center density, comparable in magnitude to center density in the Gulf of Alaska. The summer Arctic cyclone maximum has been well documented by both manual (e.g., Reed and Kunkel 1960) and automated methods (e.g., Serreze et al. 2001; Serreze and Barrett 2008). However, neither Zhang et al. (2004) nor Wernli and Schwierz (2006) present an Arctic cyclone maximum quite as strong as Serreze and Barrett (2008) or this new algorithm.

Other regions with more cyclone centers in summer include the high latitude continents and areas that exhibit a strong summer monsoon. A summer increase in cyclone center density across Siberia is common to all algorithms with available maps. The summer monsoon impact is most obvious in Wernli and Schwierz (2006), where cyclone center density is especially high in southern Asia and the southwestern USA. Although forming closed lows, these features are often unrelated to synoptic-scale cyclones. Since these monsoon areas often coincide with high topography, their prominence is reduced by employing an elevation mask. Finally, the new algorithm shows that the lee of the Mackenzie Range has higher cyclone center density in summer, which matches well with Zhang et al. (2004). Wernli and Schwierz (2006) show a slight difference. Cyclone center density patterns are by no means identical using different algorithms, but the main spatial features and seasonal contrasts show general agreement. Variation is mostly the magnitude of cyclone center maxima, not presence or absence.
Figure 5A.4. Percentage of cyclone center density (cyclones per 1000 km by 1000 km area centered on each grid cell per time interval) for winter (DJF; top) and summer (JJA; bottom) during the period 1990-2005 based on 6-hr ERA data.
5A.4.2. Event Frequency Comparison

![Diagram showing event frequency comparison between winter (DJF) and summer (JJA) for cyclone genesis and lysis events.](image)

**Figure 5A.5.** Frequency of cyclone genesis (a,b) and lysis (c,d) events in winter (DJF; a,c) and summer (JJA; b,d) for the area north of 30°N during the period 1990-2005 based on ERA 6-hr data. Frequency is calculated as the number of events per 500 by 500 km area centered on each grid cell per season.

Cyclone life cycle events such as genesis and lysis can also be compared. As with other algorithms (e.g., Wernli and Schwierz 2006; Hanley and Caballero 2012), cyclogenesis is particularly common on the leeward side of high topography such as the North American Cordillera, southeast of Greenland, and around northern Italy (Figure 5A.5a-b). Genesis is also more common on the western side of the North Atlantic and North Pacific Oceans, especially in winter. In summer, genesis along the eastern coasts of Asia and North America is diminished, but the
leeside cyclogenesis remains strong. Cyclogenesis increases over Siberia and in areas experiencing a monsoon (e.g. southern Asia and southwestern USA).

For this reason, many regions with frequent cyclogenesis also exhibit frequent cyclolysis (Figure 5A.5c-d). In winter, lysis is also relatively common around the Great Lakes, the Adriatic Sea, the northern Pacific Ocean (especially the Gulf of Alaska), between Iceland and Greenland, and in the Norwegian and Barents Seas. In summer, the Great Lakes and Adriatic Sea no longer experience notable cyclolysis, and the Icelandic and Aleutian Lows experience less. On the other hand, the central Arctic Ocean, the East Siberian Sea, and the Laptev Sea all exhibit the same frequency of cyclolysis in both seasons. These general patterns are also observed in the output from other algorithms (Trigo 2005; Wernli and Schwierz 2006; Hanley and Caballero 2012).

Two other life cycle events that are recorded using this new algorithm are the splitting and merging of cyclones (Figure 5A.6). Inatsu (2009) devised a tracking algorithm that included splits and mergers and found that merging in winter was most common east of Japan and just east of Newfoundland. Splitting was most common farther east along those main storm tracks. Hanley and Caballero (2012) also identified splits and mergers, but their detection and tracking algorithm is based on minima in SLP whereas the algorithm used by Inatsu (2009) is based on areas that exceed 10 ms^{-1} meridional wind at 850 hPa. In the Pacific Ocean, Hanley and Caballero (2012) find similar patterns as Inatsu (2009). Mergers are most common just east of Japan and splits are more common east of 180° longitude. However, the two algorithms exhibit several differences. Hanley and Caballero (2012) identify the Mediterranean as an area of high merger frequency and split frequency, whereas Inatsu (2009) shows this area as devoid of both event types. Hanley and Caballero (2012) also identify greater cyclone merge and split frequency around the Icelandic Low and the Norwegian Sea and much lower frequency around Newfoundland. For both algorithms, merging and splitting is generally more common where cyclone center density is high.

The new algorithm shows patterns of merging and splitting that likewise are highest in areas of high center density. This is not surprising because in the time interval before a merge or after a split, two cyclone centers must exist in close proximity. The new algorithm is very similar to Hanley and Caballero’s (2012) in the Atlantic, but it appears more similar to Inatsu (2009) in the Pacific. In the Pacific, it is also more likely to show merging somewhat farther northeast, nearer to Kamchatka than Japan. Like Hanley and Caballero’s (2012)
algorithm, it shows some merging and splitting in the Mediterranean in winter, but Hanley and Caballero's (2012) show substantially more.

**Figure 5A.6.** Frequency of cyclone merging (a,b) and splitting (c,d) events in winter (DJF; a,c) and summer (JJA; b,d) for the area north of 30°N during the period 1990-2005 based on ERA 6-hr data. Frequency is calculated as the number of events per 500 by 500 km area centered on each grid cell per season.

Another measure of splitting and merging events is how many different tracks experience splitting or merging during their lifespans (Table 5A.4). About of quarter of all winter cyclones detected with this algorithm experience merging or splitting, which matches the results of Hanley and Caballero (2012). In general, the new algorithm shows that one cyclone splitting into two distinct systems is more common than two distinct systems
merging together. This finding agrees with the observations of Inatsu (2009), but conflicts with the results reported by Hanley and Caballero (2012). The reasons for these differences are unclear, and the disparities amongst the algorithms indicate that more work is required to determine the best way of treating such events in cyclone tracking algorithms. Therefore, since splitting and merging of cyclones is not the focus of the current study, such events are not closely analyzed in subsequent chapters.

**Table 5A.4.** Percent frequency of merging and splitting events by season for three different input data sets. Includes only system tracks that exist above 30°N with a lifespan of at least 24 hours and a track length greater than 0 m for the period 1979-2014. Also shown are the results reported by Hanley and Caballero (2012) using a different algorithm.

<table>
<thead>
<tr>
<th>Case</th>
<th>DJF</th>
<th>JJA</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>HC2012 ERA 6-hr</td>
<td>ERA 6-hr MERRA 6-hr</td>
</tr>
<tr>
<td>Non-branching</td>
<td>76</td>
<td>76</td>
</tr>
<tr>
<td>Merge</td>
<td>20</td>
<td>14</td>
</tr>
<tr>
<td>Split</td>
<td>11</td>
<td>16</td>
</tr>
<tr>
<td>Merge and Split</td>
<td>9</td>
<td>6</td>
</tr>
</tbody>
</table>

**5A.4.3. Treatment of the Severe October 1992 Nome Storm**

Including MCCs permits the assessment of complicated interactions amongst cyclone tracks. Consider the severe October 1992 storm depicted in Figure 5A.7. If MCCs were not included in the detection phase, the algorithm would split the track of this storm into multiple parts (solid red lines), and no relation would be detected between the cyclone track over the Sea of Okhotsk on 4 October (marked A) and the cyclone track over the Chukchi Sea on 7-10 October (marked B). This matches the depiction of the cyclone track by the original Serreze (1995) algorithm.
Figure 5A.7. Track of the October 1992 storm described in Mesquita et al. (2009). The position of the storm at 0000 UTC is marked for each day from 1 October through 10 October. The track of the storm without inclusion of MCCs is indicated by the red solid lines, resulting in identification of two separate tracks (marked A and B). When including MCCs, the algorithm identifies a single cyclone track (solid red plus dashed pink lines).

Manual analysis has shown that these two tracks are really the same system (Mesquita et al. 2009). The original algorithm was fooled by temporary weakening of the cyclone as it interacted with orography. However, inclusion of MCCs allows this fairly simple SLP-based algorithm to connect the two parts into a single track (dashed pink lines). At two different times in this cyclone’s lifespan, it developed a secondary minimum. The first time was 3 October at 0000 UTC, but this secondary minimum only lasted 6 hours. The second time was 5 October at 1500 UTC. While the primary center of the cyclone stagnated in Penzhin Bay, a secondary minimum developed just offshore of Chukotka. The minimum in Penzhin Bay soon dissipated, but the secondary minimum endured and began migrating northward. Since the minimum that first appeared offshore of Chukotka was a secondary minimum of the cyclone in Penzhin Bay, the algorithm recognizes the former as a continuation of the latter. This storm thus exemplifies how the updated algorithm is less prone to ending tracks prematurely than was its predecessor.
5A.5. Influence of Input Data

The input data used in a cyclone detection and tracking algorithm can have significant impacts on the output (Hodges et al. 2003; Raible et al. 2008; Tilinina et al. 2013). All datasets used in this study are reprojected to the EASE2 grid before being used in the detection and tracking algorithm, but two different temporal resolutions are used for the MERRA inputs. As discussed in Chapter 4, finer temporal resolution leads to more accurate cyclone tracking. Since cyclones move less distance between observations, the correct match between two observation times is often less ambiguous with finer temporal resolution. Additionally, finer resolution allows more precise identification of genesis and lysis events and increases the likelihood that cyclones near the 24-hr lifespan limit might exceed it.

Figure 5A.8. (top) Average cyclone center density (% frequency per 500 km by 500 km area) for summer (JJA) 1990–2005 using three input datasets and (bottom) the difference between each pair of datasets expressed as a ratio. On the bottom, red indicates the first dataset listed has higher center frequency, whereas blue indicates the second dataset listed has the higher center frequency.
The mean summer cyclone center densities from three inputs: 3-hr MERRA data, 6-hr MERRA data, and 6-hr ERA data are compared in Figure 5A.8. The three results are very similar overall, but several differences in cyclone center density can be found around areas of high elevation, especially along the edge of Tibet. Looking more closely at the differences, the Arctic Ocean shows nearly identical results for 3-hr and 6-hr MERRA data, suggesting that the finer resolution may not add much value beyond more precision for characteristics like intensity and CAP. However, in other areas, especially over the continents, the 3-hr inputs consistently yield slightly lower cyclone center densities. The differences between MERRA and ERA 6-hr inputs, on the other hand, are more varied, with MERRA yielding more cyclones in some areas (e.g., eastern North America) and ERA yielding more in others (e.g., east-central China).

The clear bias between the 3-hr and 6-hr data is confusing because the 3-hr data actually yields more cyclone tracks (Table 5A.3). How can it possibly yield fewer cyclone centers each observation time? The answer may lie in Table 5A.5, which compares several aggregate cyclone track characteristics, including average track length, lifespan, and maximum intensity. The average maximum cyclone intensity is about the same for all inputs, but the average track length and lifespan for cyclones identified using 3-hr data are notably lower than for cyclones identified using 6-hr data. In other words, the 3-hr data yields more but shorter tracks, whereas the 6-hr data yields fewer but longer tracks. The longer tracks have more cyclone center observations, which leads to the higher cyclone center density observed in Figure 5A.8.

| Table 5A.5. Aggregate cyclone characteristics for summer (JJA) 1990-2005 using three input datasets. The mean for each variable is written on the left of each grid cell, and the 97.5% and 2.5% percentiles are written to the upper right and lower right, respectively. (Several variables have very skewed distributions.) |
|---|---|---|---|
| Variable | Units | MERRA 3-hr | MERRA 6-hr | ERA 6-hr |
| Track Count | # year⁻¹ | 380 | 338 | 345 |
| Lifespan | days | 2.7 | 7.6 | 11.5 | 3.3 | 11.3 | 1.0 |
| Track Length | km | 1912 | 5928 | 2482 | 8928 | 3.3 | 11.3 | 1.0 |
| Minimum Central Pressure | hPa | 995 | 1008 | 995 | 1009 | 995 | 1009 | 978 |
| Maximum Laplacian of Central Pressure | hPa (100 km)² | 2.2 | 4.6 | 2.2 | 4.6 | 2.4 | 4.9 | 1.0 |
| Maximum Deepening Rate | hPa day⁻¹ | -17.5 | -1.7 | -14.1 | +1.4 | -14.9 | +1.5 | -41.9 |
What might be happening here is that during the tracking phase, cyclone centers identified in high
elevation areas, especially along the edges of regions that have been masked, are often spurious centers. They are
caused by complex topography and the sometimes inaccurate extrapolation of SLP from high elevation. These
centers rarely migrate out of their high elevation areas with 3-hr data because the search radius for track
continuation is only 450 km (with 150 km hr\(^{-1}\) set as the maximum propagation speed). With the 6-hr data, this
radius is expanded to 900 km, so cyclogenesis occurring nearby is more likely to be erroneously identified as
continuation of a spurious cyclone center sitting in the high elevation zone. Accordingly, 3-hr MERRA data yields
higher cyclogenesis frequency over the oceans and low elevations whereas 6-hr MERRA data yields higher
cyclogenesis in northern Tibet and the Great Basin of the western USA (not shown).

