An analysis of Arctic climate, the intense Arctic cyclone of early August 2012, and middle to high latitude snowcover

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An Analysis of Arctic Climate, the Intense Arctic Cyclone of Early August 2012, and Middle to High Latitude Snowcover

by

Adam H. Turchioe

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ABSTRACT

On 03 August 2012, a cyclone formed over Central Siberia and progressed northeastwards. By 0000 UTC 05 August, the cyclone reached the Arctic Ocean with a mean sea-level pressure (MSLP) of 984 hPa. Once over the Arctic Ocean, the cyclone rapidly intensified and reached a minimum pressure of 966 hPa on 06 August near 83°N and 170°W. The cyclone slowly weakened, and on 0000 UTC 10 August once again had a minimum MSLP of 984 hPa. The motivation for this presentation is driven by the likelihood that this cyclone is one of the most intense storm systems to ever impact the Arctic Ocean in the modern data era. The rarity of this storm is further supported by the fact that it occurred during the summer, prior to the climatologically favored more intense cyclone-season beginning in the fall. The purpose of this thesis will be to present the results of a climatological analysis of Arctic Ocean conditions between for 1979 to 2012. Intense cyclones to occur during this period are analyzed by frequency, trajectory, and intensity. They are categorized based on track, intensity, and seasonality. 0.5° resolution NCEP Climate Forecast Reanalysis data was the primary data source for this study. Global Forecast System FNL data was used for the cast study of the cyclone of August 2012.

Changes within the climatology of snowcover extent are analyzed within this presentation, as well as a relationship between October snowcover extent and the Arctic Oscillation (AO). Suggestions will be made relating aforementioned changes in Arctic atmospheric properties and associated changes in snowcover extent. Weekly snowcover extent departures from the 25-year (1989-2013) climatology were catalogued from 1970-2014 and will be discussed.
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1. Introduction

1.1 Motivation and Objectives

For millennia, little was known about the Arctic and its associated climate. The Arctic region of the globe is one of the least densely populated regions on earth, and inherently has very sparse weather observations. Only technological advancements through the centuries have allowed us to get a peek into the mysterious “land of the midnight sun”. Prior to the installation of an Arctic weather network in the early 1950’s, high-latitude climate could only be theorized based on these few observations. With the advent of the modern meteorological satellite era in 1979, much more data has become available and thus, much more work has been done of the climatology of the Arctic. This influx of new meteorological data is very important, as the Arctic is undergoing rapid warming, especially when compared to the rest of the globe (Stroeve 2014). Several attention-grabbing news headlines have occurred over the past several years that have peaked interest in Arctic weather and climate. First, according to the National Snow and Ice Data Center (NSIDC), the areal sea-ice extent reached an all-time low in recorded history of 4.06 million km$^2$ in September of 2007. Then, on 1800 UTC 06 August 2012 a severe cyclone over the Arctic reached a minimum mean sea-level pressure (MSLP) of 962 hPa (Simmonds and Rudeva 2012). At the time of the event, it was uncertain how this extreme event fit into the climatology of intense cyclones. In order to quantify whether or not this cyclone was the most intense to occur in the month of August and the summer, several questions would need to be answered.

So, began one of the main motivations of this paper: (1) to develop a distinct Arctic cyclone climatology based on analyzing cyclogenesis, track density, and antecedent conditions
of the deepest cyclones (using MSLP) to occur in the Arctic by month and season. This climatology will allow for a method of quantifying precisely how rare the August 2012 cyclone was, and (2) a case study will be performed on the August 2012 cyclone including planetary, synoptic, and mesoscale factors that lead to its rapid intensification. Five weeks after the conclusion of the August 2012 cyclone, the sea-ice extent again reached a new all-time minimum extent at 3.17 million km$^2$ (NSIDC). This noteworthy headline occurred after, according to the National Climatic Data Center (NCDC) one of the most anomalous warm spell in March of 2012 and extreme drought/heat throughout the summer of 2012 across the United States. Media concern over these being signs of global warming went viral. This paper will also look at long-term trends within Arctic climate to judge whether significant changes are occurring within this rapidly changing part of the world. Another goal will be to assess whether or not the loss of sea-ice extent is affecting other climate and weather variables, such as October snowfall extent and upper-air flow patterns/strength. There exists a correlation between October snowfall and the Arctic Oscillation (AO), which will also be examined (Cohen 2001,2007,2012). Firstly, a literary review of scientific findings relevant to the understanding or Arctic climate will be discussed in chapter 1. Data and methodology for the studies conducted within this paper will be compose chapter 2, while chapter 3 will present the findings of the cyclone climatology, August 2012 cyclone case study, Arctic climatic changes/trends, and linkage between October Eurasian snowcover and the AO. This paper will conclude with chapters 4 and 5, which will summarize the results of this research along with key conclusions.
1.2 Literature Review

1.2.1 Early Understanding of Arctic Climate and Cyclones

Western interest in Arctic exploration first peaked in the sixteenth century as a potential trade route to China (Armstrong 1984). This so-called “Northeast Passage” lured many explorers (Brunei, Hudson, Baffin, etc.) to the region, but the lack of knowledge of Arctic climatology forced all of these missions to be abandoned when snow, cold, or ice halted further advancement into the Arctic. For the next several centuries, meteorological data was limited to basic observations from those living in the Arctic. The first attempt to attempt to understand the climate of the Arctic occurred for three years between 1881-1884 during a project dubbed the “First International Polar Year” (Barr 1985). This expedition was a joint effort between 11 nations in which over 700 men endured brutal conditions to record meteorological data. Unfortunately, the sheer distance between the 12 observation stations (see Fig. 1.1) meant that meaningful meteorological conclusions were difficult to ascertain from the study. Also, a tragic loss of life occurred to Adolphus Green’s party during the mission that, understandably, lessened the enthusiasm from the scientific community for further research to take place.

The difficulty in developing a thorough climatology of atmospheric conditions in the Arctic continued throughout the early part of the 20th century, as the only work done was the development of the now-defunct Glacial Anticyclone Theory (Hobbs 1926). This theory stated that a permanent high-pressure exists over Greenland and Antarctica, and these are the main driving forces on circulation of the Arctic, Antarctic, and mid-latitudes. As knowledge of global geography and meteorology improved, another push was made to increase understanding of the
Arctic. The Second International Polar Year began in 1932 with 94 meteorological stations in the Arctic (Brooks 1959; Laursen 1982). The aide of radio communications greatly helped observational data collection, and a vast amount of data was accumulated. The success of the mutual cooperation of several nations led to a world data center that would eventually become the International Meteorological Organization. The promise of the subsequent studies/publication releases would be hampered by the breakout of World War Two. However, all was not lost; as a side effect of the technology advancement during World War II, a major breakthrough came in the early 1950’s when the Arctic Weather Network was developed by the United States and Canada. Using naval/air weather reports and drifting buoys, it was now possible to analyze detailed weather reports across the Arctic. This aided in the development of an understanding of climate in the Arctic (Hare and Orvig 1958; Namias 1958; Reed and Kunkel 1960). The most thorough report on high-latitude cyclones yet, “Arctic Synoptic Activity in Winter”, was also published in 1958 (Keegan 1958). Keegan used weather reports from U.S., Canadian, and Soviet ice stations along with U.S. Air Force reconnaissance flights to count and track cyclones and anticyclones over the course of 15 winter months. Cyclones were identified by a hand analysis of MSLP and wind direction. Fig. 1.2 shows the results of the analysis of cyclones. It shows a climatological maximum of cyclones emerging into the Arctic from the North Atlantic between Greenland and Scandinavia. Secondary maxima appear poleward of Scandinavia and just west of Greenland. Fig. 1.3 shows the results of the analysis of anticyclones, with three distinct maxima located over northern Greenland, northwestern North America, and eastern Siberia. While this study did not specifically examine anticyclones, the Greenland high and the Siberian high-pressure systems have been identified and are semi-permanent, important features of their respective regions (Gong and Ho 2001). Results of the Keegan study were confirmed by data
gathered by the International Geophysical Year (IGY) which was a collaboration between 67 nations from 1 July 1957-31 December 1958. It studied eleven earth sciences, including meteorology, and was deemed a great success in terms of data collection. Perhaps the biggest success of the IGY was the launch of Sputnik I by the USSR. While Sputnik I did not directly study meteorology, subsequent satellites, beginning with TIROS I in 1960, would (National Aeronautics and Space Administration (NASA)). The results of early papers and data collection led the way for further advancement and refinement of knowledge of climatology in the Arctic, especially in the modern satellite era since 1979.

1.2.2 Middle to High-Latitude Cyclones

Beginning in the 1980’s renewed interest in the Arctic region, coupled with an influx of new meteorological data, resulted in a slew of papers investigating climate in the high-latitudes (Overland and Pease 1982; LeDrew 1982; Serreze and Barry 1988; LeDrew 1984; Serreze et al. 1992). Fundamentally, Arctic climate is characterized by its very cold winters, cool summers, and drastic changes in solar insolation between summer and winter. Parts of the Arctic are in polar night or day for 6 months at a time, and any latitude poleward of 67°N is witness to at least one day of total sunlight or darkness. Even when there is sunlight hitting the surface, very little heat energy is realized due to the low sun angle (only gets as high as 23°) and high reflectivity of any ice or snow located on the Pole. This results in a very low-thermal base state existing across the high latitudes. The stark contrast between the cold, polar air mass and the warm, tropical air mass results in a strong baroclinic zone in the mid-latitudes. Using the thermal wind equation:

\[
\frac{\delta u_x}{\delta v} = \frac{\delta}{\delta v} \frac{\delta y}{f \rho \delta y} \quad \text{where} \quad \frac{\delta u_x}{\delta v} \text{ is the change of zonal geostrophic winds with height (pressure)}, \quad R_d \text{ is the} \]


gas constant of dry air, $f$ is the coriolis parameter, $p$ is pressure, and $\frac{\partial T}{\partial y}$ is the meridional temperature gradient, it is determined that the strongest jet stream will coincide with the strongest meridional thermal gradient. The polar jet stream is typically between the latitudes of 50°-60°N (National Weather Service (NWS), 2011), closer to 50°N in Boreal winter and closer to 60°N in Boreal summer. Since the meridional temperature gradient is typically much stronger in the winter than in the summer, strength of the jet stream is typically much stronger in the winter. The Arctic, hereafter defined as regions of the Northern Hemisphere poleward of and including 70°N, is typically well north of the jet stream. It has long been known that extratropical cyclones are primarily associated with strong jet stream dynamics (Sutcliffe 1947; Petterssen 1956; Palmen and Newton 1969). This yields the result that most cyclones, especially in the winter, propagate to the Arctic from lower latitudes. Due to the natural baroclinic zone that exists on the east coast of North America between the cold Canadian air masses and warm tropical oceanic airmasses, a favored region for cyclogenesis exists in the western North Atlantic (Colucci, 1976). Sanders and Gyakum (Sanders and Gyakum 1980) classified the “bomb” as intense cyclones that rapidly deepen at or above the pace of 24 hPa/24 hours normalized to 60° N (latitude of Bergen, Norway). Fig. 1.4 shows the favored regions for development of these intense cyclones. Explosively deepening “bomb” cyclones maximize over the western Pacific Ocean and western Atlantic where the SST gradient is largest and the atmospheric 1000-500-hPa gradient is largest. These powerful cyclones often continue to the northeast and trek into the Arctic to the east of Greenland. Many of these cyclones continue their march to the high-latitudes and re-intensify between Greenland and Scandinavia, possibly reinforced by the natural baroclinic zone between the very cold Greenland air mass and the northern terminus of the warm Gulf Stream (Tsukernik et al. 2007). Fig. 1.5 is one of the best representations of this pattern, as
it shows emergence of these North Atlantic cyclones into the Arctic east of Greenland (Serreze and Barrett 2008) in the winter (DJF). As the jet stream migrates northward with the poleward journey of the sun in the spring, the strength of the jet itself begins to weaken in conjunction with the diminished baroclinic zone that exists during boreal summer. Because of this, during late spring to early fall storms can develop in-situ within the Arctic itself. Fig. 1.6 is from the same paper from Serreze and Barrett, but for the summer months of their study. As can be seen, a favored region for development occurs over the central Arctic Ocean between the Canadian Archipelago and eastern Siberia.