This issue in 6-hr tracking may be mitigated by lowering the elevation threshold from 1500 m to 1000 m,
by reducing the maximum number of masked grid cells allowed to border a cyclone center from four to three or
two, or by lowering the maximum allowed propagation speed. Each of these alterations would also eliminate some
true synoptic systems, but depending on the purpose of the given study, doing so may be acceptable.

For the current study, the focus is on the Arctic Ocean and adjacent high-latitude land, where 3-hr and 6-
hr data show better agreement. Therefore, these discrepancies are of lesser concern. For Chapter 6, which
requires precise locating of cyclone centers in relation to the AFZ, 3-hr data is used. However, Chapter 8 focuses on
CESM-LE, for which data are only available at a 6-hr resolution.

5A.6. Tracking Cyclone Centers Versus Cyclone Systems

Since the inclusion of MCC detection is a novel feature of the new algorithm, Figure 5A.9 shows how
aggregate cyclone track and event frequency would change in summer if MCC detection were turn off in the
algorithm. The left-hand column is compiled from a collection of cyclone tracks in which each MCC is treated as a
single system (cyclone system tracks). The center column is compiled from the same collection, only the track of
each cyclone center is treated independently (cyclone center tracks). The right-hand column shows the difference
between these two methods. North of about 40°N, most areas see an increase of about 1 to 3 tracks each season if
MCCs are split into individual tracks for each center. Overall, this represents an increase of about 10% in overall track counts.\(^{29}\) Genesis and lysis events also, logically, increase.

However, track density actually decreases in a few areas, such as just off the coast of southeast Greenland. This may occur because MCCs are more likely to have longer lifespans than their single-center counterparts. Even if one center in a MCC experiences lysis, the system will continue if the another center persists. Therefore, they are more likely to exceed the 24 hr lifespan threshold.

MCC detection has a lesser impact in the Mediterranean region and the northeast USA. MCCs will only be identified when two cyclone centers are close to each other, so they are less likely to be identified in these areas of low cyclone center density.

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\(^{29}\) An average of 418 summer center tracks and 497 winter center tracks are detected each year from June 1989 to February 2009 in ERA. The number of system tracks detected are 380 for summer and 458 for winter, as noted in Table 5A.3.
Figure 5A.9. Frequency of (top) center tracks, (middle) center genesis events, and (bottom) center lysis events for summer (JJA), averaged for the period 1990-2005, based on ERA 6-hr data. Frequencies are calculated for (left) cyclone system tracks (multi-center cyclones included) and (center) cyclone center tracks (all cyclones are single-center). The difference between center tracks and system tracks shown in the right-hand column. Units are the number of tracks or events per 500 km by 500 km area per season.
APPENDIX TO CHAPTER 6

6A.1. Thermal Tide Relationship by Season

Figure 6A.1. Average deepening rate (hPa day⁻¹) for all observations of cyclones that at any point in their tracks crossed through the AFZ by local time of day (hours) for each season 1979-2014 (from MERRA 3-hr data).

Figure 6A.1 shows the average deepening rate (hPa day⁻¹) for all observations of cyclones that at any point in their tracks cross the Arctic Ocean coastline between longitudes 41°E eastward to 126°W by local time of day (hours) for each season. Winter shows no discernible relationship between deepening rate and local time of day. A signal only becomes apparent in the Arctic when there is substantial insolation, and therefore heating of the surface by shortwave radiation. Also notable is that in spring, summer, and autumn, the thermal tide signal is diurnal. A diurnal cycle that only appears in sunlit seasons is consistent with a forcing by local sensible heat fluxes, as discussed by Dai and Wang (1999). Dai and Wang (1999) also noted a peak in the diurnal thermal atmospheric tide in mid to late morning around the Arctic landmasses, which is echoed in Figure 6A.1 by the switch from deepening to filling at about 0900 local time.

To my knowledge, no previous study of Arctic cyclones has controlled for the thermal tide’s influence on SLP. This may be in part because it is more easily detected with the 3-hr time interval used here; however, Dai and
Wang (1999) also show that thermal atmospheric tides exhibit substantial spatial variation in phase, amplitude, and frequency. For some locations the semidiurnal atmospheric tide is substantially stronger than the diurnal. Therefore, it is possible that only in a study like this, with a single season and confined spatial extent, will the impact of a thermal atmospheric tide be apparent in aggregate statistics.

### 6A.2. Residuals versus Fitted Values for Regression Model for 700 hPa AFZ

Figure 6A.2. Residuals from the regression model (Eq. 5.4) in which AFZ strength is measured at 700 hPa plotted against the model’s fitted values.

**Figure 6A.2** shows the residuals for the regression model in which AFZ strength is measured at 700 hPa plotted against the model’s fitted values. The residuals display some heteroskedasticity. When deepening (negative fitted values) is predicted by the model, weaker deepening is more often under-predicted (negative residuals), while strong deepening is more often over-predicted (positive residuals). Bias is less clear when filling (positive fitted values) is predicted by the model. Using a Breuch-Pagan test on these residuals yields a significant test statistic, which indicates that the null hypothesis of homoskedasticity can be rejected for 700 hPa and some bias is present in the model. Using the same test, the residuals from the model in which AFZ strength is measured at 500 hPa does not yield a significant test statistic, so the null hypothesis of homoskedascity cannot be rejected. Bias indicated by heteroskedasticity in the residuals of the 700 hPa model might be resolved by controlling for other influences on a cyclone’s deepening rate or by explicitly including upper-level divergence rather than implicitly including it by using measures of a cyclone’s life cycle.

Since the main goal of this paper is to assess impacts of the AFZ on the development of cyclones that impact the CAO+BCEL region, the focus is on the mid levels of the troposphere. However, traditionally the strength of the AFZ has more often been assessed at low levels of the atmosphere, such as 850 hPa, 925 hPa, or 2 m (e.g., Reed and Kunkel 1960; Krebs and Barry 1970; Serreze et al. 2001; Crawford and Serreze 2015). Interestingly, although the strength of horizontal temperature gradients, and therefore AFZ strength, at multiple levels of the atmosphere are closely aligned on a seasonal timescale, at daily and hourly timescales AFZ strength in the lower troposphere is negatively correlated with the AFZ strength in the mid troposphere.

One likely reason for this is that although cyclones are shaped by their environment, they in turn influence that environment. To illustrate, consider as a case study a cyclone that developed in July 1981, chosen because it is typical of storms generated in the lee of the Siberian mountains. The panels in Figures 6A.3-6A.5 illustrate atmospheric conditions at 0000 UTC on 20 July, 21 July, and 22 July 1981, respectively.

Those for 0000 UTC on 20 July depict conditions 24 hours before lee-side cyclogenesis. Two systems already exist in the Arctic Ocean at this time (Figure 6A.3a). The cold air advection associated with a cyclone located near the Taymyr Peninsula is particularly strong (Figure 6A.3b) and follows a strong surface front that stretches across the Kolyma Lowland (Figure 6A.3c). South of where the cyclone has just passed, the surface (2 m) expression of the AFZ is weak. However, where the coastline is free from the influence of cyclones (and in fact influenced by an anticyclone), along the East Siberian, Chukchi, and Beaufort seas, the AFZ is quite strong (8-10 K (100 km)⁻¹). Meanwhile, a trough in the 500-hPa height extends down over the Taymyr Peninsula and Lena River Delta, similar in position to the climatological Urals trough.

Over the next 24 hours, the two cyclones over the Arctic Ocean continue to circle around the North Pole. Meanwhile, the trough that was over the Taymyr Peninsula becomes accentuated as it shifts eastward (Figure 6A.4a). The downstream side of the trough passes from southwest to northeast over the Verkhoyansk Range, and a cyclone develops in the lee of the mountains, at an elevation of about 500 m. Counterclockwise motion around the incipient low brings warm air northward across the eastern Kolyma Lowland (Figure 6A.4b). This warm air over

the land surface leads to stronger coastal temperature gradients. Regional AFZ strength increases from 8-10 K (100 km)$^{-1}$ at 0000 UTC on 20 July to 10-16 K (100 km)$^{-1}$ 24 hours later.

Figure 6A.5 shows the migration of this new cyclone over the next 24 hours. Mid-level winds steer the cyclone to the northeast, crossing over the coastline and the AFZ, where the pressure tendency switches from positive to negative (filling to deepening). It continues into the Arctic Ocean and begins to merge with the two cyclones already present (Figure 6A.5a). With the cyclone now located north of the coastline, counterclockwise motion leads to cold air advection over the Kolyma Lowland (Figure 6A.5b). The impact on the AFZ is impressive, for the coastline of the Kolyma Lowland exhibits no substantial horizontal temperature gradients. Instead of the coast, the most prominent front in eastern Siberia is temporarily the cold front associated with this growing cyclone (Figure 6A.5c). Meanwhile, the Alaskan coast, which is unaffected by the cyclone’s advection, maintains a strong AFZ during all three synoptic times.

This example helps show that whereas the mid-level expression of the AFZ may influence the intensity of cyclones, temperature advection patterns associated with cyclones influence the surface expression of the AFZ, at least on day-to-day time scales. Since the AFZ involves air over land surfaces being warmer than air over ocean surfaces, warm air advection over coastal lands leads to stronger temperature gradients across the coastline. Conversely, cold air advection over land leads to weaker temperature gradients. The advection associated with cyclones works to destroy temperature gradients, and cyclones that experience more deepening are likely to induce stronger advection. Therefore, a stronger cyclone can easily lead to a weaker AFZ at the 2 m level.
Figure 6A.3. Atmospheric conditions for 0000 UTC 20 July 1981 in the western Arctic, including (a) SLP (hPa; color shading), 500-hPa height (100 m solid contours), and 1000-500-hPa thickness (100 m dashed isopleths); (b) 925-hPa temperature advection (K/day; color shading), SLP (2 hPa isobars), and areas with elevation > 500 m (gray shading); and (c) elevation (500 m filled contours), SLP (2 hPa isobars), and horizontal temperature gradient magnitude (K (100 km)$^{-1}$; red shading). The locations of cyclone centers are marked in black.
Figure 6A.4. Same as Figure 6A.3 but for 0000 UTC 21 July 1981. The green line indicates the track of the newly developed cyclone in the Kolyma Lowland.
Figure 6A.5. Same as Figure 6A.3 but for 0000 UTC 22 July 1981. The green line indicates the track of the cyclone traveling out of the East Siberian Sea.
Building on this interpretation, it might be supposed that the AFZ, at least at the surface, is simply the result of cyclone activity. However, when no cyclone influence is present, such as at 0000 UTC on 20 July for the Kolyma Lowland or all times for Alaska, the AFZ is present at moderate strength. Thermal properties of adjacent land and ocean surfaces, including the amount of snow and sea ice cover, respectively, also play a role (Crawford and Serreze 2015). Therefore, the surface AFZ can and does develop independent of cyclone activity, although passing cyclones can influence its strength near the surface.

Meanwhile, the strength of horizontal temperature gradients above the boundary layer (700 and 500 hPa) exhibit less day-to-day variability, and their relative strength is more likely to reflect lower frequency cycles, such as the larger-scale circulation patterns and seasonal patterns related to the position of the ocean and continents.
APPENDIX TO CHAPTER 7

The depiction of the AFZ in CESM-LE can be compared to reanalysis data in many detailed ways, but the five essential characteristics of the AFZ are its:

1) horizontal location, extent, and intensity,
2) seasonality,
3) vertical expression throughout the lower and middle troposphere,
4) spatial variability, and
5) interannual variability.

Each of these aspects will be considered in turn, with results from CESM-LE compared to MERRA, ERA, and CFSR.

7A.1. Near-Surface Expression

Most prior research has defined and described the AFZ based on surface charts (Reed and Kunkel 1960; Bryson 1966; Krebs and Barry 1970) or near-surface gridded fields, such as at 850 hPa (Lynch et al. 2001; Serreze et al. 2001; Liess et al. 2011) or 2 m (Liess et al. 2011; Crawford and Serreze 2015). The seasonal development of the AFZ is related to contrasts in the surface energy balance between adjacent ocean/ice and continental surfaces (Crawford and Serreze 2015). Therefore, the 2 m level is a logical starting place for comparing CESM-LE’s depiction of the AFZ to the depiction by reanalyses.

Figure 7A.1 shows the mean July 2 m temperature gradient magnitude for the period 1979-2005 for CFSR, MERRA, ERA, and CESM-LE. July was chosen because the AFZ reaches peak intensity in July in these reanalyses (Crawford and Serreze 2015). In addition to the coastline, the average July sea ice extent, northern limit of continuous boreal forest, and southern limit of continuous tundra are marked in black, green, and blue, respectively. CESM-LE is generally consistent with the reanalyses. Strong 2 m temperature gradients are observed along the Arctic Ocean coastline, which separates the warmer air over the snow-free continents from the colder air over the ocean-sea ice surface. Strong 2 m temperature gradients are also observed along mountain ranges (e.g., the Mackenzie Range in Canada and the Verkhoyansk Range in eastern Siberia), and especially surrounding the
Greenland Ice Sheet. For Greenland and the mountain ranges, steep slopes mean that the 2 m gradient is in part capturing a vertical temperature gradient, but along the coastline, these gradients are horizontal. Temperature gradients are weaker along either vegetation boundary, and the strong gradients that lie between these boundaries in some locations demarcate orography (e.g., the Putorana Plateau in central Siberia and the L-shaped Verkhoyansk Range in eastern Siberia).

Figure 7A.1. Mean July 2 m temperature gradient magnitude [K (100 km)⁻¹] for the period 1979-2005 from four datasets. Also marked are the northern limit of continuous boreal forest (green), southern limit of continuous tundra (blue), and average July sea ice extent (black).
CESM-LE’s depiction of the AFZ is distinct in two ways. First, temperature gradients hardly exceed 5 K (100 km)$^{-1}$ in CESM-LE, whereas values exceeding 8 K (100 km)$^{-1}$ are found along the shores of the Beaufort and East Siberian Seas in CFSR and MERRA. Although not part of the AFZ, Greenland and parts of the CAA also show stronger temperature gradients in CFSR and MERRA. This may appear an alarming bias at first, but looking at ERA shows that CESM-LE is not unique in having weaker temperature gradients. In fact, CESM-LE is more similar to ERA than MERRA is to CFSR.

The differences in temperature gradient magnitude amongst CFSR, MERRA, and ERA were also noted by Crawford and Serreze (2015), who concluded that spatial resolution was the main cause for the differences in temperature gradient strength. CFSR has the finest resolution (0.50° latitude by 0.50° longitude) and ERA has the coarsest (0.75° by 0.75°), which means that if two adjacent grid cells have the same temperature difference in both ERA and CFSR, the temperature gradient calculated for CFSR will be greater (exactly how much greater being dependent on latitude and the aspect of the gradient). CESM-LE has a spatial resolution of about 1° latitude by 1° longitude, so it is reasonable, and even expected, that its temperature gradients will be substantially weaker than those calculated in MERRA and CFSR.

If anything, it is surprising that the calculated gradients in CESM-LE are not also noticeably weaker than those of ERA. One reason why not may relate to bias in CESM’s sea ice cover. Figure 7A.2 compares the average 1979-2005 sea ice concentration in (a) summer (JJA) and (b) winter (DJF) to the combined passive microwave record. CESM-LE shows a significant positive bias in average sea ice concentration throughout much of the Arctic Ocean in summer, especially along the coastal seas. Crawford and Serreze (2015) observed that greater sea ice concentration can lead to a strengthening of the summer AFZ. The presence of sea ice increases the ocean’s surface albedo, decreasing that fraction of incoming shortwave radiation that is absorbed. Additionally, absorbed radiation melts the sea ice before contributing to the sensible heat content of the water column, so more summer sea ice means more time before ocean temperatures will rise above the melting point. Since the lower atmosphere is largely warmed from below, this leads to cooler air temperatures over the ocean. CESM-LE seems to capture these processes. Its positive summer sea ice bias is accompanied by a negative near-surface air temperature bias throughout the Arctic Ocean (not shown). Over European Russia, eastern Siberia, and the North Slope of Alaska,
temperatures over land have no temperature bias compared to ERA, or else a positive bias. Therefore, along some parts of the coastline, CESM-LE may be showing a stronger AFZ than it ought to based on its spatial resolution.

Figure 7A.2. Average sea ice concentration in (a) winter (DJF) and (b) summer (JJA) for (left) CESM-LE, (center) the combined passive microwave record, and (right) the difference between them for the period 1979-2005. For differences greater than 1%, stippling indicates significant change based on the 95% confidence interval of a student’s t-test in which each member is considered an independent observation (30 observations per period).

The other distinct aspect of CESM-LE is in the Taymyr Peninsula. For the reanalyses, the AFZ continues to follow the shores of the Kara Sea at these longitudes, but for CESM-LE, the area of strongest temperature gradients is more zonal, cutting across the Taymyr Peninsula instead of following the coast. The area is also broader and the temperature gradients muted compared to west and east of the peninsula. One potential source of difference for these maps is that the reference height in CESM-LE is not a perfect match for the 2 m level in the reanalyses. However, it seems unlikely that this is the cause since the same difference is also observable at near-surface pressure levels like 1000 hPa and 975 hPa (not shown).
Rather, this appears to be a real bias in CESM compared to the reanalyses. Whereas other areas are affected by a positive sea ice concentration bias, the Taymyr Peninsula has a positive bias in snow cover. Based on satellite data, snow cover is nearly absent from the Taymyr Peninsula in July; however, CESM-LE shows intermittent snow cover from about 72°N and northward (Figure 7A.3, top). This lingering snow pack may enforce a more southerly AFZ position by reflecting incoming shortwave radiation and preventing substantial warming of the lower atmosphere. CESM-LE also fixes strong temperature gradients along the southern snow limit in May and June, as shown in the bottom row of Figure 7A.3.
Starting in Section 7A.4, a geographical definition of the AFZ is employed to examine variability in temperature gradient strength throughout time. The AFZ is defined as a two-grid cell thick band extending from the Kola Peninsula (41°E) eastward to the Canadian Arctic Archipelago (234°E, or 126°W). For each longitude in between, the two adjacent grid cells that show the highest average 2 m temperature gradient magnitude north of 60°N are used as the location for the AFZ (the Bering Strait being omitted). For the reanalyses, this definition is coincident with the Arctic coastline; however, for CESM-LE, the AFZ grid cells for the longitudes of the Taymyr...
Peninsula (87°-112°E) lie to the south of the coastline. All subsequent analyses still include these longitudes as part of the AFZ, but it should be noted that the latitude of the AFZ at these longitudes is not directly comparable between CESM-LE and the reanalyses.

Figure 7A.5. As in Figure 7A.1, but for May.
7A.2. Seasonality of the AFZ

Another important aspect of the AFZ is its strong seasonality. In January, when the Arctic is in the midst of winter, atmospheric temperatures fall well below the freezing point, so any leads or polynyas in the sea ice cover provide substantial longwave and turbulent heat fluxes to the frigid atmosphere above (Serreze and Barry 2014). The continents have no such mediating factor, and the air over Siberia, Alaska, and northern Canada becomes colder than the air over the Arctic Ocean. Within the stable inversion layer, strong but shallow temperature gradients develop along the Arctic coastline (Figure 7A.4). Contrary to July, though, the temperature gradients are directed from warm in the north to cool in the south. Also differing from summer, the temperature gradients along the Arctic coastline are not the strongest coastal gradients. Coastlines along the warmer waters of the Norwegian Sea, Gulf of Alaska, and Sea of Okhotsk have much stronger temperature gradients.

As in July, CESM-LE shows a pattern of January temperature gradients that is consistent with the reanalyses, but the intensity of its temperature gradients is generally weaker, especially in comparison to MERRA and CFSR. However, CESM-LE also shows noticeably weaker temperature gradients than ERA around the Arctic Ocean coastline and CAA. Temperature gradients along the sea ice edge in the Greenland Sea and Baffin Bay, on the other hand, are equally as strong as the reanalyses. Part of the reason for this may be that the CESM-LE sea ice cover is generally higher concentration than observations, including in regions that exceed the 15% concentration threshold. Considering Figure 7A.2b, CESM-LE shows a small (1-5%) but significant positive bias in average sea ice concentration throughout the Arctic Ocean in winter. Assuming the slightly lower values in the satellite record reflect real openings in the sea ice cover, this means that CESM provides less opportunity for the ocean to transfer its stored energy into the colder atmosphere, which in January would reduce temperature gradients.
As winter turns to spring and the Arctic begins receiving appreciable shortwave radiation, the cold continents begin to warm more rapidly than the ice-covered Arctic Ocean, destroying the cold-to-the-south temperature gradients of January. The continents begin warming sooner in part because they are farther south and so receive insolation sooner, but they also have a lower albedo. Most of the snow cover in the boreal forest lies below a dark canopy, allowing the forest to absorb more energy than either the tundra or sea ice cover. This heating contrast in turn leads to substantial differences in the sensible heat flux to the lower atmosphere from the
boreal forest and surfaces lying farther north (Hare and Ritchie 1972; Pielke and Vidale 1995). The contrasts between forest and tundra are greatest in the month of May, and for a few weeks in the melt period, the differences in the surface energy fluxes of forest and tundra have been recorded at strengths adequate to induce a frontal zone were they to last longer (Beringer et al. 2001). Looking at the 2 m temperature gradients in May (Figure 7A.5), the Arctic Ocean coastline gradients become neutral, and especially in ERA, the ecotone between the boreal forest and tundra regions shows stronger temperature gradients than in either July or January. CESM-LE also depicts this transition from winter to summer temperature gradient regimes, but in Eurasia, rather than show a strong thermal transition across the gradual ecotone, CESM-LE develops notable temperature gradients narrowly fixed along the tree line. These gradients are also somewhat stronger than those depicted in ERA for the same area, averaging 2-3 K (100 km)$^{-1}$ in CESM-LE along the Eurasian forest limit and 1-2 K (100 km)$^{-1}$ in ERA throughout the Eurasian ecotone. Looking back at Figure 7A.3, the sharper gradients in CESM-LE coincide with the southern snow limit as well as the northern boreal forest limit.

The AFZ first becomes apparent in June (Figure 7A.6). With the snow cover disappearing (or disappeared) from the tundra (Robinson and Frei 2000; Armstrong and Brodzik 2001)$^{30}$ and the sea ice still lingering in the Arctic Ocean (Stroeve et al. 2016), this is the time of greatest albedo contrast across the coastline, and the air over the continents warms rapidly compared to the air over the ocean (Crawford and Serreze 2015). As in every month, the intensity of these temperature gradients is subdued in ERA and CESM-LE, reaching only 3-4 K (100 km)$^{-1}$ on average on the Pacific side of the Arctic Ocean coastline (about 100°E eastward to 234°E (126°W)). AFZ intensity is still greater, however, than the temperature gradients observed around the ecotone or the boreal forest limit. The primacy of the coastline as the most notable source of summer near-surface temperature gradients becomes even more apparent as the AFZ strengthens into July and persists through August.

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$^{30}$ According to both visible and passive microwave satellite datasets, mean monthly snow cover extent from 1978-1999 for the Northern Hemisphere (including Greenland) reaches a maximum of over 40 million km$^2$ in winter and declines to under 10 million km$^2$ by June.
Finally, as summer wanes and coastal sea ice disappears, and thermal contrasts between continents and ocean diminish, and coastal temperature gradients return to neutral in September (Figure 7A.7). Even with a positive sea ice concentration bias, CESM-LE replicates this stage in the seasonal cycle of the AFZ well. In particular, it shows that while the vegetation limits and the coastline lose all prominence as thermal boundaries, the sea ice edge becomes more prominent.
In summary, CESM-LE provides a realistic depiction of the seasonal cycle of near-surface temperature gradients in the high-latitude northern hemisphere. January shows a cold continent and (relatively) warm Arctic Ocean contrast. Cold-to-the-north temperature gradients begin to develop in May, first along vegetation boundaries, and then more strongly along the coastline. The AFZ persists from June through August and quickly diminishes in September. Meanwhile, the sea ice edge is consistently another source of strong near-surface temperature gradients, but it is least prominent in the June-August period, when the AFZ dominates. Some differences are observed between CESM-LE and the reanalyses, especially in the Taymyr Peninsula, but the general depiction of the AFZ is the same in all datasets.