When looking at cyclones overall, most recent papers have used tracking algorithms that analyze MSLP or vorticity at 850 hPa to identify and count cyclones in the Arctic. Zang et al. (2004) created a comprehensive climatology of Arctic cyclones using one of these algorithms. In general, this study agrees with the results of their paper. Especially in agreement is the result of cyclone intensity seasonality, as is shown in Figure 1.7. Intensity was defined as the mean absolute values of the difference between the central SLP of cyclones and the climatological monthly mean MSLP at corresponding grid points. Mean absolute values are averaged over all cyclones, by region, for each month. Cyclone intensity was found to be most intense during the winter and least intense during the summer. The number of cyclones entering the Arctic is highest in the summer and lowest in the winter. Cyclone duration also peaked in the summer with a minimum in the winter. The study also found that as a general rule of thumb, Arctic cyclones that originate in the lower latitudes tend to be stronger than those that originate within the Arctic. This study will look to examine further cyclones that develop both in-situ as well as those that propagate into the Arctic from mid-latitude regions.
1.2.3 Climate Change in Arctic

The Arctic has warmed over twice as fast as the rest of the globe in the 20th century (Serreze and Barry 2011; Screen et al. 2013). The main reason for this is because of albedo. Albedo is a unitless measure of the amount of solar radiation that gets absorbed into a surface (Uttal et al. 2002). As the earth warms, a small amount of ice and snow extent is lost in the high-latitudes. As a result, instead of highly reflective snowcover ($\alpha = 0.9$) and less-so reflective sea-ice ($\alpha = 0.5$) reflecting solar insolation back to space, solar energy is absorbed into the surface of the ocean ($\alpha = 0.06$). A positive feedback then occurs, as the newly created open-water will warm faster, further melting ice and snow more quickly. It appears as if this positive feedback may have begun to run rampant, as the pace of sea-ice loss is at an all-time high since observations began in 1979. According to a recent paper by Stroeve et al. (Stroeve et al. 2014), the Arctic melt season has lengthened more than five days decade$^{-1}$. This is due to the fact the upper-ocean is storing much more heat. Over the past decade, sea-surface temperatures have increased 0.5°C-1.5°C and approximately 752 MJm$^{-2}$ of increased heat energy is contained within the top layer of the Arctic Ocean. From 1979-2001, sea-ice loss was recorded at -7.0% decade$^{-1}$. Now, the pace of sea-ice loss is an astounding -14% decade$^{-1}$. Figure 1.8 clearly shows this trend of accelerated sea-ice loss through the observable period. The rapid pace of ice loss has been accelerating since the early part of the 21st century. Astoundingly, some peer reviewed publications (Stroeve et al. 2007; Boe et al. 2009; Wang and Overland 2009) suggest that the entirety of the Arctic basin could be ice-free by the middle to end of this century. This albedo feedback is not the only factor in the impressive sea-ice loss, as will be discussed in later chapters.
A lot of attention has been paid to the idea of Arctic amplification over the past several years. Arctic amplification is a term regarding changes in temperatures in the Arctic relative to the rest of the globe. A positive polar amplification effect occurs when the Arctic is warming faster than the rest of the globe. Figure 1.9 clearly shows that the arctic has been warming at a much faster pace than the rest of the world in the period from 1960-2009. Going back to the equation for thermal wind as shown on pg. 6, as the temperature contrast weakens, as does the jet stream above the baroclinic zone. It has not been explicitly proven, but there is some evidence that the Northern Hemisphere Polar jet stream may be weakening in response to Arctic Amplification. Francis and Vavrus (2012) showed evidence that not only is the jet stream observably weakening in strength, but the pace at which it propagates eastwards due to amplification of the Rossby wave flow has lessened. A schematic of what their results have shown are displayed in Figure 1.10. If correct, this observation would affect the Arctic regions in several ways. The amplified flow would allow for more heat and moisture to advect into the Arctic, possibly furthering the positive feedback of sea-ice loss. The amplification would also allow for the high-latitude landmasses, especially Eurasia, to heat up more distinctly. This could begin increasing the baroclinic zone that naturally exists within this region. As will be discussed later, this baroclinic zone is one of the reasons for cyclone development over the Arctic, especially in the summer months (Jun-Aug). A less progressive Rossby wave pattern would also in theory reduce the number of cyclones that occur within the Arctic for all months of the year. Keeping all of this in mind will help to answer questions about relating results of this survey to possible climate change.
1.2.4 Characteristics of the Arctic Oscillation (AO)

A “see-saw” of atmospheric mass between the Arctic and mid-latitudes was first discovered by Edward Lorenz (Lorenz 1951), and has become to be known as the Arctic Oscillation (AO) (Thompson and Wallace 1998). The AO is a measure of pressure over the polar regions vs. pressure in the mid-latitude regions. When the AO is positive, low pressure is the dominant feature of the Arctic while high pressure is dominant in the mid-latitudes. This strengthens the jet stream as it becomes more progressive, restricting the flow of cold air from the high-latitudes into the mid-latitudes. The opposite is true when the AO is negative, as high pressure is dominant over the high-latitudes with low-pressure over the mid-latitudes. The jet stream weakens, amplifies, becomes less progressive, and expels cold air from the Arctic into the mid-latitudes. The AO has not only huge implications for the human populations across the middle latitudes, but also has big implications for cyclone activity across the Arctic (Rigor et al. 2002; Comiso 2003; Rigor and Wallace 2004). When intense cyclones occur in Arctic regions, the precipitation and heavy wave action wreaks havoc on coastal communities on the Arctic. Coupled with longer and longer periods of ice-free coastal waters (Stroeve et al. 2014), coastal erosion has become a big problem in these areas. When the AO is positive, the number of cyclones in the Arctic dramatically increases (Zhang et al. 2004) having major implications for the quality of life for members of Arctic coastal communities. A negative AO decreases the frequency of cyclones through the Arctic, but can cause severe cold and intense cyclones to impact populations in the mid-latitudes (Thompson and Wallace 1998, 2000).

Due to the huge implications by the AO on weather patterns, the ability to predict the AO in long-range forecasts would be very beneficial to anyone involved in the energy, insurance,
governmental, research, etc. communities. Long thought of as stochastic and random, prediction of the AO long-term was thought to be impossible. Cohen and Entekhabi (1999) published the first paper linking the relationship between snow cover and variability of the AO. In theory, the main diagnostic used is the wave activity flux (WAF) that is given as

\[ F = (\Omega \sin(2\phi)/S)\left(\nu'' T'' - [2\Omega \alpha \sin(2\phi)] - \frac{15\Gamma' \Phi' \lambda'}{\xi \lambda} \right) \]

where \( \Omega \) is the angular velocity of the earth, \( \rho \) is pressure, \( \phi \) is latitude, \( \nu \) is meridional velocity, \( T \) is temperature, \( \alpha \) is the mean radius of the earth, \( \Phi \) is geopotential height, \( \lambda \) is longitude, and \( S \) is the static stability of the atmosphere. When upward wave activity flux occurs, eastward momentum is deposited into the stratosphere. This acts to weaken the polar vortex and in some cases can even result in a sudden stratospheric warming event (SSW). Regardless of if a SSW event occurs, the continued upward WAF can weaken the polar vortex resulting in a prolonged state of negative Arctic oscillation. Using this theory, the more(less) upward WAF, the weaker(stronger) the polar vortex will become and the AO will respond in more of a negative(positive) state (Cohen et al. 2007). Cohen and other authors have found that a great way to increase upward WAF in the fall and winter, during the development of the polar vortex, is to increase snowcover extent in the month of October. A vast and rapidly expanding snowcover will strengthen the Siberian high and its associated cold-pool. In stark contrast to the very warm tropical air mass of the western Pacific, a vigorous baroclinic zone begins to set up. A very intense baroclinic zone will lead to a pronounced jet stream across eastern Eurasia, and more upward WAF will occur. A schematic of the process is displayed in Fig. 1.11. Looking towards the future, almost every climate model shows a decrease in areal Arctic ice extent (Stroeve et al. 2007; Boe et al. 2009; Wang and Overland 2009). It has been found that there exists a direct correlation between ice loss over the Arctic Ocean and snowcover in Eurasia (Liu et al. 2012). Fig. 1.12 shows the correlation
between actual and detrended autumn sea ice area anomaly and the AO (a), and the linear regression of winter snow cover anomalies on the detrended Arctic sea ice area anomaly (b). This could have large implications regarding the AO, as decreasing Arctic ice-extent could lead to more snowcover across Eurasia in the fall. In concert with previous studies, this paper will examine the relationship between October Eurasian snowfall and the following winter’s AO.
1.3 Research Goals

This paper will aim to: (1) Create a comprehensive list of the most intense Arctic cyclones using MSLP data; (2) Classify these cyclones in the manner of which they formed or propagated into the Arctic; (3) Determine the frequency of intense cyclones in the Arctic; (4) Attempt to create composites of cyclones of a similar genesis; (4) Create a case study of the early August 2012 intense Arctic cyclone. Planetary, synoptic, and mesoscale processes that led to its intense development will be discussed. The cyclone’s context within the newly created climatology will help to distinguish this cyclone as either a rare event or common occurrence. Its impact on the record sea-ice loss of 2012 and associated downstream impacts are included; (5) Examine long-term climatic trends within the Arctic that could be associated with future cyclogenesis, as well as identify trends of the Arctic within our warming world; (6) Examine the relationship between October Eurasian snowcover and the Arctic Oscillation. We will also assess whether there is a correlation between snow cover trends over the past 30 years with loss of sea-ice and increasing atmospheric thermal properties of the high-latitudes. The main goal of this paper is to examine a region of the earth that, compared to other parts of the globe, have not been examined quite as often. Changes in the Arctic have significant ramifications involving mankind and the environment. As sea-ice continues to recede, the “Northeast Passage” is beginning to open up. This will lead to crude oil/natural gas drilling sites forming and shipping lanes opening up across the Arctic. Due to the severity of these cyclones, knowledge of Arctic weather and climate will become very important to individuals associated with these endeavors. Meteorologists will soon be explicitly instructed to produce forecasts for the very high-latitude regions; in fact, some companies have already begun to make the Arctic a part of their forecast
territories. Climate trends will become ever so important as new sources of CO₂ emissions are introduced into the Arctic. Climate change will yield impacts on the local populations, as well as flora and fauna of these regions and it is best we understand what has already been occurring in the high-latitudes. This paper will strive to tackle all of these ultimate goals in the hope of furthering academic discovery in the land of the midnight sun.
Figure 1.1. Eleven nations established 14 principal research stations across the Polar Regions. 12 were in the Arctic, along with at least 13 auxiliary stations. Over 700 men incurred the dangers of Arctic service to establish and relieve these stations between 1881 and 1884 [adapted in part from NOAA Arctic Research Office].
Figure 1.2. Per cent frequency of cyclones in winter north of 60°N per 100,000 sq. mi. One of the first publications developing a cyclone climatology in the winter [Fig. 1 and adapted caption from Keegan (1958)].
Figure 1.3. Per cent frequency of anticyclones in winter north of 60°N per 100,000 sq. mi [Fig. 2 Adapted in part from Keegan (1958).]
Figure 1.4. Distribution of bomb events during three cold seasons. Raw non-zero frequencies appear in each $5^\circ \times 5^\circ$ quadrilateral of latitude and longitude. Isopleths represent smoothed frequencies, obtained as one-eighth of the sum of four times the raw central frequency plus the sum of the surrounding raw frequencies. The column of numbers to the left and right of the heavy line along longitude $90^\circ W$ represent, respectively, the normalized frequencies for each $5^\circ$ latitude belt in the Pacific and Atlantic regions, using a normalization factor of $(\cos 42.5/\cos \phi)$. Heavy dashed lines represent the mean winter position of the Kuroshio and the Gulf Stream [adapted in part from Sanders and Gyakum, 1980].
Figures 1.5 & 1.6. Counts of surface cyclone centers in equal area boxes of 250 km x 250 km for (a) winter [December-February (DJF)] and (b) summer [June-August(JJA)] totaled over the period of 1958-2005 for the region north of 60°N. Results are based on NCEP reanalysis data. Fields have been smoothed by first averaging counts for each grid cell, along with counts at the four adjacent cells, and then applying a nine-cell center weighted average. Bold contours indicate regions with at least 135 cyclone centers. Due to spurious cyclones associated with the reduction of surface pressure to sea-level, Greenland is masked [adapted in part from Serreze and Barrett, 2005].
Figure 1.7: The long-term mean seasonal cycle of cyclone (a) intensity, (b) trajectory count, and (c) duration in the Arctic region (60°-90°N) and its two subregions, the Arctic Ocean (70°-90°N) and the Arctic marginal zone (60°-70°N). The units of intensity and duration are in hPa and h, respectively [adapted from Zhang et al., 2004].
Figure 1.8. Graph of average monthly sea-ice extent for September 1979-2013. A steady decrease in average sea-ice extent can be seen, with a more dramatic pace of ice loss since the turn of the century [Image from NSICD].
Figure 1.9. Linear trends in annual mean surface air temperature for the period 1960-2009, based on the National Aeronautics and Space Administration Goddard Institute for Space Sciences (NASA GISS) temperature analysis (http://data.giss.nasa.gov/gistemp). The inset shows linear trends over the 50-year period averaged by latitude. [Adapted in part from Serreze, 2009].