Figure 7A.8. Latitudinal cross sections along 152°E of mean July meridional temperature gradient and zonal wind velocity averaged for the period 1979-2005 from (left) MERRA and (right) CESM-LE. Also marked are the Arctic Ocean coastline (C), the southern boundary of tundra (T), and the Okhotsk Sea coastline (O).
7A.3. Vertical Expression

Thus far, all focus has been on the near-surface expression of the AFZ, but as discussed in Chapter 6, what makes this feature important to synoptic-scale weather patterns is its vertical extent. The vertical expression of the AFZ along 152°E longitude (eastern Siberia) is presented in Figure 7A.8. As shown in Figure 7A.1, this is a particularly strong section of the AFZ. Because of limited space, CESM-LE (b,d) is compared only to MERRA (a,c), the reanalysis with the median spatial resolution of the three examined, as well as the median AFZ intensity. For both datasets, the July average meridional temperature gradient (a,b) and zonal wind velocity (c,d) for 1979-2005 are shown. The meridional gradient is shown instead of gradient magnitude so that the direction of the temperature gradient can be discerned.31

CESM-LE shows better agreement with MERRA at higher latitudes than lower latitudes. The mean polar front position, for instance, is about 42°N for both datasets, but this zone is narrower and the average temperature gradient is stronger in MERRA than in CESM-LE. Since the cross sections are showing climatology, perceived intensity of the polar front is a result of both its actual average intensity and the variability of its location. The intensity of a more mobile polar front will be smeared out across more latitudes, giving the appearance of lower values. This mixed effect is especially likely since the CESM-LE plot is built from 30 times as many observations.32

The AFZ, on the other hand, is nearly identical in the two datasets. In each plot, the average temperature gradients are actually more intense at 72°N than 42°N up to about 700 hPa, with the absolute strongest temperature gradients lying over the coastline. In each plot, although the AFZ is clearly most intense near the surface, a narrow zone of strong meridional temperature gradients that is north of and distinct from the polar front exists all the way up to the tropopause, which sits at a lower altitude so far north. And in each plot, the AFZ extends farther inland than offshore, which may indicate some secondary influence of the nearby southern limit of the tundra augmenting the coastal temperature gradients (Liess et al. 2011).

31 However, at this longitude the AFZ is essentially zonal, so the temperature gradient magnitude is nearly identical.
32 Having 30 times as many observations also explains why CESM-LE contour plots throughout this study are often smoother in appearance than their reanalysis counterparts.
In addition to meridional temperature gradients, CESM-LE also matches well to MERRA in terms of zonal wind velocity. As indicated by the thermal wind equation (Equation 3.3), the strength of the zonal wind aloft is proportional to the intensity of the meridional temperature gradient throughout the troposphere. The polar jet stream is the best-known manifestation of this relationship, but the summer AFZ is also associated with a jet-like feature. The AFZ’s near-zonal orientation, its wide extent over 190° of longitude, its seasonal persistence from June through August, and its reinforcement of the general equator-to-pole temperature gradient conspire to create unique conditions appropriate for the seasonal development of this secondary jet-like feature in the Arctic.

Figure 7A.9. Same as Figure 7A.8, but for January.
Figure 7A.10. Latitudinal cross sections along 206°E (154°W) of mean July meridional temperature gradient and zonal wind velocity averaged for the period 1979-2005 from (left) MERRA and (right) CESM-LE. Also marked are the Arctic Ocean coastline (C), the southern boundary of tundra (T), and the southern shore of Kodiak Island (K).

Making it clearer that CESM-LE follows a similar seasonal cycle as the reanalyses, Figure 7A.9 shows the same cross sections, only for the month of January. In winter, the polar front becomes more prominent, increasing in apparent strength, becoming narrower, and shifting southward by about 10° latitude. Correspondingly, the polar jet intensifies and shifts southward. Another point of consistency is that a distinct subtropical jet, which CESM-LE showed more prominently than MERRA in July, is absent from both cross sections in January.
Finally, the AFZ is conspicuously missing from the January cross sections, and not just at the surface.

Meridional temperature gradients are slightly negative or neutral throughout most of the troposphere north of 60°N. The one exception is close to the surface, where temperature gradients are positive (cold to the south) from near the apex of the Verkhoyansk Range to the coastline. Cold-to-the-south temperature gradients counteract the general equator-to-pole temperature gradient and would actually induce easterly winds aloft if they were able to persist above the stable inversion layer, so no westerly jet-like feature exists in winter for CESM-LE, as with the reanalyses.

Figure 7A.11. Latitudinal cross sections along 280°E (80°W) of mean July meridional temperature gradient and zonal wind velocity averaged for the period 1979-2005 from (left) MERRA and (right) CESM-LE. Also marked are the northern shore of Ellesmere Island (E) and the coast of James Bay (J).
Looking at cross sections through other longitudes of the AFZ, such as 206°E (154°W) through Alaska (Figure 7A.10), tell a similar story. CESM-LE reproduces the same basic temperature gradient and zonal wind patterns as the reanalyses, and the AFZ is actually one of the most consistently depicted features. Also consistent is that along longitudes that lack a distinct boundary between the Arctic Ocean and the continents, no AFZ is apparent. A good example of this is along 280°E (80°W), through the CAA (Figure 7A.11). Between 60°N and 85°N, the surface alternates between islands and straits and inlets. There is no clear and narrow boundary between warmer continental air and cooler Arctic Ocean air, and no AFZ is present in the reanalyses. CESM-LE, which is coarser resolution than MERRA, shows less variability in the temperature gradients at these latitudes but likewise depicts no AFZ. Therefore, in every aspect of the vertical expression of the AFZ described by Crawford and Serreze (2015), CESM-LE is consistent with the reanalyses.

7A.4. Spatial Variability of the AFZ

7A.4.1. Near the Surface

Crawford and Serreze (2015) found that although in the climatology the AFZ develops as a single coherent feature each summer, variability in AFZ strength from one summer to the next is heterogeneous. This behavior is apparent examining the monthly July anomalies in AFZ strength for almost any year, such as 1988 (Figure 7A.12). In 1988, the AFZ was much weaker than normal along the Taymyr Peninsula and the Canadian Beaufort Sea coast. However, it was stronger than normal along most of eastern Siberia and Chukotka.

By performing a cluster analysis, Crawford and Serreze (2015) found that this spatially heterogeneous variability is systematic and that the summer AFZ can be divided into several distinct sectors, each of which experiences interannual variability separately from the others. These sectors are shown in Figure 7A.13, which depicts the clustering results from each of the three reanalyses based on Ward’s method. Results using k-means are comparable. The reanalyses disagree on how many sectors should divide the AFZ, with ERA indicating nine, MERRA seven, and CFSR only six. However, all sector boundaries in CFSR and MERRA results are within two grid cells of the sector boundaries in ERA, showing strong consistency in where the statistically meaningful divisions exist. Also notable is that the sector boundaries align with physically meaningful disruptions to the Arctic coastline,
such as Bering Strait (between sectors 7 and 8), the terminus of the Ural Mountains (1 and 2) and Verkhoyansk Range (4 and 5), and Khatanga Bay (3 and 4). With two clustering methods and three reanalyses yielding similar results, Crawford and Serreze (2015) concluded that it is often more appropriate to consider interannual variability of the AFZ in terms of these regional sectors as opposed to treating the feature as a single unit.

Figure 7A.12. July 1988 2 m temperature gradient anomaly (K [100 km]⁻¹) for (a) CFSR, (b) MERRA, and (c) ERA.
Figure 7A.13. Results of Ward’s method cluster analysis for (a) CFSR, (b) MERRA, and (c) ERA. The grid cells used for each cluster analysis are colored to represent their clustering assignment. The white shading is the mean July sea ice extent for the period 1979-2014. (d) A schematic relates the clusters in each analysis to the nine sectors identified by Crawford and Serreze (2015).

Figure 7A.14 is a schematic comparing the consensus sector breaks from the three reanalyses to the sector breaks identified in CESM-LE data using the same Ward’s clustering method. The height of each column shows how many members identify that longitude as a sector break when the clustering analysis is performed on each individual member. The black dots along the top of the plot show the sector breaks when the clustering analysis is performed on all 30 members together (treating CESM-LE as though it is 810 observations of a persistent 1979-2005 climate instead of 30 separate sets of 27 observations each).

The degree of internal variability represented in CESM-LE is clearly shown in this Figure 7A.14. Only one longitude is a sector break for every CESM-LE member, and over 40 different longitudes are a sector break for at least one member. The divisions in eastern Siberia between sectors 4 through 7 are especially muddled, with no single break point identified in more than 10 out of 30 members. This is the same stretch of coastline that shows the least amount of agreement amongst the reanalyses (Figure 7A.13). Although ERA indicates the presence of
four separate sectors here, MERRA indicates three, and CFSR indicates two. Moreover, the one sector boundary identified in CFSR is not shared by MERRA. All of the breaks in MERRA and CFSR are within two grid cells of a break in ERA, but as represented by the gaps in the colored rectangles in Figure 7A.14, there is not perfect agreement on exactly where the sector breaks occur. To account for this, Crawford and Serreze (2015) omitted grid cells without certain sector assignment from further analysis. If that same tactic is taken with CESM-LE, the three boundaries between sectors 4 through 7 have better agreement (the minimum number of members being 12 out of 30).

Besides Siberia, strong agreement exists for the placement of sector boundaries. Baydaratskaya Bay and the terminus of the Ural Mountains is a clear boundary between sectors 1 and 2. The boundaries for sector 3 are somewhat constricted in CESM-LE, but this can be traced directly back to the more southerly latitude of the AFZ across the Taymyr Peninsula. Rather than Pyasina Bay, the boundary of sectors 3 and 2 is marked by where the AFZ switches from coastal to continental. Khatanga Bay is clearly the boundary between sectors 3 and 4, the small shift observed being a result of different grid cell alignments. The Bering Strait is the most common division between sectors 7 and 8, but several CESM-LE members actually include a few grid cells from the western tip of Alaska in
sector 7, although none east of Icy Cape. Finally, the division between sector 8 and 9 is most often shifted by one grid cell to the east.

**Figure 7A.15.** Mean July 700 hPa temperature gradient magnitude [K (100 km)$^{-1}$] for the period 1979-2005 from four datasets. Also marked are the northern limit of continuous boreal forest (green), southern limit of continuous tundra (blue), and average July sea ice extent (black).

If all CESM-LE members are combined into a cluster analysis with 810 observations, the preferred number of sectors jumps to 11 instead of the nine (based on average silhouette width). However, nine of those 11 sectors match very closely with the divisions identified by the reanalyses. Especially notable is that the three divisions in
eastern Siberia line up perfectly. The other two extra divisions line up with the Poluostrov Peninsula and Yenisei Bay, two other prominent disruptions to the Arctic Ocean coastline.

The clustering analyses show the spatial heterogeneity of the AFZ’s interannual variability is similarly systematic in both the reanalyses and CESM-LE. Especially when combining all members into a single analysis, the resulting sector divisions from ERA are both physically meaningful and replicated by CESM-LE. These similarities justify the use of the same sector definitions for CESM-LE as the reanalyses, which will make subsequent results easier to compare to the findings of Crawford and Serreze (2015). Using a consistent set of sectors also makes comparisons between the various CESM-LE members more straightforward.

7A.4.2. In the Mid-Troposphere (700 hPa)

With increasing altitude above the surface, variability in the AFZ becomes less responsive to variations at the surface. For instance, the climatological position of the strongest horizontal temperature gradients at 700 hPa lies north of the coastlines of the Barents and Chukchi Seas, conforming to a more zonal pattern that matches the mean coastal latitude instead of the precise coastline (Figure 7A.15). Near the surface, the strongest temperature gradients north of 60°N are co-located with the coastline not only in their climatology, but also in nearly any given summer month. However, higher up in the troposphere, the AFZ and its attendant jet-like feature behave more like the polar jet to the south, exhibiting substantial ridges and troughs from month to month even though the mean position and seasonal development appear tied to energy balance differences between the continental and oceanic surfaces (as discussed in Chapter 3). Therefore, whereas year-to-year variability in the summer AFZ at the surface is well described by a cluster analysis of temperature gradient variability at fixed locations along the coast, at 700 hPa the AFZ may be better described as a sinuous band of particularly strong temperature gradients that exhibits monthly variability in both intensity and position.
Figure 7A.16. Mean latitude of the migrating Arctic Front (defined as grid cells within the area bounded by 42°E to the west, 234°E (126°W) to the east, 60°N to the south, and 88°N to the north that exceed the 90th percentile of horizontal temperature gradient strength at 700 hPa) for (a-c) the summer months and (d) the combined summer season for MERRA (red; 1979-2014), ERA (blue; 1979-2014), and all 30 CESM-LE members (green; 1979-2005).

An alternative metric called the MAF (the “migrating Arctic Front”) was devised to capture this combination of spatial and intensity variability. The MAF is defined as the grid cells within the bounding box of 42°E to the west, 234°E (126°W) to the east, 60°N to the south and 88°N to the north that exceed the 90th percentile of horizontal temperature gradient magnitude for a given level within those bounds. Since the value of the 90th percentile may vary with each field examined, MAF strength can be measured as the mean temperature gradient magnitude of those grid cells. The location of the MAF can also be assessed by, for example, recording the average latitude of MAF grid cells each month at the 700 hPa level (Figure 7A.16). The mean latitude of the Arctic Ocean coastline is about 71°N, which compares especially well with the mode of average MAF latitude in MERRA (red) and ERA (blue) in July, August, and JJA combined. In other words, these results agree with the position of the climatological AFZ, as described by Reed and Kunkel (1960), Crawford and Serreze (2015) and others. In June, the
histograms for all three datasets are shifted south of the coastline, especially CESM-LE and MERRA. This mean position around 69°N may reflect the lingering presence of snow cover along the coasts, especially of the Taymyr Peninsula.