Figure 1.10. Region of study: 140°W to 0°W. (a) Asterisks illustrate an example of a selected range of 500 hPa heights used in the analysis. (b) Schematic of ridge elongation (dashed vs. solid) in upper-level heights caused by enhanced warming in Arctic relative to mid-latitudes. Higher amplitude waves progress eastward more slowly, as indicated by arrows [adapted in part from Francis and Vavrus, 2012].
Figure 1.11. Conceptual model for how fall snow cover modifies winter circulation in both the stratosphere and the troposphere; case for extensive snow cover illustrated: 1) Snow cover increases rapidly in the fall across Siberia, when snow cover is above normal. 2) Diabatic cooling helps strengthen the Siberian high and leads to below normal temperatures. 3) Snow-forced diabatic cooling in proximity to the high topography of Asia increases upward flux of wave activity from the troposphere, which is absorbed into the atmosphere. 4) Strong convergence of WAF leads to higher geopotential heights, a weakened polar vortex, and warmer temperatures in the stratosphere. 5) Zonal mean geopotential height and wind anomalies propagate down from the stratosphere into the troposphere all the way to the surface. 6) Dynamic pathway culminates with strong negative phase of the Arctic Oscillation at the surface [adapted from Cohen et. al, 2007].
Figure 1.12. (A) Time series of actual and detrended autumn Arctic sea ice area anomaly ($\times 10^6 \text{ km}^2$) and winter AO index and (B) linear regression of winter snow cover anomalies (%) on the detrended autumn Arctic sea ice area anomaly (regions within contours denote the regression above 95% confidence level).
2 Data and Methodology

2.1 Data Sources

A climatology of the number/frequency of Arctic cyclones was developed using gridded data from the four-times daily (0000, 0600, 1200, and 1800 UTC) 0.5° x 0.5° National Centers for Environmental Prediction (NCEP) Climate Forecast System Reanalysis (CFSR; Saha et al. 2010). Mean sea level pressure (MSLP) data was the only variable used for the cyclone climatology study. If uncertainty existed regarding whether or not a cyclone counted for the climatology, manual inspection of daily averaged data from the National Centers for Environmental Protection (NCEP)/ National Center for Atmospheric Research (NCAR) Reanalysis 1 2.5° x 2.5° resolution (http://www.esrl.noaa.gov/psd/data/gridded/data.ncep.reanalysis.html) MSLP aided in producing this full climatology.

Cyclones were tracked and monitored for the period 1979-2013, again using MSLP data from the CFSR dataset. A tracking algorithm scripted within NCAR Command Language (NCL) displays the latitude and longitude of each cyclone for every six-hour period the storm existed within the Arctic. The algorithm examined every single grid point within the 0.5° gridded dataset domain poleward of 70°N. If the MSLP at any single point in the Arctic was lower than every single immediate neighbor to that point, then that low-pressure minimum would count as a possible cyclone candidate (http://www.ncl.ucar.edu/Document/Functions/Built-in/local_min.shtml).

CFSR Data was also used to examine long-term changes within the Arctic. For this effort, the 1000, 925, 850, 700, 500, and 300 hPa levels were used. Variables used for some of these
levels included: geopotential height, temperature, total columnar precipitable water, u-wind, v-wind, and omega.

A case study of the August 2012 Arctic cyclone used the NCEP FNL (Final) operational global analysis 1° x 1° grids data. It is produced every six hours (0000, 0600, 1200, 1800 UTC) from the Global Data Assimilation System (GDAS) which continually collects observational data from the Global Telecommunications System (GTS). FNL analyses are made with the same model that NCEP’s Global Forecast System (GFS) but are released an hour later than the GFS data is released so that more observational data can be used. FNL data would have been used to complete the entire climatology since it has more observational data, but its temporal range only goes back to 1997. The case study used the 1000, 925, 850, 700, 500, 300, and 10 hPa isobaric levels, and surface, 10m levels within the dataset. Variables used included: MSLP, U wind, V wind, omega, geopotential heights/anomalies, Convective Available Potential Energy (CAPE), Precipitable Water (PW’s), sea-ice concentration, temperature, and thickness calculated between the 1000-hPa and 500-hPa isobaric levels. Atmospheric upper-air raidiosonde data was used for detailed observation of conditions leading up to cyclogenesisis and intensification. Data was accessed from the University of Wyoming College of Engineering (http://weather.uwyo.edu/upperair/sounding.html). The location chosen for radiosonde analysis in the August 2012 cyclone case is displayed in Figure 2.1. Backward trajectories were produced using the National Atmospheric and Oceanic Administration (NOAA) Air Resources Laboratory Hybrid Single-Particle Lagrangian Integrated Trajectory (HYSPLIT) model (Draxler and Hess 1997; Draxler 1999; http://ready.arl.noaa.gov/HYSPLIT.php).

Storm-relative composites were computed using the NCEP/NCAR 2.5°x 2.5° 4x daily Reanalysis I dataset as described above. 500 hPa standardized heights/anomalies, 300 hPa wind
speed, MSLP and anomalies, and 850 hPa temperatures/anomalies, and Q vectors were used to
determine differences between the two cyclone groups. Q-vectors were calculated using the Q-
vector parameter in General Meteorology Package (GEMPAK; DesJardins et al. 1991). The
script created the Q vectors using:

$$
\vec{Q} = \begin{pmatrix}
\frac{\partial \vec{v}_g}{\partial x} - \vec{v}_p \theta \\
\frac{\partial \vec{v}_g}{\partial y} - \vec{v}_p \theta \\
\end{pmatrix} = \begin{pmatrix}
Q_1 \\
Q_2 \\
\end{pmatrix}
$$

Determining forcing for vertical motion was calculated using the right-hand side of the Q-vector
form of the QG omega equation:

$$
\left( \sigma v_p^2 + f_0^2 \frac{\partial^2}{\partial p^2} \right) \omega = -2h \vec{v}_p \cdot \vec{Q}
$$

as defined in Keyser et al. (1992) where \( h = \left( \frac{\kappa}{\mu_b} \right)(\rho_p \xi_p) \) and \( \rho_p \) is a reference pressure (1000
hPa). \( \sigma \) is the static stability coefficient, as defined by \(-h d\Theta/dp\) where \( \Theta \) is a pressure-
dependent reference profile of potential temperature. In conjunction with common practice,
convergence (divergence) indicates regions of forcing for ascent (descent). The storm-relative
composites shifted the latitude/longitude grids to a common center based on where the minimum
MSLP of each cyclone occurred, whereas geographic-relative composites did not shift the
latitude/longitude grids. These also used the NCEP/NCAR Reanalysis I 2.5°x2.5° dataset.

Arctic Oscillation data was obtained from the Climate Prediction Center (CPC;
http://www.cpc.ncep.noaa.gov/products/precip/CWlink/daily_aot_index/ao_index.html). The
daily AO index projects the 1000 hPa 0000UTC geopotential height anomalies onto the loading
pattern (first EOF of monthly mean 1000 hPa geopotential height anomalies) to gauge the polarity and strength of the Arctic Oscillation. Snowfall data was derived from the Rutgers University Northern Hemisphere (NH) Snow Cover Extent (SCE) Climate Data Record (CDR). Early snowfall data was digitized (Robinson 2000) from hand drawn maps onto an 89x89 NH grid that is still used today. Data is collected using shortwave satellite sensors, and a grid cell qualifies as snow covered if more than 50% of the cell is covered in snow. The data is relatively course, but is suitable for continental-scale climate studies such as this. NCEP/NCAR Reanalysis data was the source for identifying decadal trends and differences between decades in an attempt to identify.

2.2 Methodology

2.2.1 Extreme Cyclone Identification and Frequency

As per one of the main inspirations for this paper, a comprehensive climatology of cyclones, based on MSLP, needed to be constructed. Since the CFSR data used for this analysis ran at 4x daily intervals since 1979, the climatology produced is sub-daily; every six-hour period from 1979-2013 is included. This leads to a total of 49,640 data points that lead to the absolute lowest MSLP for each 6-hour period over the 34 years. This method is the primary way to identify the most intense cyclones based on MSLP. Frequency of these cyclones was calculated by looking at the 34-year climatology of minimum pressure within the Arctic and using the 2, 1.5, and 1-sigma thresholds for cyclone identification. The average absolute minimum MSLP in the Arctic was used for every period within each month since 1979 to calculate the mean value. Standard deviation (sigma) was calculated with these same data points to calculate sigma for
each month. Table 1 shows these results. 6-hour periods in which the Arctic had a lower MSLP than 2, 1.5, and 1 sigma values lower than that month’s mean value were counted as a “cyclone period”. The data was tested to diagnose if it followed a normal bell-curve; 6-hourly data for each month using 2, 1.5, and 1-sigma thresholds are within 0.5% and contain 16%, 12%, and 3%, respectively, of all data points for that month. The results section will describe how the 6-hourly cyclone period counts have changed over the 34-year period.

The ten most intense cyclones for each month were studied in finer detail. In order to track location of maximum intensity for each cyclone, as well as its genesis and movement, the location of each relative minimum MSLP was given in terms of latitude and longitude for every six hours. Moving backwards from the point of maximum intensity of each cyclone yielded track data for all intense cyclones. Utilizing this knowledge allowed for two questions to be answered: Where in the Arctic did the cyclones reach maximum intensity and what part of the globe did the cyclones enter into the Arctic from? A few problems arose from this methodology, however. Often times, especially in the summer, the intense cyclone formed in response to an emergent low pressure joining with a parent low pressure antecedently located over the Arctic. Figure 2.2 is an example of this; shown is MSLP, 1000-500 hPa thickness, and total column precipitable water. For these cases, the emergent low-pressure track was counted as the pathway for which that particular intense cyclone entered the Arctic. Another issue with this technique deals with tracking of these cyclones. Many cyclones, especially in winter, reached maximum intensity prior to entering the Arctic, but still might have been a top-10 most intense cyclone for that month when they entered the Arctic. These cyclones were included in the climatology since technically speaking they were still in the upper-echelon of Arctic cyclones. In these cases clearly the longitude of where the cyclone entered the Arctic, as well as location of maximum
intensity, would be the exactly the same. The Arctic was broken up into four subregions to aide in classification of storms. The cyclones were classified by the locations of which they reached maximum intensity. Figure 2.3 displays the four regions; (1) Oceanic region, which entails cyclones that reached maximum intensity polewards of 80°N; (2) North Atlantic region, which entails cyclones that reached maximum intensity between 70°N-80°N and 60°W-60°E; (3) Eurasian region, which entails cyclones that reached maximum intensity between 70°N-80°N and 60°E-180°E; (4) North American region, which entails cyclones that reached maximum intensity between 70°N-80°N and 180°W-60°W. A secondary analysis of where these cyclones emerged into the Arctic did not have certain regions, but simply counted by what longitude at 70°N the center entered into the region.

2.2.2 Long-term Climate Trends Within Arctic

In light of the rapid warming of the Arctic, reliable climate data going back to 1979 needed to be used to assess possible trends in the Arctic beyond simple warming, as has been discussed. The author did not want to use any data previous to 1979, as accuracy of this data could be called into question. Using CFSR data, several atmospheric variables were averaged over the region polewards of, and including, 70°N for every six-hour period through the year. Atmospheric variables examined include: precipitable water (PW), 500-hPa geopotential heights, 300-hPa U-wind, 700-hPa, 850-hPa, and 925-hPa temperature. This created an annual climatology, including the average, maximum value, and minimum value for every variable. Temporal trends were also examined; monthly averaged data from 1979-2013 allowed for significant trends to appear within the climatology, the most notable of these being PW and 850
hPa temperatures. Lapse rates between 925 hPa and 700 hPa were derived by subtracting the temperature at 700 hPa from the temperature at 925 hPa. Statistic significance was tested using the one-sample student’s t-test, given by \( z = \frac{\bar{y} - \mu}{\sigma / \sqrt{n}} \) where \( \bar{y} \) is the mean, \( \mu \) is the anticipated value, \( \sigma \) is the standard deviation, and \( n \) is the sample size. If any trend within the climatic record was not statistically significant at 95% confidence level, it was not included within this paper.

\[2.2.3 \text{ Eurasian Snowfall Climatology and Snow Growth Index}\]

Snow extent data goes back to 1966 and is produced weekly. Because of this, snowfall climatology is computed using the week number of year instead of monthly or daily climatology. The snow extent climatology is presented in percentage based on how frequently snow existed for a given point. Since the data is originally in binary format, only whole numbers could be used to plot the climatology. To avoid any issues associated with decimal points, each data point of 1 or 0 was multiplied by whatever factor was needed to equate the values to a percentage out of 100. For example, if looking at the 25-year climatology of snowfall for week 43, each data point was multiplied by four and added together. Thus, if 20 of the 25 years had snowfall for a given point this method displays 80%, and it is now easy to compare the snowcover frequency for varying periods of length. The Snow Growth Index (hereafter SGI) is the percentage snowcover over the region bounded by 40°N-70°N and 85°E-135°E. Analysis shows that the best
relationship between the AO and Eurasian snowcover is the sum of weekly percentage of snowcover for weeks 42-44 multiplied by the slope of weekly percentages from weeks 42-47. The index is then converted into sigma, using the standard deviation and average for all 34-years worth of SGI data. The data is inverted to produce a positive correlation between the two time series. The correlation has been tested for significance at the 95th% level.
Table 1. Average minimum MSLP and standard deviation in the Arctic for each month. Subtracting 2, 1.5, and 1 sigma from the mean allowed for identification of intense cyclones for each 6-hour period for that month. Counting of these “cyclone periods” allowed for analysis of trends within frequency of intense cyclones in the Arctic.
Figure 2.1. Location of radiosond launches on 5 August 2012 used in gathering data regarding atmospheric setup leading to intense cyclogenesis. Location was chosen because of its proximity to the cyclone during maximum intensification rate.
Figure 2.2. Mean Sea-Level Pressure (MSLP; hPa), 1000-500 hPa Thickness (dam), and Total Column Precipitable Water (mm) on 0600 UTC 4 Aug 2012. Two cyclone centers exist prior to intensification of intense cyclone. For tracking purposes, the longitude for the emergent low-pressure was used as it crossed 70°N, not the antecedent cyclone.
Figure 2.3. Cyclone classification regions for location of maximum intensity for the most intense cyclones by month and season. Region 1 is the Oceanic sector, Region 2 is the North Atlantic sector, Region 3 is the Eurasian sector, and Region 4 is the North American sector.
3. Results

3.1 Climatology

3.1.1 Intense Cyclone Identification and Classification

Cyclone periods were identified using the methodology described in section 2.2.1. The most intense cyclones tracked were those that were equal to or greater than 2 standard deviations below the monthly mean minimum sea-level pressure within the Arctic. A total of 1646 6-hour periods were identified, which is 3.3% of all periods within this study. The next-most intense cyclones, those that were 1.5 standard deviations or greater, accounted for 7.6% of all periods and entails 3,812 6-hour periods. The least intense cyclones tracked were greater than or equal to 1 standard deviation from the monthly mean. 7,940 periods met this criteria, which is 16.0% of the total periods in this study.

3.1.2 Temporal Trends

Figs. 3.1.1a, 3.1.1b, and 3.1.1c show the annual number of cyclone periods that were lower than 2-sigma, 1.5 sigma, and 1 sigma respectively for the period, 1979-2013. While there is no metric that analyzes vorticity, winds, or shear, for the sake of this study it is assumed that any pressure reading in the Arctic lower than 1 sigma is associated with a cyclone. Visual analysis of several 1-sigma intensity cyclones for each month reveals that the even the weakest of pressure depressions analyzed still contains three or more 4-hPa closed MSLP contours. The highest year in terms of cyclone activity was 1990 for cyclones of all intensities. This year was significantly above any other year in terms of cyclone activity. At the two-sigma level, 2005 was
the lowest year in terms of cyclone activity. The next-lowest year at this level was just one year later in 2006. 1987 was the lowest year in cyclone activity for the 1.5 and 1-sigma level. Interestingly, 2005 was not in the top-5 lowest years for these levels, indicating that while this year had a significant lack of very intense cyclones, it did not have a lack of less-intense cyclones.

While none of the trends are statistically significant at the 95\textsuperscript{th} percentile, there appears to be a diminishing trend for cyclones of all intensity thresholds. Looking at trend, the number of cyclone periods averaged above 50 in the late 1970’s and diminished to around 45 in recent years. The trend appears to lessen when looking at the lower-intensities, but speaking in terms of percentage they are very similar. Since there are a lot more cyclones at the lower intensities, it is harder to determine visual trends as the scale diminishes in size. Temporal trends in cyclone activity are difficult to determine since there is so much interannual and apparent interdecadal variability within the dataset.
Figure 3.1.1a, 3.1.1b, and 3.1.1c (top to bottom): The number of 6-hourly periods for each year since 1979 in which MSLP in the Arctic was below the monthly 2, 1.5, and 1 sigma pressure threshold, respectively. The red line indicates trend for the period. The scale for 3.1.1c is increased to 25 cyclone periods to compensate for increased number of cyclones.
3.1.3 Relationship to Arctic Oscillation

Several papers have mentioned that the phase and amplitude of the Arctic Oscillation may be indicative of Arctic cyclone activity (Thompson and Wallace 1998; Moritz et al. 2002; Serreze and Barrett 2007). Looking at Figs. 3.1.1a, 3.1.1b, and 3.1.1c, when observing the interannual variability within the temporal cyclone trends, there also appears to be an interdecadal element of variability. The 1990’s had the highest number of cyclone periods, followed by the 1980’s and lastly the 2000’s. The number of 1-sigma cyclone periods during the period from 1989-1995 were all above average. In fact, this 6-year period contains the first, second (tied with 2011), fourth, sixth, and seventh highest number of cyclone periods annually in for the entire 34-year period. This period of high Arctic cyclone activity corresponds well with an increase in the positive phase of the Arctic Oscillation in the early 1990’s.

The comparison of the AO and cyclone activity and its correlation is shown in Fig. 3.1.2. The AO is standardized and is the average monthly value from January-December for each year shown. The 1-sigma period departures are the total number of 6-hourly periods above or below the yearly average for the 34-year mean value. The correlation is strongest when the AO is extreme in either polarity. When the AO is plus (minus) 0.5, cyclone activity was correspondingly plus (minus) 50 cyclone periods at the one-sigma level. When the AO was closer to zero, correlation was lower. The result of this analysis further supports findings from earlier papers about the relationship between cyclone activity and the AO; if the average AO value is positive (negative), it will result in a greater ((lesser) number of periods of cyclone activity with a correlation of R=0.82 since 1979.
Figure 3.1.2: Comparison between yearly cyclone activity at the 1-sigma level and yearly Arctic Oscillation values from 1979-2012. AO values are shown in red with the scale on the right and cyclone period departures are shown in blue with the scale on the left. Correlation between the two exists at R=0.82.
3.1.4 Location of Maximum Intensity

The locations of intense Arctic cyclones were just as important as the frequency of intense cyclones. Fig. 3.1.3 shows the locations of the top-ten most intense cyclones to occur in the Arctic in each month for the entire year (January-December). Each star indicates the exact location of the cyclone center when each cyclone was at its maximum intensity north of and including 70°N. There is a definitive cluster of cyclone locations just to the east of Greenland extending towards Scandinavia. A secondary maxima appears over the open Arctic Ocean between 165°E and 140°W. No intense cyclones have every occurred in the Canadian Archipelago or the immediate coastline of far eastern Russia extending to the Northwest Territories of Canada. Cyclone distribution is fairly even from the coastline towards the open waters of the Arctic Ocean and from Scandinavia to central Russia. The intense cyclones just to the east of Greenland appear to reach maximum intensity within 5° latitude of entering the Arctic.

In order to depict seasonal changes with locations of most intense cyclones, Figs. 3.1.4a, 3.1.4b, 3.1.4c, and 3.1.4d show the locations of intense cyclones for spring (March-May), summer (June-August), fall (September-November) and winter (December-February), respectively. Intense cyclone locations in spring are clustered over the Greenland Sea and Scandinavian Sea, as well as another loose clustering of cyclones over the central Arctic Ocean and western Russia. No cyclones reached their maximum intensity lower than 80°N between the longitudes of 90°E and 50°W. One cyclone reached maximum intensity over Greenland, a region typically dominated by prolific high-pressure systems. The 30 most intense cyclones during the summer months are centered over the heart of the Arctic Ocean, with a secondary group that
appears to reach maximum intensity near the coast of central Russia. Only one cyclone reached maximum intensity in the region between Greenland and Scandinavia. Generally speaking, in the summertime cyclone distribution is the most evenly distributed across the Arctic. Also, the fewest number of cyclones reached maximum intensity in the lower latitudes of the Arctic. Instead, a good number of cyclones reached maximum intensity at very high latitudes, including some as far north as 88°N. The green star indicates the August 2012 case-study cyclone, which will be discussed in more detail in Ch. 3.3. Autumn witnesses a resurgence of intense cyclones east of Greenland, but still holds on to several intense cyclones occurring over the central Arctic Ocean. Compared to the summer, there are more cyclones on the Atlantic side of the Arctic versus the Pacific side. Only one cyclone reached maximum intensity after emerging of the coast of eastern Siberia. The winter is heavily dominated (73%) by cyclones occurring between Greenland and Scandinavia. A few cyclones occur just west of Greenland, as well as central Russia and the central Arctic Ocean. One cyclone again reached maximum intensity over the typically high-pressure dominated Greenland. The justification for the locations of the clusters of intense cyclones will be shown by looking at 500-hPa height and anomaly composites in Chapter 3.2 (Composites).
Figure 3.1.3: Location of maximum intensity for the 10 most-intense cyclones per month, denoted by the purple stars. All 12 months are shown, leading to 120 total locations.
Figures 3.1.4a, 3.1.4b, 3.1.4c, and 3.1.4d: As in Fig. 3.1.3, but broken up into seasons. Spring, summer, fall, and winter are shown, respectively. Green star located in 3.1.4b is case study cyclone of August 2012.
3.1.5 *Cyclone Classification by Region*

As discussed in Chapter 2, the Arctic was broken down into the (1) Oceanic region, (2) North Atlantic region, (3) Eurasian region, and (4) North American region to quantify favored regions of maximum intensification. Fig. 3.1.5 shows the distribution of where maximum cyclone intensity occurs in these regions throughout the entire year. The majority of cyclones reached maximum intensity in the North Atlantic region. Once again, seasonality is important when looking at Arctic cyclones.