However, there is some variation detected from the most common position. CESM-LE results are more likely than either reanalysis to show the strongest temperature gradients south of 71°N, especially in July and August. On the other hand, the two reanalyses are more likely than CESM-LE to show the strongest temperature gradients north of 71°N (over the Arctic Ocean). This discrepancy is also visible in Figure 7A.15, where the climatological temperature gradient magnitude for CESM-LE is greater than the reanalyses over much of Siberia and lower than the reanalyses over most of the Arctic Ocean in July. Although these differences appear minor, they may relate to differences in Arctic cyclone development between the reanalyses and CESM-LE, as discussed later in Section 8A.2.

7A.5. Interannual Variability of the AFZ

Sectoralized interannual variability of monthly near-surface AFZ strength from 1979-2005 for both CESM-LE and the reanalyses is shown in Figure 7A.17. Meridional temperature gradients are shown to highlight the seasonal cycle of positive gradients in winter and negative gradients in summer. Each CESM-LE box plot is built from 810 observations (27 years for each of 30 members), whereas each reanalysis box plot is built from 81 observations (27 years for each of 3 reanalyses). The boxes mark the median and interquartile range (IQR), while the whiskers extend to the most extreme observations within $1.5 \times$ IQR. The data are divided by AFZ sector, as defined in Section 7A.4.

As indicated in Section 7A.2, CESM-LE shows the same seasonal cycle of AFZ development as the reanalyses. Temperature gradients are cold-to-the-south in winter, become very weak in spring, and then rapidly turn to cold-to-the north in summer before diminishing again in September. In most cases, CESM-LE is a better match with the reanalyses in June and July, when the AFZ is strongest. The interquartile range of the boxplots overlaps in 16 of 18 instances, the exceptions being sectors 2 and 7 in June. In these sectors, the reanalyses, especially CFSR and MERRA, show stronger temperature gradients in June than does CESM-LE.
Figure 7A.17. Seasonality of near-surface AFZ strength based on monthly means (1979-2005) of nine AFZ sectors for both CESM-LE (gray) and atmospheric reanalyses (blue). CESM-LE box plots are constructed from 810 observations (27 years and 30 members), whereas reanalysis box plots are constructed from 81 observations (27 years and 3 reanalyses). Each box contains the median and first and third quartiles; the whiskers extend to the most extreme observation within 1.5 times the interquartile range (outliers beyond that point being omitted).

The most drastic difference between CESM-LE and the reanalyses is sector 3, which comprises the Taymyr Peninsula. Temperature gradients are very weak in CESM-LE throughout winter, whereas the reanalyses show a switch to positive gradients. This bias occurs because CESM-LE observations are from the interior of the peninsula instead of the coastline. Unlike the other eight sectors, sector 3 does not provide a direct comparison of the same exact locations between CESM-LE and the reanalyses. However, since the sector definition is based on July temperature gradients, CESM-LE is a good match with the reanalyses for the summer Taymyr Peninsula.

CESM-LE also shares several other notable features with the reanalyses, such as the spatial variability in both median AFZ strength and the timing of AFZ development. For instance, in both CESM-LE and the reanalyses, the median July AFZ strength is stronger than -4 K (100 km)$^{-1}$ for sectors 5, 6, 8, and 9, about -4 K (100 km)$^{-1}$ for sector 4, and weaker than -4 K (100 km)$^{-1}$ for sectors 1-3 and 7. In general, the sectors with stronger peak AFZ strength are also the sectors with earlier onset of cold-to-the-north temperature gradients, but in both CESM-LE
and the reanalyses, sector 1 is a notable exception. Bordering the Barents Sea, this sector is the most influenced by the North Atlantic and lies the farthest south. It loses its snow cover early, allowing for earlier warming of continental air than for other sectors, but it also is the one sector without a significant summer sea ice cover, which diminishes the thermal contrast between air overlying the continent and ocean.

Finally, these box plots offer some intriguing insights regarding both interannual variability and spread amongst CESM-LE members versus spread amongst the reanalyses. Focusing on summer in particular, the IQR for CESM-LE is comparable to the IQR for the reanalyses, and more often than not the IQR for CESM-LE is actually smaller. This appears largely due to greater variability in CFSR. Comparing the standard deviation of near-surface July AFZ strength in each CESM-LE member to that of the reanalyses, CFSR has a standard deviation greater than the maximum CESM-LE members for eight out of nine sectors. By contrast, ERA and MERRA show a greater standard deviation than the maximum CESM-LE member only for sectors 5 and 6. In other words, interannual variability of July AFZ strength in ERA and MERRA is actually more similar to CESM-LE than to CFSR.
APPENDIX TO CHAPTER 8

This appendix compares the Arctic cyclone regime in CESM-LE to that of MERRA and ERA. Most of the methods used and figures shown in this section are repeated from Chapter 6 only with different data sets. Although results are broadly similar, some notable differences exist between the MERRA data used in Chapter 6 and in this appendix. First, data are only taken from 1990-2005 instead of 1979-2014 in order to match the period for CESM-LE. This leads to aggregated fields that are less smooth and wider confidence intervals for statistical tests.

Second, although MERRA data is available at a 3-hr temporal resolution, data are examined only every 6-hr for this appendix to match the temporal resolution of CESM-LE. The change in temporal resolution has a greater impact on results. Most notably, finer temporal resolution increases the accuracy the cyclone tracking and makes the continuation of a cyclone track from one observation time to the next more likely (Blender and Schubert 2000). Since track continuation is more likely, cyclone tracks are more likely to exceed the minimum track length required during post-detection filtering. This leads to slightly higher track density with finer temporal resolution (Blender and Schubert 2000). However, as shown in the Appendix to Chapter 5 (e.g., Table 5A.3), differences between various datasets are greater than differences between 3-hr MERRA and 6-hr MERRA, so comparisons between temporal resolutions are rarely made throughout the following sections.

8A.1. Basic Cyclone Characteristics of CESM-LE

8A.1.1. An Initial Look at Northern Hemisphere Cyclone Frequency

Before focusing on the Arctic region, it is worthwhile to compare the underlying input data for the cyclone and detection algorithm: the instantaneous SLP fields. More specifically, CESM-LE SLP fields have higher frequency spatial variation (i.e., more noise) than their counterparts in MERRA and ERA. Generally speaking, a SLP field from CESM-LE has more peaks and depressions, more maxima and minima than a SLP field from MERRA or ERA in the same season. As shown in Figure 8A.1 this difference is especially great in summer, when CESM-LE fields have on
average 45% more SLP minima than ERA fields and 63% more than MERRA fields. In winter, the difference is about the same with ERA (44% more minima in CESM-LE fields) but much reduced with MERRA (only 25% more).

**Figure 8A.1.** Histograms of the average number of minima per SLP field, aggregated by month, for summer (JJA; top) and winter (DJF; bottom) 1990-2005 using CESM-LE (black), MERRA (red), and ERA (blue). Since density is used for the y-axis, the total area covered by each set of columns equals one.

Having more SLP minima means that CESM-LE has the potential for greater cyclone frequency, but the cyclone algorithm has several steps between the detection of minima and the calculation of track density that help to mitigate these initial differences. **Table 8A.1** displays comparisons between CESM-LE and the reanalyses for six major steps in the process, from detecting minima, to cyclone centers, to cyclone tracks. The second step in the
algorithm attempts to filter the detected SLP minima so that the only ones remaining represent true synoptic-scale cyclones. Some resulting cyclone centers are then grouped together into multi-center cyclones (MCCs). These steps go a long way toward reducing the differences between CESM-LE and the reanalyses. In summer, CESM-LE fields have on average 19% and 25% more cyclones detected than MERRA and ERA fields, respectively. In winter, MERRA and CESM-LE have essentially the same number of cyclones per field, meaning no significant bias exists.

However, it is important to note that both center filtering and MCC detection are imperfect processes, as exemplified in Figure 8A.2, which shows an example SLP field with cyclone centers and areas marked for each dataset. Although the time is the same for each (2 June 1990 at 1200 UTC), the cyclones in the CESM-LE field do not directly correspond to those in MERRA and ERA because CESM is not forced by observations. These fields are only meant as illustrative examples. For this time, both CESM-LE and the reanalyses show a case of the MCC detection failing to combine multiple centers that by visual inspection appear to be part of the same cyclone. In CESM-LE, five cyclone centers are detected along the North Atlantic storm track, but they likely only represent two different systems: one east of Greenland and one southwest of Greenland. In the reanalyses, four cyclone centers are detected around Scandinavia, but none are combined to form a MCC. The algorithm fails in these complex cases, which exhibit a large number of centers with large differences in central pressure.

Table 8A.1. Average cyclone detection and tracking counts for each dataset. The 95% confidence interval for CESM-LE is also reported. “Valid” tracks have a track length of at least 100 km, a lifespan of at least 24 hr, and cross at least one grid cell with an elevation less than 500 m. (Only valid tracks are used in the construction of other figures and tables unless otherwise noted.)

<table>
<thead>
<tr>
<th>Variable</th>
<th>MERRA</th>
<th>ERA</th>
<th>CESM-LE</th>
<th>C/M</th>
<th>C/E</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Summer (JJA)</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1 Minima Per SLP Field</td>
<td>176.6</td>
<td>199.1</td>
<td>287.8 ± 1.5</td>
<td>1.63</td>
<td>1.45</td>
</tr>
<tr>
<td>2 Cyclone Centers Per SLP Field</td>
<td>19.4</td>
<td>18.5</td>
<td>23.4 ± 0.8</td>
<td>1.21</td>
<td>1.27</td>
</tr>
<tr>
<td>3 Cyclones Per SLP Field</td>
<td>17.9</td>
<td>17.0</td>
<td>21.3 ± 3.6</td>
<td>1.19</td>
<td>1.25</td>
</tr>
<tr>
<td>4 Center Tracks Per Season</td>
<td>1660.0</td>
<td>1442.4</td>
<td>2149.0 ± 33.4</td>
<td>1.29</td>
<td>1.49</td>
</tr>
<tr>
<td>5 Cyclone Tracks Per Season</td>
<td>1451.4</td>
<td>1258.8</td>
<td>1887.4 ± 28.8</td>
<td>1.30</td>
<td>1.50</td>
</tr>
<tr>
<td>6 Valid Cyclone Tracks Per Season</td>
<td>379.8</td>
<td>337.9</td>
<td>402.3 ± 8.7</td>
<td>1.06</td>
<td>1.19</td>
</tr>
<tr>
<td><strong>Winter (DJF)</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1 Minima Per SLP Field</td>
<td>179.1</td>
<td>155.5</td>
<td>224.4 ± 1.6</td>
<td>1.25</td>
<td>1.44</td>
</tr>
<tr>
<td>2 Cyclone Centers Per SLP Field</td>
<td>24.2</td>
<td>20.4</td>
<td>24.6 ± 1.6</td>
<td>1.01</td>
<td>1.20</td>
</tr>
<tr>
<td>3 Cyclones Per SLP Field</td>
<td>22.6</td>
<td>18.8</td>
<td>22.7 ± 0.5</td>
<td>1.00</td>
<td>1.21</td>
</tr>
<tr>
<td>4 Center Tracks Per Season</td>
<td>2066.3</td>
<td>1607.7</td>
<td>2257.3 ± 63.4</td>
<td>1.09</td>
<td>1.40</td>
</tr>
<tr>
<td>5 Cyclone Tracks Per Season</td>
<td>1878.3</td>
<td>1443</td>
<td>2037.8 ± 57.1</td>
<td>1.08</td>
<td>1.41</td>
</tr>
<tr>
<td>6 Valid Cyclone Tracks Per Season</td>
<td>449.4</td>
<td>422.8</td>
<td>456.8 ± 10.8</td>
<td>1.02</td>
<td>1.08</td>
</tr>
</tbody>
</table>
Figure 8A.2. An example of a typical SLP field (2 Jun 1990 1200 UTC) with cyclone centers and areas marked and in CESM-LE member 20, MERRA, and ERA. Areas with elevation exceeding 1500 m are masked. Note than while both ERA and MERRA are forced by observations, CESM-LE is not. The reanalysis fields show corresponding weather conditions, while the CESM-LE example shows an entirely independent weather scenario.

The tracking stage adds more complexity to the identification of cyclones. Isolated cyclone centers are the easiest to track because little ambiguity exists regarding which centers in consecutive times correspond to each other. Cyclone centers in high density areas, on the other hand, can result in complicated situations. In summer, CESM-LE exhibits more cyclone centers than either reanalysis, increasing the number of ambiguous tracking
situations. As is the case for the example in Figure 8A.2, these extra cyclone centers are also often clustered near other cyclones, as opposed to being dispersed or randomly placed. (Notice how the CESM-LE field has only three isolated cyclones, located in Baffin Bay, western Russia, and central Siberia.) When attempting to resolve ambiguous tracking situations with congested cyclone centers, tracking mistakes become more common, which leads to premature lysis of some cyclone tracks (Mesquita et al. 2009; Rudeva et al. 2014). In aggregate statistics, this is manifested as inflated track density and shorter tracks. As shown in Table 8A.2, the population of cyclones detected in CESM-LE has an average lifespan and average track length that are both shorter than for the cyclone populations in either MERRA or ERA, corroborating the idea that more congested cyclone centers in CESM-LE explain, at least in part, the increased difference between CESM-LE and the reanalyses from step 3 to step 4.