Fig. 3.1.6a depicts the same as the previous figure, but for the spring. The N. Atlantic cyclones account for 80% of all cyclones, with the rest split (10%) between the Siberian region and the Oceanic region. Fig. 3.1.6b shows intense cyclones for the summer. The Oceanic region accounts for half of all intense cyclones, with the Siberian, N. American, and N. Atlantic regions, respectively, following suit. Intense cyclone distribution maximizes over the high-latitude oceanic and North Atlantic (90%) regions in the fall (Fig. 3.1.6c). Only 10% of all cyclones to occur in the fall reach maximum intensity in the Siberian sector. The North Atlantic sector is once again heavily dominated in the winter (Fig. 3.1.6d) with only 4,3, and 1 cyclones occurring over the Oceanic, Siberian, and North American sector, respectively.
Figure 3.1.5: Locations of where 10 most-intense cyclones, by month, occur throughout the entire year. There are a total of 120 cyclones, same as Fig. 3.1.3. Scale is on left by 10 from 0-80.
Figures 3.1.6a, 3.1.6b, 3.1.6c, and 3.1.6d: As in Fig. 3.1.5, broken up into seasons.
3.1.6 Entrance Locations for Intense Cyclones

As stated in previous chapters, intense cyclones formed in one of two ways; either they were very intense storm systems that reached maturity as they crossed the 70°N threshold, or they were the result of an emergent low from lower latitudes combining forces with a parent low already in place over the Arctic. The result of cyclone track analysis is shown in Fig. 3.1.7. The deepest closed MSLP contours were traced backwards from when the cyclone reached maximum intensity to where it crossed the 70°N latitude. Cyclones that did not maintain at least one closed contour were not counted in this study. Shown in Fig. 3.1.7 is a histogram depicting the number 2-sigma intensity cyclones and the longitude of where the low, or the emergent low, crossed into the Arctic (70°N). Longitudes are broken up into 15° segments, leading to a total of 24 locations for crossing into the Arctic. The entire year (January-December) is shown in this figure, and once again the period of record is from 1979-2013. It is evident that a majority of cyclones throughout the year emerge into the Arctic from the North Atlantic between the longitudes 30°W to 15°E. The remainder of cyclones are spread out fairly evenly extending from 15°E to 165°W. Three sectors did not witness any 2-sigma cyclone crossings: 150°W-135°W, 120°W-105°W, and 105°W-90°W.

Generally speaking, cyclones throughout the year appear to enter into the Arctic between the North Atlantic and the Eurasian Arctic coastline. However, as will be shown, there is great seasonality regarding where cyclones enter the Arctic. Shown in Fig. 3.1.8a is the longitude distribution for cyclones entering the Arctic in Winter (DJF). The winter season is heavily dominated by cyclones that originate in the North Atlantic region. In fact,
47% of all 2-sigma intensity cyclones in the winter emerge between 0°-15°W. Statistically speaking, at random chance there is only a 4% probability of a cyclone emerging into the Arctic at any given longitude division. The fact that almost half of all cyclones in the winter emerge between 0°-15°W is a significant finding given random-chance probability and resultant prominence of cyclone track density. The westernmost cyclone crossing occurs between 75°-60°W and the easternmost crossing is between 45°-60°E. One “rogue” cyclone crossed between 165°E-180°, which statistically speaking (odds of occurrence) is the greatest outlier for any cyclone during any season.

Looking at spring (MAM; Fig. 3.1.8b), the dominance of North Atlantic cyclones begins to diminish. While still the most favored entrance region, the 0°-15°W sector accounts for roughly half of the cyclones of which it did for the winter at 23.7% of all cyclones. The longitudinal region from 60°E-135°E averaged 4.8 cyclone crossings per sector in spring, whereas these same regions in the winter observed a total of zero cyclones crossing in the entire region. There was once again a “rogue” cyclone crossing from 165°W-180°, but it was not as statistically anomalous as the winter cyclone.

The distribution of summertime (JJA) cyclone crossings is drastically different from any other season. Fig. 3.1.8c is the distribution of summer cyclones entering the Arctic. The dominance of the North Atlantic cyclones from winter and spring is completely extinguished; only 15.6% of cyclones emerged between 30°W and 15°E. The rest of the distribution of cyclones is dispersed between 15°E and 165°W at 79.2% of all summer cyclones. This is the region from western Russia through Alaska. Three cyclones were unusual because they entered the Arctic via the Canadian Archipelago between 90°-60°W. Separated from a natural baroclinic zone such as the coastline of northern Eurasia, further
work must be done to understand what factors lead to such intense cyclones in an unusual place.

The reemergence of the North Atlantic cyclones occurs in the fall, shown in Fig. 3.1.8d. The 0°-15°W sector is once again heavily favored (29% of all fall cyclones), followed by the neighboring regions to the east and west, respectively. Western Russia and Eastern Europe are much more favored for cyclone tracking than its Eastern Russian counterparts. A few cyclones were unusual, with one occurring between 150°E-165°E, two occurring from the dateline to 165°W, and one crossing between 135°W and 120°W. Overall, it appears that there are two main distributions of cyclone tracks into the Arctic. The first is the wintertime pattern, where most cyclones emerge into the Arctic from the North Atlantic. The second is the summertime pattern, where a majority of cyclones enter the Arctic in the Eurasian sector, followed by the North Atlantic sector. The transitional seasons of spring and fall appear to by exactly that; transitional between these two favored patterns, perhaps with a bias towards the wintertime distribution.
Figure 3.1.7: Longitude entrance-regions for where very intense (2-sigma) cyclones crossed the 70°N latitude band. The 0° longitude represents the Prime Meridian and the 180° longitude represents the International Date-line. Scale is on the left, contoured every 10 cyclones. Negative longitude values are in the Western Hemisphere and positive longitude values are in the Eastern Hemisphere.
Longitude of 70°N Crossings for 2 Sigma Cyclones-
Spring

Longitude of 70°N Crossings for 2 Sigma Cyclones-
Summer
Figures 3.1.8a, 3.1.8b, 3.1.8c, and 3.1.8d: As in Fig 3.1.7, but broken up into seasons. Shown is spring, summer, fall, winter, with scale contours of 5, 2, 5, and 5, respectively.
3.2 Composites

3.2.1 Geographic Relative Composites

As stated in chapter 2, the most-intense cyclones per season (N=15; N=60 for entire year) were used to create composites. Fig. 3.2.1 is the composite analysis at 500 hPa with geopotential heights (contoured; dam) and geopotential height anomalies (shaded; dam) with blue colors indicating negative height anomalies and red colors indicating positive height anomalies. The projection is a from a “satellite” viewpoint and is centered over 90°N. The vertical center is along the 90°W/90°E longitude while the horizontal center is along the Prime Meridian and the International Date Line. Continental features are outlined in the magenta color, and latitude/longitude lines are spaced every 10° and 15°, respectively. The lag is given at day 10, meaning that this figure is showing the daily averaged 500-hPa geopotential heights/anomalies 10 days before each cyclone reached maximum intensity. The 500-hPa geopotential height anomalies are relative to each date’s climatology; the daily anomaly for June 1st and August 31st is relative to each day respectively, not the entire summer’s average 500-hPa geopotential height for each grid point. The top left of the figure encompasses the 15 most intense spring cyclones, the top right is summer cyclones, the bottom left is fall cyclones, and the bottom right is the winter cyclones. What is apparent is that in all seasons lower heights exist across the highest latitudes, with general ridging occurring across the mid-latitudes. The most eye-catching feature between the four seasons at 10 days out is the strong ridge across eastern Russia and the Gulf of Alaska in the composites for spring. The combination of this ridge in close proximity with the trough across the Arctic Ocean results in a strong height gradient and a fair amount of cross-polar
flow. Synoptically speaking, 10 days out from lowest MSLP of these intense cyclones, ridging is evident across most of Europe and the eastern half of the United States. This is supportive of the earlier finding that intense Arctic cyclones occur more frequently in a positive AO background state, which once again features negative height anomalies over the high-latitudes and positive height anomalies towards the mid-latitudes.

Moving forward temporally to a 7-day lag, Fig. 3.2.2 shows the same variables and projection as Fig. 3.2.1 except for the difference in lag. Once again, the most notable and significant feature between the four panels occurs in the springtime composites. Relative to 10 days lag, the ridge has intensified and moved even further poleward, while the trough has deepened and is centered right over the North Pole. Cross-polar flow at 500 hPa is clearly visible, as a strong jet can be assumed to exist from eastern Russia into northern Canada per thermal wind arguments as discussed in Chapter 2 of this paper. Cross-polar flow is notable because it is not a typical occurrence over the Arctic. Typically the Arctic is surrounded by a vortex at 500 hPa during all seasons, so cross-polar flow is a significant departure from the norm. Low heights are prevalent over the Arctic in the rest of the seasons as well. Another interesting feature is a deep trough over the eastern U.S. in the winter composites. Seeing as this is the coldest time of the year, a trough in this location would have the greatest impact both economically and socially. Also, as has been discussed, the eastern Seaboard of the United States is a breeding ground for very intense, rapidly deepening cyclones. Lower heights over the continent of North America would aide in cyclogenesis. It would appear that an intense cyclone occurrence in the Arctic is likely preceded by an intense cyclone, or at least an environment conducive for intense cyclones, across the coastline of eastern North America.
Moving forwards to 4 days prior to the event (Fig. 3.2.3), disparities begin to emerge between summer and the rest of the three seasons. Fall, winter, and spring all begin to feature a developing trough across Greenland and northeastern Canada accompanied with weak ridging equatorwards of this trough. Spring, and to a lesser extent fall, also feature a pronounced Rossby wave train across central Russia. This enhanced meridional movement of the 500-hPa-height field would promote cyclogenesis across Eurasia, perhaps suggesting a source for intense cyclones that enter the Arctic in this region. The summer composites do not have an intense trough over Greenland, but instead over the Eurasian side of the Arctic Ocean. While intense ridging is not evident in the composites, ample sunshine this time of year across the vast continent of Eurasia combined with the positioning of this trough can help enhance the natural baroclinic zone in the region.

Fig. 3.2.4 and Fig. 3.2.5 show the lag=1 and lag=0, respectively, for the intense cyclones. For both figures, the disparity between summertime cyclones and the rest of the seasons grows. The fall, winter, and spring seasons all show a pattern indicative of a very intense cyclone to over the far North Atlantic. An exceptionally deep trough reinforces cold air on the poleward side of the cyclones, while a very prominent ridge exists on the warm side of the cyclone. In the summer, however, no intense trough exists over Greenland, instead an elongated trough extends from Arctic coast in western Russia to the open Arctic Ocean polewards of Alaska. Weak ridging exists over the continental regions of Eurasia and the Gulf of Alaska in the summer composites. The winter composite is entirely dominated by an intense trough extending from the Canadian Archipelago to Greenland. A broad ridge extends from the Great Lakes all the way to Eastern Europe. This same pattern across the North Atlantic is also apparent in the spring and fall. However, unlike the winter, the spring
and fall also have troughing across the Arctic Ocean and ridging in interior regions of Eurasia. This could explain why these seasons are not quite as North Atlantic-dominant in terms of location of intense cyclones compared to the winter. However, the biggest similarity between all four seasons is that they all have a highly positive AO pattern. Just like at 10 days lag, lower geopotential height anomalies occur over the high-latitudes, with higher geopotential height anomalies equatorwards of these low heights, indicative of a classic highly positive Arctic Oscillation.

Looking forwards to see what occurs after these intense cyclones, Fig. 3.2.6 is the composites for 500 hPa geopotential heights and anomalies 9 days after the occurrence of the event. Trough amplitude across the high-latitudes has diminished, but still exists around Greenland in fall, winter, and spring. It is also noteworthy that in the fall, winter, and spring composites, eastern North America is influenced by ridging 9 days after the event. Since troughing over this area 9 days before event is replaced by ridging 9 days after the event, intense Arctic cyclones could signal a transition from cooler than normal temperatures across the eastern half of the U.S. to warmer than normal temperatures. These seasons also show a trough over the subtropical regions of the eastern Pacific. This arrangement would help to enhance the subtropical jet and could produce an atmospheric river into California and the rest of the West Coast. This is especially notable since this type of pattern would help the prolonged drought across this area, were it to set up this upcoming cool season. The summertime composite shows broad troughing across the Arctic Ocean with no other significant features. However, lower heights across the Arctic Ocean, combined with the warmer continental airmass still promote increased cyclone activity in the Arctic. The winter composite also has a Rex-block signal over eastern Russia
with higher heights located polewards of lower heights in the North Pacific. Springtime composite once again has the most distinguished feature, as a deep trough in Eastern Europe is immediately upstream to an intense ridge in central Russia. Recognition of these patterns following intense Arctic cyclones could aide in forecasting across US and Eurasia up to 10 days after an intense cyclone is observed in the Arctic.
Figure 3.2.1: Composite geopotential heights and anomalies at 500 hPa for 15 most-intense cyclones 10 days prior to maximum intensity. Figure is broken up into seasons, with spring, summer, fall, and winter in the top-left, top-right, bottom-left, and bottom-right, respectively. Blue(red) shading indicates 500 hPa heights below(above) daily average. Mean geopotential heights are solid lines, contoured every 6 dam.
Figure 3.2.2: As in Fig. 3.2.1, except for Lag=7 days prior to event.
Figure 3.2.3: As in Fig. 3.2.1, except for Lag=4 days prior to event.
Figure 3.2.4: As in Fig. 3.2.1, except for Lag=1 day prior to event.
Figure 3.2.5: As in Fig. 3.2.1, except for Lag=0, day of event.
Figure 3.2.6: As in Figure 3.2.1, except for Lag=-9 days after event.
3.2.2 Cyclone Relative Composites

As will be discussed in chapter 4, the result of the lag-composites was very useful in determining what features aide in cyclogenesis of Arctic cyclones, especially in the fall, winter, and spring. However, it was much more difficult to ascertain what exactly preludes Arctic cyclones in the summertime besides a basin-wide trough across the Arctic. Fig. 3.2.7 once again shows the locations of the 30 deepest summer cyclone locations in the Arctic. One observation made is that there appear to be two distinct groups of intense cyclones. Group 1, denoted by the red circle on the left, encompasses cyclones that reached maximum intensity over, or near to, the coastline of Russia. These will be hereby classified as the coastal group. The second group, denoted by the number 2 and the red circle on the right, is the oceanic group. These cyclones reached maximum intensity well away from the coastline, and perhaps away from the baroclinic influences produced by the varying thermal properties of the warm continent and the marine-cooled air across the Arctic Ocean.

Fig. 3.2.8 (1) depicts the storm-relative composites for Group 1 (Coastal) cyclones at maximum intensity. The location is irrelevant, but is used for scale and comparison purposes and is located in central Russia just west of the Laptev Sea and the islands of Severnaya Zemlya. Shown in the figure are the MSLP (contours, every 4 hPa), and 300-hPa wind (barbs, kts; shaded, kts). The coastal group of cyclones reach a minimum MSLP of at least 984 hPa, aided by a large 300-hPa jet with winds of over 50 kts in the left exit region of a cyclonically-curved jet streak. The oceanic group storm-relative composite at maximum intensity is produced in Fig. 3.2.8(2). The same exact features are plotted as in the previous figure. The MSLP of these cyclones is lower than 984 hPa and is aided by a strong jet at 300 hPa. However, the breadth of
50+kt winds at 300 hPa is not as expansive. However, the cyclone is much more enveloped by the jet streak, perhaps increasing upper-air support for cyclonic flow.

Looking at Q-vectors, Figs. 3.2.9(1) and 3.2.9(2) show regions of Q-vector convergence (divergence) in the warm (cool) colors, contoured every 2 units. Solid lines are the MSLP and the dashed lines are 700 hPa heights contoured every 3 dam. Looking at the two figures, the amount of Q-vector convergence is much greater for the coastal group than with the oceanic group. The distribution of convergence ahead of the motion of the cyclone in the coastal group and aligned with the coastline could show direct contribution from the baroclinic zone associated with the Eurasian coast in the Q-vector field.

500 hPa mean height and anomalies are described in Figs. 3.2.10(1) and 3.2.10(2). The solid black contours are geopotential heights contoured every 6 dam, and the shaded areas are the height anomalies for each point. The height anomalies use the departure from the daily mean, and are standardized. The coastal group has heights lower than 528 dam, which produces a 2.5-sigma negative anomaly. The oceanic group is driven by a deeper trough at 500 hPa, with heights down to 516 dam and a 3.0-sigma negative height anomaly. For the coastal composite, a weak ridge is situated equatorwards from the deep trough, while the ridge in the oceanic composite is oriented further away and not as expansive. It appears that at 500 hPa the oceanic cyclone group is better supported aloft than the coastal cyclone group.

Moving closer to the surface, 850 hPa (Figs. 3.2.11(1) and 3.2.11(2)), 850 hPa heights (solid black, contoured every 4 dam), winds (barbs, kts), temperatures (red, contoured every 2°C), and temperature anomaly (shaded, °C) were examined for the two groups. The center of the cyclones for the coastal group appears to be in a much more baroclinic environment than the oceanic group. The coastal cyclone group is negatively tilted and very mature, with +8°C
temperature anomalies north of the center and over -10°C temperature anomalies south of the composite cyclone. Compared to the oceanic cyclone group, these temperature anomalies are exceptional. Group 2 features at the lowest a -6°C temperature anomaly and only up to a +4°C anomaly at the highest. More cold air at 850 hPa is evident in the oceanic group, as the area of -4°C temperatures is much larger. The position of the oceanic group of cyclones could aide in production of cold air at 850 hPa, as these cyclones are located over more of the ice-cooled air mass of the Arctic Ocean. The cold air advection region of the coastal group is either over warmer continental regions, or over the Arctic Ocean in areas that are much less prone to be ice-covered.
Figure 3.2.7: Locations of the 30 most-intense cyclones to occur in the summer months, as has been shown in Fig. 3.1.4b. The red circles indicate the cyclones used for the coastal group 1, and the oceanic group 2. Group 1 contains 13 cyclones, and group 2 also has 13 cyclones. The green star denotes the case-study cyclone of August 2012.
Figure 3.2.8: Storm-relative composites during maximum intensity showing 300-hPa winds (barbs; shaded in blue) and MSLP (contours) for coastal group 1(top) and oceanic group 2(bottom). Center of cyclone has no significance geographically.
Figure 3.2.9: As in Fig. 3.2.8, except for 500-hPa heights (contours; every 6 dam) and standardized anomalies (shaded; every 0.5 sigma).
Figure 3.2.10: As in Fig. 3.2.8, except for 700-hPa Q-vectors (arrows and shaded, red (blue) indicates convergence(divergence)), MSLP (solid contours), and 700-hPa height (dashed contours).
Figure 3.2.11: As in Fig. 3.2.8, except for 850-hPa temperatures (red contours), 850-hPa temperature anomalies (shaded; cold(warm) colors indicate below(above) normal temperatures), 850 hPa heights (black lines; contoured every 4 dam), and 850 hPa winds (barbs).
3.3 Intense Cyclone of August 2012 Case Study

3.3.1 Overview

The cyclone of early August 2012 was chosen for analysis because it was a massive, intense cyclone centered at an incredibly high latitude of 82°N when it attained maximum intensity. Moving off the coast of eastern Siberia and merging with a parent cyclone over the Arctic Ocean, the cyclone deepened significantly and reached a maximum intensity of at least 964 hPa on 1800 UTC 06 August. At the time it was theorized to be the most intense cyclone to occur in the Arctic during the month of August. Confirmation of this inference is provided in Fig. 3.3.1. The graph begins on January 1st and ends on December 31st. Using the 34 years of data since 1979, each point on the graph represents each 6-hourly period for the entire year. The black line indicates the average minimum MSLP in the Arctic, while the red line indicates the absolute minimum MSLP in the Arctic. Cyclones are on average found to be most intense in January and least intense in May. The most intense cyclone to ever occur was on 1200 UTC 08 December 1993 (Fig. 3.3.2). The geographic North Pole is centered in the image, with continents outlined in purple. Solid black lines are the MSLP contoured every 4 hPa, and the dashed lines are 1000-500 hPa thicknesses (blue is less than or equal to 540 dam, red is greater than 540 dam). Shaded is precipitable water in mm, starting at 20 mm. This is a classic intense North Atlantic cyclone, centered at 70°N with a minimum MSLP of 934 hPa. The tightening of the thickness field indicates immense baroclinicity associated with the very deep cyclone. Going back to Fig. 3.3.1, drawn on the graph is a big black circle on the absolute minimum MSLP line. Looking at the rest of the month of August, this point appears to be the lowest surface pressure observed for that month. This point is actually the lowest MSLP for the entire summer (JJA).
Figure 3.3.3 is the same image as before, but only for the month of August. Plotting the lowest MSLP in the month of August in 2012 yields our result that this is indeed not only the most intense Arctic storm to occur in the month of August, but for the entire summer. The cyclone had many impacts, including erosion of the Arctic coastline due to incredible wave action attributed to the cyclone and the winds it produced. Afterwards, the rate of loss of sea-ice extent increased dramatically, leading to a new record minimum sea-ice extent the following September (Fig. 3.3.4; NSIDC). This was not all bad, however, as for the first time in recorded history the Northwest Passage was finally opened up through the Canadian Archipelago.
Figure 3.3.1: Climatology of minimum MSLP in the Arctic from 1979-2013. Chart begins on 01 Jan and ends on 31 December with every point representing one 6-hour period. Black line is the average minimum MSLP over 34-year period, with red line being the absolute minimum MSLP over said period.
Figure 3.3.2: The most intense storm to ever occur in the Arctic on 19 December 1993. Shown is the MSLP (black lines; contoured every 4 hPa), 1000-500 hPa thickness (dashed lines; contoured every 6 dam), and precipitable water (shaded). Lowest MSLP is 934 hPa.
Figure 3.3.3: As in Fig. 3.3.1, except only for the month of August. Thin black line is the average minimum MSLP and thick black line is the absolute lowest MSLP for each 6-hour period from 1979-2013. Red line plots the minimum MSLP for every 6-hourly period in 2012 in August. Scale is on the left, and is MSLP contoured every 4 hPa.
Figure 3.3.4: Daily areal Arctic sea-ice extent for June-October. Shown is the 1979-2000 average with 2 sigma (grey) and the three lowest minimum sea-ice extent on record through 15 September 2012. With a value much less than the previous minimum record set in 2007, 2012 set the record for lowest areal sea-ice extent since 1979 [adapted in part from National Snow and Ice Data Center: http://nsidc.org/arcticseaicenews/2012/09/arctic-sea-ice-extent-settles-at-record-seasonal-minimum/].
3.3.2 *Antecedent Conditions*

Prior to cyclogenesis, conditions across the Eurasian continent were very conducive for intense cyclogenesis. Fig 3.3.5 displays 300 hPa heights (dam) and anomalies for the week preceding cyclogenesis of the August cyclone. Ridging is prevalent across the entire Eurasian continent with positive height anomalies upwards of 120 dam. To the north of this expansive ridge is an equally expansive trough with negative anomalies upwards of 100 dam, especially just to the north of central Russia. It is interesting to note that the positioning of the baroclinic zone aligns perfectly with the thermal boundary that exists along the coastline of the Arctic Ocean. Warm, humid air over the continent is enhanced by the ridge, while dry, cool air marine air is reinforced by this trough. Owing to the thermal wind equation, an enhanced jet stream can be expected in this region of the globe. Moving lower in the atmosphere, Fig. 3.3.6 indicates the level of warmth across eastern Russia. For the week prior to cyclogenesis, 850 hPa mean temperatures (solid, °C) are shown along with 850 hPa temperature anomalies (shaded, °C). A very impressive region of +8°C temperature anomaly is in place through the eastern fifth of Russia. Keep in mind that these are the average temperature anomalies for an entire week; this intense heat was very pronounced for an extended period of time. Cooler than normal temperatures are found along the central Russian coastline and the open water of the Arctic Ocean polewards of eastern Russia. The thermal boundary found at 300 hPa extends all the way down to 850 hPa and is indicative of the extreme baroclinicity that aided in cyclogenesis of this intense cyclone.
Figure 3.3.5: 300 hPa heights (left) and anomalies (right) in dam for the first week of August 2012.