Limiting tracks to one for each MCC (step 5) makes little difference to the ratios between CESM-LE and the reanalyses, but the final step, including filtering by elevation, track length, and life span, substantially reduces the differences for summer, as well as for CESM-LE and ERA in winter. CESM-LE still shows a significantly higher number of tracks per season than ERA in both seasons and than MERRA in summer; however, the variation amongst these three datasets is less than the variation amongst different cyclone detection and tracking methods using the same data (see Table 2 in Neu et al. (2013)). Therefore, although CESM-LE SLP fields clearly contain notable differences in terms of the number of minima and initial cyclone tracking, the filters in place in the detection and tracking algorithm go a long way toward making results from the different datasets more comparable.

<table>
<thead>
<tr>
<th>Variable</th>
<th>MERRA</th>
<th>ERA</th>
<th>CESM-LE</th>
</tr>
</thead>
<tbody>
<tr>
<td>Summer (JJA)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Lifespan (days)</td>
<td>1.36</td>
<td>1.35</td>
<td>1.14 ± 0.01</td>
</tr>
<tr>
<td>Track Length (km)</td>
<td>1171.91</td>
<td>1102.26</td>
<td>1012.12 ± 12.61</td>
</tr>
<tr>
<td>Winter (DJF)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Lifespan (days)</td>
<td>1.30</td>
<td>1.35</td>
<td>1.12 ± 0.01</td>
</tr>
<tr>
<td>Track Length (km)</td>
<td>1440.89</td>
<td>1510.76</td>
<td>1291.49 ± 13.63</td>
</tr>
</tbody>
</table>
8A.1.2. Other Hemisphere-Wide Cyclone Characteristics

In addition to cyclone track frequency, other aspects of a cyclone’s track, including lifespan, track length, maximum deepening rate, minimum central pressure, and maximum Laplacian of central pressure are aggregated for each dataset and season in Table 8A.3. The average minimum central pressure for cyclone tracks is marginally higher in CESM-LE than the reanalyses in both seasons, but CESM-LE shows good agreement for the Laplacian of central pressure. The discrepancy between these two intensity measures likely comes from a positive background SLP bias throughout the continents and Atlantic Ocean in winter and everywhere but Europe and central Asia in summer (not shown). This bias only affects the central pressure, so the Laplacian is the truer comparison of intensity.

Table 8A.3. Average cyclone track characteristics for each dataset. All tracks included in the averages have a lifespan of at least 24 hr and a track length of at least 100 km and spend at least one observation time at elevations less than 500 m. The 95% confidence interval for CESM-LE is also reported.

<table>
<thead>
<tr>
<th>Variable</th>
<th>MERRA</th>
<th>ERA</th>
<th>CESM-LE</th>
</tr>
</thead>
<tbody>
<tr>
<td>Summer (JJA)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Tracks Per Season</td>
<td>337.94</td>
<td>345.25</td>
<td>402.30</td>
</tr>
<tr>
<td>Lifespan (days)</td>
<td>3.33</td>
<td>3.33</td>
<td>3.32</td>
</tr>
<tr>
<td>Track Length (km)</td>
<td>2482.48</td>
<td>2396.50</td>
<td>2362.90</td>
</tr>
<tr>
<td>Maximum Deepening Rate (hPa day⁻¹)</td>
<td>-14.12</td>
<td>-14.87</td>
<td>-11.97</td>
</tr>
<tr>
<td>Minimum Pressure (hPa)</td>
<td>995.41</td>
<td>995.13</td>
<td>997.98</td>
</tr>
<tr>
<td>Maximum Laplacian (hPa [100 km]⁻²)</td>
<td>2.18</td>
<td>2.36</td>
<td>2.24</td>
</tr>
<tr>
<td>Winter (DJF)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Tracks Per Season</td>
<td>449.38</td>
<td>422.75</td>
<td>456.76</td>
</tr>
<tr>
<td>Lifespan (days)</td>
<td>3.08</td>
<td>3.04</td>
<td>3.09</td>
</tr>
<tr>
<td>Track Length (km)</td>
<td>2949.90</td>
<td>3024.27</td>
<td>3033.26</td>
</tr>
<tr>
<td>Maximum Deepening Rate (hPa day⁻¹)</td>
<td>-20.56</td>
<td>-20.18</td>
<td>-18.21</td>
</tr>
<tr>
<td>Minimum Pressure (hPa)</td>
<td>986.97</td>
<td>987.96</td>
<td>989.73</td>
</tr>
<tr>
<td>Maximum Laplacian (hPa [100 km]⁻²)</td>
<td>3.00</td>
<td>3.07</td>
<td>3.01</td>
</tr>
</tbody>
</table>
Figure 8A.3. Cyclone track density for (a) summer (JJA) and (b) winter (DJF) averaged for the period 1990-2005 in CESM-LE, MERRA, and ERA, as well as the differences between CESM-LE and the two reanalyses. Difference based on a 95% confidence interval. Units are tracks per 250,000 km$^2$ per season.
After limiting tracks to those that have a lifespan of at least 24 hr and a track length of at least 100 km and spend at least one observation time at elevations less than 500 m, the lifespan of cyclones is about the same for each dataset in both seasons, and CESM-LE and ERA have similar track lengths. Compared to ERA and CESM-LE, MERRA cyclone tracks tend to be longer in summer and shorter in winter. Finally, the average maximum deepening rate for cyclone tracks is significantly smaller in CESM-LE than in either reanalysis for both seasons. This may in part be an effect of the biases in the underlying SLP fields, but a simple shift in SLP should not affect the rate of change in SLP. Therefore, deepening rate in CESM-LE should be viewed with more caution than the more robust cyclone characteristics, such as lifespan or the Laplacian of central pressure. Overall, CESM-LE presents reasonable values for the Northern Hemisphere, which justifies comparisons with the reanalyses at regional levels.

8A.1.3. Regional Track Density

Figure 8A.3a shows the average track density during summer (JJA) months for CESM-LE, MERRA, and ERA on the top row and the difference between CESM-LE and each reanalysis on the bottom row. In the difference maps, red indicates that more cyclone tracks are observed in CESM-LE; blue indicates fewer. Difference based on a 95% confidence interval.

All three datasets show very similar spatial patterns over the oceans, including the three key maxima for cyclone activity. First, the greatest track density is observed in the North Atlantic storm track (over 6 tracks per month). Second, the North Pacific storm track is present but substantially weaker (mostly 4-6 tracks per month). Third, the CAO is a relative maximum; its track density being comparable to the storm track in the Pacific. The North Atlantic storm track is denser in CESM-LE around the Gulf of St. Lawrence, along the southeast coast of Greenland, and in the northern Norwegian Sea. The start of the Pacific storm track (around the Sea of Japan) is denser in CESM-LE, but its terminus in the Gulf of Alaska is much stronger in the reanalyses. The reanalyses show greater track density in the CAO, the Canadian Arctic Archipelago, and along the coast of the Barents and Kara Seas. The greatest differences, however, lie over high elevation areas, especially in eastern Siberia and northwestern North America, where CESM-LE exhibits about twice as many storm tracks as do the reanalyses.
Figure 8A.4. As in Figure 8A.3, but for cyclogenesis frequency.
Figure 8A.5. As in Figure 8A.3, but for cyclysis frequency.
Figure 8A.6. Cyclogenesis frequency for the subset of cyclones whose tracks intersect the CAO+BCEL region for (a) summer (JJA) and (b) winter (DJF) averaged for the period 1990-2005 in CESM-LE, MERRA, and ERA, as well as the differences between CESM-LE and the two reanalyses. Difference based on a 95% confidence interval. Units are tracks per 250,000 km$^2$ per season.
In winter (DJF; Figure 8A.3b), the general patterns are again the same for all datasets. Compared to summer, cyclone activity is much reduced over the continents and strengthened in the two main storm tracks. Track density maxima in the Gulf of Alaska, the north end of Baffin Bay, and in the Icelandic Low region are all prominent in both the reanalyses and CESM-LE. Track density is greater in winter on the Atlantic side of the Arctic Ocean but lesser for the Pacific side. As in summer, CESM-LE shows fewer cyclones tracking into the Kara Sea and Gulf of Alaska and more cyclones around Kamchatka and along the southeast Greenland coast. It also has greater winter track density southwest of Greenland, throughout the Icelandic Low region, and in the Beaufort Sea. Also similar to summer, the areas with greatest differences between CESM-LE and the reanalysis are also generally areas with high track density in all datasets, although the Beaufort Sea and north of Greenland are notable exceptions to this pattern. These areas have fewer than four tracks per winter month for the reanalyses.

CESM has some of the same biases as other models. The HadGAM1 model for instance, also underestimate Gulf of Alaska cyclone frequency in winter (Greeves et al. 2007). However, CESM appears to improve on the orientation of the North Atlantic storm track. Most CMIP3 and CMIP5 models (including the Community Climate System Model version 4, CESM’s predecessor), typically show an Atlantic storm track that is too zonal, with too many cyclones over central Europe and too few in the Norwegian and Beaufort Seas (Ulbrich et al. 2008; Zappa et al. 2013). CESM-LE shows fewer cyclones at the extreme end of the North Atlantic storm track, but it actually overestimates track density in the Norwegian Sea. Underestimation of cyclone track density in the CAO is also a common bias in climate models (Zappa et al. 2013; Nishii et al. 2015).

8A.1.4. Regional Cyclogenesis & Cyclolysis

Maps of cyclogenesis (Figure 8A.4) and cyclolysis (Figure 8A.5) show a consistent story with track density: CESM-LE matches the general spatial patterns and seasonal contrasts as the reanalyses, but it has a tendency to yield more cyclone tracks in high elevation areas in summer and shows differences in several other high density areas for both seasons. Some of the differences reveal the reasons behind the differences in track density observed in Figure 8A.3. For instance, CESM-LE shows excessive cyclolysis in Scandinavia and deficient cyclolysis along the Barents and Kara Sea coastlines in summer. This suggests that the Atlantic storm track is truncated in
CESM-LE, with more storms slamming into Scandinavia instead of tracking northward in the Arctic seas. In other words, CESM-LE still exhibits some “zonalization” of the North Atlantic storm track, but it is much less than previous models showed for winter (see Figure 1 in Ulbrich et al. (2008) and Figure 2 in Zappa et al. (2013)). Excessive summer track density around Labrador in CESM-LE can be attributed to an increased tendency for cyclogensis in that region. Both Siberia and the North American cordillera generate many more cyclones each summer in CESM-LE than either reanalysis, and in neighboring regions due east of these areas, such as the Bering Sea, cycloysis is exceptionally high in CESM-LE.

In winter, differences in cyclogenesis and cycloysis are patchier. The Barents and Kara Seas witness less cycloysis in CESM-LE, but they also experience significantly less genesis. This suggests that unlike summer, where a shifted storm track can be blamed, the differences in winter are the result of local generation of cyclones in the reanalyses. Again, the zonalization issue is much reduced compared to other models. One similarity with other models is that CESM-LE shows less winter cyclogenesis in the lee of the Rocky Mountains (Ulbrich et al. 2009).

Local generation of cyclones over the Arctic Ocean is often linked to the presence of open water, which increases atmospheric instability by enhancing upward turbulent heat fluxes (Ledrew 1984; Overland and Wang 2010; Long and Perrie 2012; Jaiser et al. 2012). Figure 7A.2b shows that mean winter sea ice concentration in CESM-LE is higher for much of the Barents Sea, which corresponds to the area with lower cyclogenesis than the reanalyses. Although the difference is small, greater sea ice cover could encourage more stability and less cyclogenesis by inhibiting upward energy fluxes.

A similar argument cannot, however, be made for the bias observed in the Beaufort Sea. In this area, CESM-LE shows a positive bias for track density despite the sea ice concentration being similar to or greater than that of the reanalyses. However, MERRA and ERA also have notable differences in this area. Unlike MERRA, ERA shows a similar local maximum in cyclogenesis to CESM-LE near the Beaufort Sea coast just west of the USA-Canada border. It is simply weaker in ERA. Therefore, it is possible that the depiction of this region in CESM-LE is actually more accurate than in MERRA. However, excessive cyclone frequency along the Alaskan coast has been noted in other models without a certain explanation (Ulbrich et al. 2009; Zappa et al. 2013).
Summer sea ice concentration is biased high in CESM-LE (Figure 7A.2a), and the CAO has some negative bias for cyclogenesis (Figure 8A.4a). However, turbulent energy fluxes are often directed into the Arctic Ocean during summer, so a bias in sea ice cover is less likely to impact cyclone development in summer than in winter.

8A.2. Arctic Cyclone Origins

With the more general differences in the depiction of cyclones between CESM-LE and the reanalyses described, focus now turns to the origin of cyclones that enter the CAO+BCEL domain. Figure 8A.6 shows the average cyclogenesis frequency for cyclones that at any point in their lifespan enter the CAO+BCEL region for a) summer and b) winter. Depending on the ensemble member, CESM averages 42 to 50 cyclones entering or forming in the CAO+BCEL region each summer. ERA (42) and MERRA (44) fall on the lower end of this range. All three datasets show that many of the cyclones entering the CAO+BCEL region are externally sourced, especially from central and eastern Siberia. They also show that some CAO+BCEL cyclones are locally sourced, forming over the Arctic Ocean. These findings are consistent with Serreze and Barrett (2008), who showed that most cyclones entering the CAO originate either locally or over Eurasia.

However, despite these broad similarities, CESM-LE exhibits two distinct and significant differences from the reanalyses. First, it overestimates the number of cyclones originating over the continents, especially along the Beaufort Sea coastline. Second, it underestimates the number of cyclones being generated locally. Combined, these differences provide a disparate picture of Arctic cyclogenesis. Unlike those from the reanalyses, results from CESM-LE suggest that the coastline (and therefore the AFZ) is a preferred area of cyclogenesis.

Some of these discrepancies are also observed in winter, such as the overestimation of cyclogenesis along the Alaskan coast and the underestimation of local genesis throughout parts of the Arctic Ocean. Winter also sees excessive cyclogenesis around northern Greenland and Ellesmere Island, although all three datasets indicate that area as a cyclogenesis maximum.

Despite some clear differences between CESM-LE and the reanalyses in each season, the seasonality of cyclogenesis patterns is consistent in CESM-LE. For instance, cyclogenesis is only common in Siberia in summer. For all CESM-LE members and both reanalyses, the number of CAO+BCEL cyclones coming from Asia averages between
30 to 40 in summer and no more than 11 in winter. By contrast, the number of cyclones entering the Arctic Ocean from the Atlantic and Pacific storm tracks is much greater in winter. The number of locally sourced cyclones is about the same in each season.

Figure 8A.7. Daily geopotential height at 500 hPa for three datasets averaged for (top) all summer months 1990-2005 and (middle) days in which a cyclogenesis event occurs within the Kolyma Lowland (red outline). The bottom row shows the middle row subtracted from the top row. Contour intervals are 25 m in the top and middle rows and 10 m in the bottom row.
8A.2.1. Focus on Local Cyclogenesis

CESM-LE may be broadly similar to the reanalyses, but understanding the reasons for its differences with the reanalyses is important if the ensemble is to be used as a climate projection tool. One of the main differences that needs explaining is the underestimation of local cyclogenesis. This difference may relate to the behavior of the AFZ at mid levels of the atmosphere. As noted in Section 7A.4.2, the Arctic front is neither perfectly circular nor stationary at sub-seasonal time scales. At higher levels of the troposphere, being less strongly constrained by the coastline, it expresses more monthly variability in both mean position and sinuosity. Examining the MAF metric shows that the Arctic front is less likely to meander north of the coastline in CESM-LE than it is in ERA or MERRA (Figures 7.15 and 7.16). Therefore, classic top-down cyclone development downwind of a trough may be less likely to occur over the Arctic Ocean in CESM-LE than in the reanalyses.

8A.2.2. Focus on Eastern Siberia

Another area of difference in summer is eastern Siberia. CESM overestimates the number of cyclones entering the CAO+BCEL from this region each summer (averaging 14-19, depending on the member) compared to MERRA (10) and ERA (11). Recalling again Figure 7A.16 and the MAF, CESM is more likely to see the front shifting south of the coastline. Therefore, this bias may simply represent a southward shift in cyclogenesis, with more cyclone generation in eastern Siberia compensating for less over the Arctic Ocean.

However, it seems that CESM does more than favor cyclogenesis over the continent; it also favors cyclogenesis specifically along the coastline. Consider, for instance, the differences in the upper-level circulation during cyclogenesis events in the Kolyma Lowland, outlined in red in Figure 8A.7. The top row of Figure 8A.7 shows the summer mean GPH at 500 hPa for 1990-2005. The middle row shows a composite of GPH on days in which cyclogenesis occurs within the Kolyma Lowland. Finally, the bottom row shows their difference. In the climatology, all three datasets show a trough extending over the Taymyr Peninsula and central Siberia. However, CESM-LE shows somewhat different behavior during cyclogenesis events. In ERA and MERRA, the trough is deeper, narrower, and shifted eastward during cyclogenesis events so that contours downwind of the trough axis are crossing nearly perpendicular to the Verkhoyansk Range. This suggests that orography plays a notable role, leading
to lee cyclogenesis. For CESM-LE, the trough does not deepen, narrow, or shift nearly as much. The contours downwind of the trough are more zonal and align better with the coastline than across the mountains, suggesting that cyclogenesis often occurs along the coast instead of directly in the lee of the mountains. In other words, Figure 8A.7 provides evidence that the AFZ is more directly responsible for cyclogenesis in CESM than in the reanalyses.

![Figure 8A.8](image)

**Figure 8A.8.** (a) Elevation (m) from the ETOPO1 digital elevation model compared to surface geopotential (m) in (b) CESM, (c) MERRA, and (d) ERA. The Brooks Range (BR) and Verkhoyansk Range (VR) are marked for each.
The greater preference for topography-related cyclogenesis in the reanalyses may in part relate to topographic resolution. CESM has a coarser grid than either ERA or MERRA, so the elevation of narrow mountain ranges like the Verkhoyansk Range is reduced (Figure 8A.8). Diminished barriers have a diminished ability to disrupt large-scale flow, so CESM may assign less importance to the Siberian mountains because in CESM those mountains are smaller. Results may also differ based on the how each model incorporates sub-grid-scale topographic information for calculating wind drag and other variables (Dee et al. 2011; Rienecker et al. 2011; Lauritzen et al. 2015).

The underlying models for MERRA and ERA also have simplified topography, but since observational data are assimilated with output, the final results from these reanalyses are better able to reflect the true complexity of Earth’s surface. In other words, because it blends model output with observations, the reanalysis process has a finer effective topographic resolution than the underlying model. CESM, on the other hand, is limited to the information in the model.

The differences in cyclogenesis in eastern Siberia should not be overstated, though. Figure 8A.9, which shows track density (top row) and spaghetti plots (bottom row) for all tracks generated in summer in this source region, except that the CESM-LE spaghetti plot only shows tracks from one member.33 The yellow arrows in the lower plots are meant to summarize the main routes travelled by cyclones generated in eastern Siberia during summer. In all three datasets, the majority of cyclones migrating out of this region enter the CAO+BCEL (60.0% in MERRA, 62.6% in ERA, 55.7 ± 8.4% in CESM-LE). Their tracks begin with a northeastward propagation but tend to turn to the left over and across the Arctic Ocean. MERRA and ERA exhibit a narrower lane for preferred propagation into the Canadian Arctic Archipelago. This track is still most common in CESM-LE, but the model output also shows cyclones tracking both into the Beaufort Sea and toward the Atlantic side of the Arctic. CESM-LE also shows a few more cyclones taking a more eastwardly track into the Bering Sea. CESM-LE agrees well with the reanalyses in terms of seasonality. Between 4 and 7 CAO+BCEL cyclones are sourced from eastern Siberia each summer. This drops to between 2 and 3 each winter.

33 Only member 30 is shown so that exactly 16 years of data is represented by each spaghetti plot. The choice of CESM-LE member does not make a substantial difference, as can be seen by the close agreement between the spaghetti plot and the track density plot above it.
Figure 8A.9. (top) Contour plots and (bottom) spaghetti plots of cyclone tracks originating in eastern Siberia and Chukotka (dark blue outline) during winter according to (left) CESM-LE member 30, (center) MERRA, and (right) ERA for the period 1990-2005. The CAO+BCEL focus region is outlined in bold gray on the top row and shaded dark blue on the bottom row. Yellow arrows in the spaghetti plots are stylized depictions of preferred cyclone track direction(s).

8A.2.3. Focus on Northwest North America

CESM-LE overestimates the number of cyclones entering the CAO+BCEL region from northwest North America in both seasons, but the greatest differences occur in winter (Figure 8A.6b) around the Alaskan Beaufort Sea coast, where CESM-LE exhibits a particularly strong genesis maximum in winter. ERA also shows a strong maximum here, although it is weaker and smaller. MERRA, meanwhile, shows barely any difference between cyclogenesis along the Beaufort Sea coast and the rest of the Arctic Ocean. Therefore, although CESM-LE may show significant differences from the reanalyses, it is not clear which data set is most accurate.

Past studies that show maps of Arctic cyclogenesis also yield conflicting results. Using an older version of ERA and a detection algorithm based on the Laplacian of SLP, Simmonds et al. (2008) found a relative maximum in cyclogenesis along the North Slope of Alaska, similar to the result shown here for ERA. On the other hand, neither
Serreze (1995) nor Gulev et al. (2001) identified any preference for cyclogenesis in this location based on National Meteorological Center SLP analyses and the NCEP/NCAR Reanalysis, respectively. Since the cyclone detection and tracking algorithm used here is based on that of Serreze (1995), the presence or absence of a relative cyclogenesis maximum on the Alaskan Coast appears to be a matter input data. Reanalyses from ECMWF and CESM identify a maximum, but not MERRA, NCEP, or the National Meteorological Center SLP analyses.

**Figure 8A.10.** As in Figure 8A.9, but for cyclones originating in northwest North America north and east of the Alaska Range, Wrangell Mountains, and Mackenzie Range.

Besides this one small area, the behavior of cyclones generated over northwest North America is consistent amongst these three datasets. **Figure 8A.10** shows track density and spaghetti plots in the same style as Figure 8A.9 only for northwest North America in winter. The vast majority of cyclones generated in this region form on the leeside of the Canadian Rockies and Mackenzie Range. MERRA shows the most winter cyclones with 21.4 per winter, ERA the fewest with 15.5, and CESM-LE in between at 18.6 ± 2.3. Most of these cyclones track east
into Baffin Bay or southeast toward Hudson Bay and the Great Lakes. However, a much higher percentage of cyclones leaving this region migrate into the CAO+BCEL for CESM-LE (38.3 ± 10.6%) than for ERA (19.8%) or MERRA (13.4%). Figure 8A.10 also shows that many of the extra cyclones migrating into the CAO+BCEL region for CESM-LE follow a path that hugs the northern edge of the Canadian Arctic Archipelago. CESM-LE has about the same number of cyclones striking a more direct path into the center of the Arctic Ocean.

Finally, this region is similar for all datasets in that, unlike for eastern Siberia, only the slightest seasonal difference exists between winter and summer. For MERRA, this region contributes an average of 2.3 cyclones per winter to the CAO+BCEL and 1.6 per summer. ERA has slightly more, showing 2.7 per winter and 2.0 per summer. CESM-LE has the most, showing 5.2 ± 1.7 per winter and 4.2 ± 1.1 per summer.

8A.3. The Relationship between the AFZ and Arctic Cyclones

8A.3.1. Spearman Correlations

Recall from Chapter 6 that, contrary to earlier suggestions, the AFZ does not appear to be a region of preferred summer cyclogenesis despite being a zone of baroclinicity conducive to cyclone development. However, the AFZ does impact cyclone development by acting as an intensifier. Looking at MERRA data for all summers 1979-2014, it was found that many CAO+BCEL cyclones originate south of the Arctic coastline. Such cyclones must cross the AFZ before entering the CAO+BCEL domain, and if the AFZ is stronger, those cyclones are likely to experience less filling or more deepening.

As is evident from Figures 8A.9 and 8A.10, CESM-LE shows a similar situation. Overall, only a few more cyclones cross the AFZ each summer on their way to the CAO+BCEL domain in CESM-LE than in either MERRA or ERA. The average track length and lifespan of cyclones crossing the AFZ are also very similar for all three datasets (Table 8A.4).