Figure 3.3.6: 850 hPa temperatures (left) and anomalies (right) in kelvin for the first week of August 2012.
3.3.3 Development of Intense Cyclone

With the atmosphere primed for intense cyclogenesis, all it needed was a trigger. The trigger for cyclone development was a digging trough over central Russia. 500-hPa geopotential heights on 0000 UTC 03 Aug 2012 are produced in Fig. 3.3.7a; contoured are 500-hPa heights in dam and shaded is the height anomalies given in m. Fig. 3.3.7b is for the same time period but shows the MSLP (contoured, hPa), thickness (dashed, dam) and total columnar precipitable water (shaded, mm). A deep elongated trough at 500 hPa extends from the Laptev Sea to eastern Kazakhstan. Combined with ridging to the east of this trough, southwesterly flow pushes warm, moist (over 40 mm PW) air up to the Arctic Ocean along this baroclinic zone. A low pressure is beginning to form over central Russia aided by jet stream dynamics. Fig. 3.3.8a is the 300-hPa heights (contoured, dam), wind speed (shaded, kts) and 500-hPa vertical velocity (blue contours, pa/s). The location of the developing cyclone coincides perfectly with positioning of the right-entrance region of a powerful uncurved jet streak upwards of 120 kts. 24 hours later, jet stream dynamics become even more favorable for intensification of the cyclone. Fig. 3.3.8b is the same map as 3.3.8a just one day later on 0000 UTC 04 August 2012. Highlighted in black is a beautifully coupled jet stream system. Powerful winds with two jet streaks of 140 kts and 120 kts, respectively, are positioned perfectly such that the jet streaks can be described as coupled. Not only is the cyclone in the right-entrance region of a 140 kt. jet, but in the left exit region of a 120 kt jet. Steep ascent of air parcels at 500 hPa indicates continued deepening of the cyclone. Confirmation of this remarkable ascent is shown in Fig. 3.3.9 that shows the NOAA HYSPLIT model representation of the LaGrangian viewpoint of air parcels located within the warm sector of the cyclone. The exact location chosen for analysis was the northernmost extent of the tongue.
of high precipitable water values greater than 20 mm and the greatest thickness values closest to the center of the cyclone. The center of the cyclone is indicated by the red letter ‘L’ in the figure. The backwards trajectories are chosen from 1200 UTC 05 August 2012 as this is the period of most rapid intensification. Air parcels located at 7000, 5000, and 3000 meters above ground level are shown in green, blue, and red lines, respectively. The source region for these levels going back five days is central Russia and eastern interior China. While surely these locales are far from tropical, precipitable water values of over 30mm indicate these air parcels are laden with moisture.

Ascent is observed for all of the height levels, especially during the period from 0000UTC to 1200 UTC 05 August. Steep ascent was also favored thermodynamically. Fig. 3.3.10 is a sounding from far northeastern Russia (see inset) in Tiksi on 1200UTC 05 August 2012. Located close to the coast of the Arctic Ocean, this station observes surface temperatures on 05 August at 1200 UTC near 20°C amid very moist air with precipitable water values close to 40mm. The amount of Convective Available Potential Energy (CAPE) is close to 900 J/Kg. It was difficult to find any studies analyzing CAPE in very high latitudes, but CAPE values this high are likely extremely unusual this far north. 900 J/Kg indicate an environment of instability and buoyant lift, regardless of what latitude it is found. 1000-500 hPa shear is close to 60 kts, indicating the potential for convective activity given the amount of buoyancy in the atmosphere. Fig. 3.3.11 is an infrared satellite image taken from polar orbiting satellites on 2100UTC 04 August 2012, courtesy of University of Wisconsin-Madison Antarctic Meteorological Research Center for the Arctic composite data set (Kohrs et al. 2014). Approximate locations of the two low pressures in this case study are shown, with cold, occluded, and warm fronts shown in blue, purple, and red lines, respectively. The expansive nature of very cold cloud tops along the cold
front is indicative of possible thunderstorm activity associated with the cold-frontal passage. Also, it is incredible to note how far south this cold front appears to exist, stretching from northeastern Siberia all the way to China, a distance of over 2,000 miles. Model analysis, balloon soundings, and satellite data all agree that instability, Jetstream dynamics, and convection along the cold front worked in concert to deepen this cyclone explosively. The satellite image shown in Fig. 3.3.12 is the same as Fig. 3.3.11, except for 0000 UTC 06 August 2012. A large, mature, and intense low pressure can be seen over the waters of the Arctic Ocean. The deepening phase of the cyclone commences on 1200 UTC 06 August 2012, as shown in Figs. 3.3.13a and 3.3.13b (same variables as Figs. 3.3.7a and 3.3.7b, respectively). The center of low pressure begins to separate from the baroclinic zone that aided in its intensification as it becomes enveloped in an equivalent barotropic environment. Heights at 500 hPa are much below climatology, with a minimum height value of 506 dam, which is over 30 dam below normal. Moist air treks to incredibly high latitude, as 25+mm precipitable water values are observed polewards of 80°N. Moist, humid air is very efficient at melting ice and the fact that such plentiful moisture worked its way so far north could have contributed to sea-ice loss.
Figures 3.3.7a and 3.3.7b: 500 hPa heights (contoured, dam) and anomalies (shaded; warm (cool) colors indicated geopotential heights above(below) normal, dam) on the left, and MSLP (solid black lines; contoured every 4 hPa), 1000-500 hPa thickness (dashed lines, dam), and precipitable water (shaded, mm). Location of developing cyclone shown with red ‘L’.
Figures 3.3.8a and 3.3.8b: 300 hPa geopotential heights (black lines, every 6 dam), winds (shaded, kts), and 500 hPa vertical velocity (blue contours, pa/s). Figure 3.3.7a (left) is on 0000UTC 03 Aug 2012 and Figure 3.3.7b is 24 hours later on 0000UTC 04 Aug 2012. The black circle highlights the coupled nature of two jets at 300 hPa.
Figure 3.3.9: 5-day NOAA HYSPLIT backwards trajectories from 1200UTC 05 August 2012 for air parcels at the 7000, 5000, and 3000 meters above ground level shown in green, blue, and red lines, respectively. Red ‘L’ is approximate location of center of low pressure at given time.
Figure 3.3.10: Radiosonde data taken on 1200UTC 05 August 2012 at Tiksi, Russia, as located by the black dot in inset on upper-right. Skew-t plot is shown, with temperature in °C on bottom and height (hPa) on left. The black line is the temperature, blue line is dew point, and the dashed red line indicates the saturation-cooling rate taken from LCL. Winds are shown as the barbs along the right side of the Skew-t.
Figure 3.3.11: Infrared satellite image taken on 2100UTC 04 August 2012. Locations of low pressure is shown with red ‘L’s, and cold, occluded, and warm fronts are indicated with blue, purple, and red lines, respectively. White cloud tops along cold front indicate very cold clouds, indicative of convection.
Figure 3.3.12: As in Figure 3.3.10, except on 0000UTC 06 August 2012. Image was taken just as cyclone was reaching maximum intensity.

Figures 3.3.13a and 3.3.13b: As in Fig. 3.3.6a and 3.3.6b, except on 1200UTC 06 August 2012.
3.3.4 Impact of Cyclone on Sea-Ice Loss

Along with beach erosion, one of the largest impacts of the cyclone in the Arctic was the loss of sea-ice. Besides the incredible strength of this cyclone, the amount of ice loss during the summer of 2012 was a major news headline across world media. It was more than just a coincidence that the all-time minimum sea-ice extent occurred the same summer as the most intense cyclone to ever occur in the Arctic. Figs. 3.3.14a, 3.3.14b, and 3.3.14c show MSLP (dashed black contours, hPa), 10m-winds (barbs, kts) and sea-ice concentration (shaded, %) on 0000 UTC August 04, 06, and 10, respectively. While sea-ice extent was already well below normal by the time the cyclone emerged into the Arctic, it would be significantly reduced by this cyclone as it began to emerge off the Siberian coast on 04 Aug 2012. Winds at 10 m across the Arctic are light, with a few locations north of Alaska observing 20 kt winds. However, as the cyclone progressed and intensified, winds would significantly increase. By 06 August, the cyclone is nearing maximum intensity and is a remarkable 967 hPa cyclone located at 81°N (Fig. 3.3.14b). Winds are now sustained at 30 kts over a large area of the Arctic Ocean.

It is also important to note the orientation of the wind field. It is perpendicular to the edge of the ice sheet, which means the potential of the wind to move and compress the ice is at a maximum. The wind also results in extreme wave action, which breaks up the ice and makes it easier to melt through the rest of the summer. The winds and resultant wave action also act to mix the upper levels of the Arctic Ocean. This is important because in the Arctic, warmer water typically lies below cooler waters at the surface in the summer. This is because the salinity of the uppermost layer of the Arctic Ocean is lower than the subsurface ocean water. When the ice melts into the Arctic Ocean, it is unable to sink due to the salinity barrier and remains near the
surface. Thanks to the intense cyclone mixing the basin, these warmer waters in the subsurface ocean can emerge on the surface. Sea-surface temperatures further warm, and ice loss is accelerated. Finally, of importance is the length of which the cyclone remained in the Arctic. As can be seen on 0000 UTC 10 August (Fig. 3.3.14c), the cyclone is much weaker but has remained relatively stationary for the past four days. Winds have lessened, but there still exists a broad area of 20 kt winds pushing up against the edge of a significantly diminished ice sheet. The cyclone would remain for another three days over the Arctic before moving towards North America.

The total impact of the cyclone can be seen in Fig. 3.3.15, as the percentage change of sea-ice extent is shown for the 10 day period between the 1st and 11th of August. Red colors indicated a loss of sea-ice, and blue colors indicated a gain of sea-ice greater than 10%. A huge area of the Arctic lost a large percentage of sea ice, especially over the East Siberian Sea to the Beaufort Sea. Any locations in the darker reds were ones that likely lost the entirety of its ice concentration. The thick gold line shows the track of the cyclone from 01 August through maximum intensity on 06 August. The location of maximum ice loss is aligned very well with the track of the cyclone. To help quantify the impact the cyclone had on sea-ice loss, Fig. 3.3.16 is the percentage of sea-ice loss in the Arctic from the seasonal maximum in March through the seasonal minimum in September. The summer of 2012 lost an extraordinary 79% of its ice extent through the melt season. The next closest year was 2007 with 69% ice loss, and in the 1980’s the average percentage of ice-loss through and Arctic summer was 51%. During the first 2-week period of August 2012, almost 8% of the entire ice sheet was lost. The case can be made that although the summer of 2012 would have been an incredible year for ice loss, the intense cyclone to occur in early August helped push Arctic ice-loss into uncharted territory.
Figures 3.3.14a, 3.3.14b, and 3.3.14c: Percentage of sea-ice concentration as shown in percentage (shaded, every 20%), MSLP (dashed lines, every 4 hPa), and 10m winds (barbs, kts). Times are noted as 0000UTC on August 04 (a), August 06 (b) and August 10 (c).
Figure 3.3.15: Sea-ice concentration change from August 01 to August 11 2012. Red colors indicate a loss of sea-ice percentage and blue colors indicated a gain of sea-ice greater than 10%. The gold line is the track of the cyclone from 01 Aug to maximum intensity on 06 Aug.
Figure 3.3.16: Percentage of Arctic sea-ice loss each year from the climatological peak in March through the climatological low in September. Shown in red is the amount of sea-ice that was lost during the first two weeks of August 2012.
3.3.5 Impact on North America