<table>
<thead>
<tr>
<th>Variable</th>
<th>Units</th>
<th>MERRA</th>
<th>ERA</th>
<th>CESM-LE</th>
</tr>
</thead>
<tbody>
<tr>
<td>Cyclones Crossing the AFZ</td>
<td>#</td>
<td>20.9</td>
<td>19.3</td>
<td>23.5 ± 2.4</td>
</tr>
<tr>
<td>Their Average Lifespan</td>
<td>days</td>
<td>5.3</td>
<td>5.5</td>
<td>5.2 ± 0.3</td>
</tr>
<tr>
<td>Their Average Track Length</td>
<td>km</td>
<td>4342</td>
<td>4462</td>
<td>4178 ± 269</td>
</tr>
</tbody>
</table>
However, two notable differences exist between CESM and the reanalyses that may impact the AFZ-cyclone relationship. First, on monthly time scales the location of the MAF at 700 hPa or 500 hPa in CESM is biased to the south, meaning that the strongest mid-troposphere baroclinicity in any given month is less likely to lie over the Arctic Ocean and more likely to lie over the continents in CESM than in the reanalyses. Second, and perhaps related to this, cyclones that enter the CAO+BCEL domain in CESM are less likely to originate locally and more likely to originate along the coastline. This bias is consistent with a MAF that makes fewer northern excursions since it means troughing is more likely to occur over the continent. In short, CESM appears to assign more emphasis to the AFZ because it shows the coast as a cyclogenesis region. Therefore, it might be expected that the AFZ will have a significant relationship with cyclone frequency in CESM-LE.

In Chapter 6, the hypothesis that variability in AFZ strength modifies cyclone intensification was assessed by calculating the correlation between summer AFZ strength averaged over the entire Arctic Ocean coastline and cyclone intensity in the CAO+BCEL region. When AFZ strength was measured at 700 hPa, the Spearman’s correlation was -0.37 for central pressure and +0.53 for the Laplacian of central pressure, indicating a positive relationship between AFZ strength and Arctic cyclone intensity. Figure 8A.11 shows the results of similar tests for each member of CESM-LE (gray) compared to ERA (blue) and MERRA (red). The correlation estimate for MERRA is not an exact match to that of Chapter 6 because only 16 years (1990-2005) are used in the analysis. Nevertheless, the MERRA estimates in Figure 8A.11 show consistency with their counterparts in Table 6.2. Results from ERA show close agreement with those from MERRA for these correlations.

CESM-LE results, however, show substantial range in correlations. Most members have a positive and significant relationship between summer AFZ strength and the Laplacian of central pressure (Figure 8A.11a), but several members have insignificant results. One member actually has its best estimate as negative. Much of this variability in results can be attributed to the small number of observations. The green line, which shows the correlation coefficient that results if all CESM-LE members are combined in one correlation (n = 16 years/member × 30 members = 480 observations), is nearly identical to the ERA estimate at about +0.40. If using central pressure

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34 MERRA results in Figure 8A.11 are also compiled from 6-hr data instead of 3-hr data, but as noted at the beginning of this appendix, the difference in temporal resolution does not make a notable difference in results.
instead of the Laplacian, results are similarly robust (Figure 8A.11c). Only two CESM-LE members show a positive correlation, and the coefficient using all members together is again very close to the coefficient observed for the reanalyses.

**Figure 8A.11.** Frequency plots of Spearman’s correlation coefficients between AFZ strength at 700 hPa and average (a) Laplacian of central SLP, (b) CAP, and (c) central pressure for cyclones in the CAO+BCEL region, as well as AFZ Strength at 700 hPa and the average number of tracks that (d) cross the AFZ (red outline in inset map), (e) cross the CAO+BCEL domain (light blue in inset map), or (f) form in the AFZ for all summer months (JJA) 1990-2005. The height of each bar indicates how many members for which the correlation coefficient falls within the bar’s width. The green line marks the correlation coefficient that results when combining observations from all members for the correlation, and the blue and red lines mark the correlations for ERA and MERRA, respectively.
CAP, which is influenced in part by cyclone intensity and frequency, also shows consistent correlations with AFZ strength (Figure 8A.11b). Twenty-four of thirty members yield a correlation greater than +0.30, and eight yield a correlation greater than +0.60. In CESM, as with the reanalyses, cyclones that are intensified during strong AFZ years produce more precipitation over the CAO+BCEL region.

Finally, Figure 8A.11d-f show the correlation between AFZ strength and three cyclone frequency measures. In Chapter 6 it was reported that AFZ strength had no significant correlation with track frequency according to MERRA for the period 1979-2014. Looking at the shorter period (1990-2005), the correlations between AFZ strength and the number of tracks crossing (Figure 8A.11d) or forming in (Figure 8A.11f) the AFZ are insignificant for both reanalyses. ERA actually shows a strong positive correlation (+0.49) between AFZ strength and CAO+BCEL cyclone track frequency (Figure 8A.11e), but this correlation disappears (+0.07) if the full record (1979-2014) is considered. Taken as a whole, CESM-LE shows no meaningful relationship between AFZ strength and either track frequency measure (Figure 8A.11d-e). Some individual ensemble members show large correlations, but these may be positive or negative. These findings agree with the reanalyses: the AFZ is a cyclone intensifier, but not a direct cyclone generator.

However, looking at the correlations with AFZ cyclogenesis (Figure 8A.11f), CESM-LE shows a clear preference for positive correlations, with only three members yielding a negative correlation and about half of all members yielding correlations exceeding +0.30. Combing all CESM-LE members results in a correlation of only +0.26 (p < 0.01), though, which is a weaker correlation than observed for any of the intensity measures. As show in Section 8A.2, the coastline is a preferred source region for cyclones entering the CAO+BCEL in CESM-LE, so these results may be reflecting that tendency.

In summary, CESM-LE agrees well with the reanalyses in its depiction of a positive relationship between AFZ strength and summer Arctic cyclone intensity. Unlike MERRA and ERA, however, which showed no evidence for a relationship with cyclone frequency, the CESM-LE data present some evidence that a stronger AFZ also may lead to greater summer cyclone frequency. This is likely because CESM has a bias toward greater coastal cyclogenesis.
Complicating the AFZ-cyclone relationship is the co-variance of both AFZ strength and Arctic cyclone activity with large-scale atmospheric circulation indices like the SVNAM index. In Chapter 6, this potential confounding variable was addressed by removing the linear relationship between the SVNAM index and both AFZ strength and cyclone-related variables prior to calculating the correlation. The index constructed by Ogi et al.
(2004) was used in Chapter 6 for observational data, but this index is not applicable to the model results from CESM-LE. Therefore, in order to perform the same correction on CESM-LE results, a separate SVNAM index must first be constructed for the CESM-LE data.

Creating a time series of an index for an oscillation like the SVNAM has two distinct steps. First, EOF analysis is performed on a set of monthly SLP anomaly fields. The result of this analysis is a solver with multiple EOFs, each of which can be used to define an oscillation index. Figure 8A.12 shows the loading pattern of the leading EOF for MERRA, ERA, and CESM-LE for each summer month. Consistent with the SVNAM identified by Ogi et al. (2004), all three datasets show the main center of action over the CAO. When the SVNAM is positive in summer, SLP is lower than normal over the Arctic Ocean, CAA, Baffin Bay, Greenland, and Greenland Sea and higher than normal at mid latitudes. The loading patterns from MERRA and CESM-LE match particularly well. The main mid-latitude center of action is focused on the North Atlantic, and the Arctic Ocean coastline largely confines the area positively correlated with the CAO. ERA shows a stronger signal in the CAO in July and August, and in July the CAO center of action is shifted toward Siberia. Also, the area positively correlated with the CAO extends farther southward in ERA than in MERRA, CESM-LE and the results shown in Ogi et al. (2004). Lastly, ERA shows a strong center of action in the Bering Sea in August that is absent from the other loading patterns. The main point, however, is that, as with many other aspects of the Arctic climate system, CESM-LE produces a realistic SVNAM comparable to observational datasets.

The second step to defining an index is mapping the appropriate SLP anomaly fields to the desired principal component for each month. Since all atmospheric reanalyses ingest observational data and have similar loading patterns, their resulting index time series are strongly correlated, as shown in Figure 8A.13a. The MERRA and ERA time series are nearly identical. Since Ogi et al. (2004) used the period 1948-2002 to construct the SVNAM loading patterns rather than the 1979-2014 period used for MERRA and ERA, the NCEP results have a somewhat lower (although still high) correlation with the other reanalyses. The significant difference between NCEP and MERRA/ERA indices indicates that NCEP tends toward more positive (or less negative) SVNAM values.

35 The first principal component is associated with the leading EOF.
The indices derived from the CESM-LE EOF solver cannot be directly compared to the reanalyses because unlike the reanalyses, CESM-LE is not influenced by observations. However, so long as they have the same projection, grid cell size, and dimensions, the SLP anomaly fields that are mapped to the EOF solver need not be the same fields that were used to construct the solver. Therefore, CESM-LE SLP anomaly fields can be mapped to the EOF solver based on MERRA, ERA, or CESM-LE data. The mean differences and correlations for these methods are also shown in Figure 8A.13. The mean difference between CESM-LE and the reanalyses is about +1 standard deviation in June and July and +1.5 standard deviations in August. Since the mapped fields are the same for each time series, the discrepancies must come from inputs to the EOF solvers.

![Figure 8A.13](image-url)

**Figure 8A.13.** (a) Pearson’s correlation and (b) mean difference (paired t-test) amongst various SVNAM time series for June (blue), July (red), and August (green). Significance at 95% and 90% confidence intervals is marked by a dot and X, respectively. Reanalyses are compared by using time series generated completely independently for each reanalysis. CESM-LE is compared to reanalyses by mapping CESM-LE fields to the EOF solver derived from CESM-LE, MERRA, and ERA. Time periods are 1979-2011 for test involving NCEP, 1979-2005 for tests involving CESM-LE, and 1979-2014 for the ERA-MERRA comparison.

More specifically, they arise from the strong bias in CESM-LE SLP fields relative to the reanalyses. As discussed in Section 8A.1.1, CESM-LE shows positive bias in SLP over the Arctic Ocean and negative bias in SLP over Eurasia in summer. Therefore, when using an EOF solver based on ERA or MERRA data, the mean CESM-LE state
appears to be the negative phase of the SVNAM. When using an EOF solver based on CESM-LE data, the mean CESM-LE state appears to be neutral.

Despite the presence of bias, the loading patterns for CESM-LE and the two reanalyses are quite similar (Figure 8A.12). For that reason, the correlation between the time series is exceptionally high, exceeding 0.95 for all months (Figure 8A.13a). Therefore, the SLP bias in CESM-LE does not compromise the realism of large-scale circulation variability, such as represented by the SVNAM index.

8A.3.3. Spearman Correlations after SVNAM Adjustment

The results in Chapter 6 showed that, after accounting for the effect of the SVNAM on cyclone characteristics and the AFZ, the relationship between the AFZ and cyclone intensity remained robust, although correlation coefficients were generally lower magnitude. The same pattern is observed for MERRA results limited to the period 1990-2005, as shown in Figure 8A.14a-c.\textsuperscript{36} ERA also shows a decrease in coefficient magnitude, especially for the central pressure of cyclones (Figure 8A.14c), for which the Spearman’s coefficient is essentially zero. Although simple and easily measured, central pressure is not an ideal metric for cyclone intensity because of its dependence on latitude and background SLP (Murray and Simmonds 1991; Serreze et al. 1997). The Laplacian of central pressure, which is independent of latitude and proportional to geostrophic relative vorticity, is a more robust intensity measure. Therefore, more weight should be given to results relating to the Laplacian (Figure 8A.14a).

As with the un-adjusted correlations, CESM-LE members demonstrate a wide range of correlations, but most members are consistent with the results from the reanalyses. Additionally, when combining all members into a single correlation analysis, the final result is very close to the reanalysis results for the Laplacian of central pressure, and CAP.

\textsuperscript{36} However, note the presence of several differences between the correlations for MERRA using the full record of 1979-2014 and the shorter record of 1990-2005. For example, the correlation between AFZ strength and cyclone characteristics in the CAO+BCEL region varies and the Laplacian of central pressure is weaker in MERRA for the period 1990-2005 (+0.21) than for the period 1979-2014 (+0.40). The latter value better matches the value observed in the combined CESM-LE record (+0.36).
Results for AFZ strength and track frequency (Figure 8A.14d-e) and coastal cyclogenesis (Figure 8A.14f) are only slightly affected by the SVNAM adjustment. As with the un-adjusted correlations, little evidence exists for a relationship between AFZ strength and the number of cyclones crossing into the CAO+BCEL, but CESM-LE does indicate a positive relationship between AFZ strength and the number of cyclones generated along the coast.

In summary, as do the reanalyses, CESM-LE shows that the relationships (or lack thereof) between AFZ strength and both cyclone intensity and frequency are independent of large-scale circulation as measured by the SVNAM. Since the AFZ appears to influence both cyclone intensity and frequency in CESM-LE (as opposed to just
intensity), CESM may over-emphasize the AFZ’s role in the Arctic climate system. However, any cyclogenesis related to troughs in the Arctic’s jet-like feature and its interactions with topography can still be indirectly linked to the ocean-continent heating contrasts that initially inspire the seasonal and regional formation of an Arctic front and Arctic jet-like feature. Therefore, rather than over-emphasizing the AFZ’s importance, it may be better to characterize CESM as emphasizing more direct effects of coastal heating contrasts, whereas MERRA and ERA emphasize more indirect effects.