Another possible impact of the Arctic cyclone was to help break down the North American upper-level ridge that had been in place from late spring through the majority of the summer. The persistent ridge resulted in a lack of rainfall and record heat that was significantly impacting United States crop conditions negatively. However, welcome heat relief would come in the middle of August as the remnant vortex from the Arctic cyclone at 500 hPa moved into central Canada. As seen in the 500-hPa height anomalies in Fig. 3.3.17a on 1200 UTC 15 August 2012, a positively tilted shortwave digs into the upper Midwest in response to the deep trough over north-central Canada. Two days later (Fig. 3.3.17b) the shortwave has rounded the base of the trough, deepened and has become negatively tilted, and will eventually cut off and produce cooler temperatures and rainfall over areas of the US that had barely seen rain all summer. While there is no way to definitively prove the Arctic cyclone was directly responsible for crops-salvaging rain without additional analysis and numerical modeling, there is strong evidence that without its influences the shortwave responsible for the trough might not have dug as far south. Its associated rainfall would have thus missed the major agricultural region of the US and the crops would have suffered even more than they already had.
Figures 3.3.17a and 3.3.17b: As in Fig. 3.3.7a, but for the North American projection and on 1200 UTC 15 Aug 2012 and 1200 UTC 17 Aug 2012.
3.4 Snowcover and its Relationship to the AO

3.4.1 Snow Growth Index

The motivation for looking at Eurasian snowcover has been discussed in Ch. 1. As snowcover across Eurasia grows rapidly and expands, it strengthens the Siberian high-pressure system and the associated baroclinic zone with the warmer tropical Pacific basin. The increase in baroclinicity has very important seasonal repercussions due to the meridional nature of the thermal gradients. As has been discussed in Ch. 1.2.2, a stronger meridionally-based jet stream will occur in concert with a strong meridionally-based baroclinic zone, owing to the thermal wind equation. The right-hand side of the WAF equation (Ch. 1.2.4),

$$v' T' - \left[ 2 \chi \alpha \sin(2\phi) \right] - \frac{16(f')^2}{\beta^2};$$

is crucial for determining if upward wave flux will occur. A positive WAF value is necessary for upward wave flux, and as such the \(v' T'\) section of the WAF equation must be positive. Because of the meridional orientation of the eastern seaboard of Asia (and thus a meridional baroclinic zone in the cool season), it is quite frequent to get positive meridional heat advection such that the WAF equation will be positive. The amount of snowcover over Asia is important, because positive snowcover anomalies reinforce cold air in close proximity to the warmer West Pacific airmass. This intensifies the baroclinicity of eastern Asia, intensifying the positive temperature advection \((v' T'\) term. As wave flux increases, easterly momentum can be deposited into the upper troposphere and lower stratosphere with an attendant weakening of the Polar Vortex, resulting in a more negative Arctic Oscillation (AO) (Charney and Drazin 1961; Martius et al. 2009). A negative AO, especially during the winter months, may be associated with high-impact extreme weather events over North America. These extreme weather events can range from intense cold
air outbreaks, crop damage in subtropical regions of the US and Europe, and to prolific snowstorms. A negative wintertime AO is a major contributor to major weather challenges. Evidence of this link between the AO to Eurasia is explained in Fig. 3.4.1. This is the first empirical orthogonal function (EOF) of temperatures at 2m for winter (DJF) from 1979-2012. Explaining nearly 30% of the variance, it appears that very cold temperatures across central Eurasia can result in what appears to be a negative AO and cold temperatures across the mid-latitudes. Cold temperatures in Eurasia can be enforced by an early start to snow expansion in the fall, as solar radiation is lost to space more readily and cold temperatures are reinforced by the snowpack.

Using Fig. 3.4.1 as a guide for what part of Asia to focus the study on, correlations between October snowcover growth and the AO were calculated. Adjusting the region of study to various locations, it was found that correlations between the AO and snowcover tended to decrease significantly when longitudes less than 60°E were included in the analysis. The region of Eurasia that produces the best correlation between snowcover and the Arctic Oscillation was found to be the area from 40°N-70°N and 85°E-135°E. The best combination of rate of snow growth, combined with areal extent of snow, was found to be the multiplication of the integral extent of snow from weeks 41-44 and rate of snow expansion from weeks 40-45. When this value was compared the average AO value for the following winter (DJF), the correlation is shown in Fig. 3.4.2. The two values for Snow Growth Index (SGI) and AO are shown for each year from 1973-2013. The blue line is the SGI and the red line is the AO for the following winter. The correlation for the entire period is found to be R=0.49, however there are certain periods that exhibit greater correlation. The mid-to-late 1970’s appear to have significantly lower correlation than the rest of the period. This could be attributed to a few things; degradation of
data, as snow extent quality was not nearly as high of quality as it is in the modern satellite era and sampling differences between current-day techniques and those used in the 1970’s. When looking from the modern period of 1997-2013, correlation vastly improves to R=0.83 a statistically significant value. The yearly chart for this period is found in Fig. 3.4.3, same chart as for the 1973-2013 period. With the exception of a couple years like 2011 and 2005, the SGI does a pretty good job of accurately predicting the state of the AO for the following winter. A scatterplot is given in Figure 3.4.4, with the SGI on the X-axis and the AO index on the Y-axis. There is a clear correlation, with a variance of $R^2=0.61$. Values as high as this are generally accepted to be statistically useful in determining forecast skills in a model. It appears that the SGI works best at the extremes. When the value is either very high or very low, skill improves with forecasting the AO value. Values closer to zero can still be useful, but perhaps confidence decreases in years with these types of values.
Figure 3.4.1: Empirical Orthogonal Function (EOF) of 2m temperatures for winter (DJF). Cold (warm) colors indicate temperatures below (above) average for the winter period.
Figures 3.4.2 and 3.4.3: Snow growth index vs. AO index from 1973-2013 (top) and 1990-2013 (bottom). Red line shows the winter (DJF) AO and the red line is the SGI for the preceding October. SGI is on the left axis, and AO is on the right index. Both are standardized.
Figure 3.4.4: Scatterplot of AO vs. SGI with AO as the vertical axis and SGI as the horizontal axis.
3.4.2 Changes in Snowcover Extent

While examining the Snow Growth Index, it became evident there certain changes had taken place across Eurasia relating snow extent. When comparing average snowcover extent during the 1980’s, 1990’s, and 2000’s, each decade successively became more prone to snowcover in October with time. A clear example of this can be seen in Figs. 3.4.5a, 3.4.5b, and 3.4.5c, which show the percentage difference between the average snowcover extent between 2003-2012 and 1980-1989 in weeks 42-44 respectively. Week 42 is not overly impressive regarding frequency of areas with increased snowcover but more so the scope of territory that has seen increased snowcover extent over the past decade than during the 1980’s across Eurasia. Canada has also observed an increase in snowcover extent frequency, albeit a smaller scale owing to less land-area. Moving forward to week 43, a huge area of Eurasia has seen an increase of 30%-60% in percentage frequency of snowcover extent. Areas from Scandinavia to central and eastern Russia all have shown a huge increase of snowcover frequency. Even week 44 has significant increase in snowcover, although it is less expansive and confined to mostly China.

It is difficult to ascertain what is driving these changes, but a connection can be made to precipitable water. As more open water in Arctic Ocean for extended periods of time, more latent heat flux is transferred from the ocean to the atmosphere in the Arctic. Precipitable water trends are displayed in Fig. 3.4.6, which is the average precipitable water in the Arctic in the fall (SON) from 1979-2012. A statistically significant increasing trend is apparent, as values the past decade are higher than in the 1980’s. While an increase of 1mm of precipitable water may not seem like much, this is the increase over the entire Arctic basin including very dry Greenland. Also, because of the fact the Arctic is so cold and therefore dry, a 1 mm increase when the average is
between 5-6 mm is a significant increase percentage-wise. The link has been shown between decreasing sea-ice extent and increased snowcover. It is well known that there is significantly less ice extent of late, but these results show that there is also a correspondingly significant increase of snowcover. More work would need to be done to prove a definitive correlation between increasing Arctic moisture and increased snowcover, but these are signs pointing to a connection between the two.
Figures 3.4.5a, 3.4.5b, and 3.4.5c: Percentage difference in snowcover extent between the 2003-2012 period and the 1980-1989 across Eurasia. Blue (red) colors indicate an increase (decrease) of snowcover frequency contoured every 10%. Weeks shown are week 42, 43, and 44 in figure 3.4.5a, 3.4.5b, and 3.4.5c, respectively.
Figure 3.4.6: Average precipitable water (mm) across the Arctic for the fall (SON) from 1979-2012. While not statistically significant, a steady increasing trend is seen, as denoted by the solid black line. Precipitable water values are contoured every 0.5 mm beginning at 4 mm.
3.4.3 *Weekly Snowcover Extent Anomalies*

A final word on snowcover extent comes regarding anomalies on a weekly scale in a very user-friendly manner. Images can be found on the website [http://aturchioe.weebly.com](http://aturchioe.weebly.com) under the “Research Tools” link. Going back to 1970, weekly snowcover anomalies have been produced relative to the 25-year climatology from 1989-2013. Figs. 3.4.7a, 3.4.7b, 3.4.7c, and 3.4.7d show 2014 week 1 as an example of what the images look like. A broadened view of Eurasia and North America are shown, with more detailed insets of Europe and the United States available as well for more detailed study. Red colors indicate an absence of snowcover (negative snowcover extent) with the percent anomaly shaded every 10%. Blue anomalies show a positive snowcover extent contoured every 10% as well. The effect of synoptic events, such as midlatitude cyclones, on snowcover can be studied alongside the effects of El Nino Southern Oscillation, North Atlantic Oscillation, AO, and an endless variety of other indices.
Figures 3.4.7a, 3.4.7b, 3.4.7c, and 3.4.7d: Weekly anomalies from the 25-year climatology for week 1 2014. The Eurasian, North American, European, and United States projections are shown in Fig. 3.4.7a, 3.4.7b, 3.4.7c, and 3.4.7d, respectively. Blue (red) colors indicate positive (negative) anomalies. Contours are in percentages every 10%.
4. Discussion

4.1 Climatology

This study is the first since the modern satellite era to manually count the frequency and intensity of cyclones. Early studies (Keegan 1958) had manually tracked and counted cyclones, but the quality of the data was subject to error and was much more sparsely distributed than data available today. Several other studies (Zang et al. 2004; Serreze and Barrett 2008) used modern data, but only used computer algorithms to track cyclones that are naturally subject to error. Because the results of these studies cannot be manually quality-checked, any failures of the algorithm in identification of cyclones will be included within their findings. A vast majority of intense Arctic cyclones occurred in the North Atlantic sector of the Arctic (see Fig. 3.1.5). This finding is consistent with the results of Sanders and Gyakum (1980) who showed that the most intense cyclones in the Northern Hemisphere occur along the natural baroclinic zones along the eastern seabords of Asia and North America. Although this study did not consider bomb cyclones specifically, intense cyclogenesis occurred in the region of study that extended into the Arctic between Greenland and Scandinavia.

There exists a large seasonality regarding locations of maximum intensity of cyclones and where these cyclones enter into the Arctic (Wernli and Schwierz 2006; Allen et. al 2010). The winter season is heavily dominated by North Atlantic cyclones that reach maximum intensity typically in the N. Atlantic sector or the Arctic or the Oceanic sector of the Arctic. The transitional seasons of the spring and fall are still heavily influenced by North Atlantic cyclones, but there are a much greater number of cyclones that emerge over the Eurasian continent and reach maximum intensity fairly evenly in the N. Atlantic, Oceanic, and Eurasian sectors of the
Arctic. Summertime is witness to a much more evenly distributed clustering of Arctic cyclones. While there still exist cyclones that emerge into the Arctic via the North Atlantic region, a majority of intense cyclones now emerge off the coast of central and eastern Eurasia. A majority of summertime Arctic cyclones reach maximum intensity in the Oceanic region polewards of and equal to 80°N and in the Eurasian sector (see Fig. 3.1.8b). The development of Eurasian cyclones coincides well with the development of the warm-continent cool-ocean baroclinic zone along the northern Eurasian coastline. The seasonality can be attributed to the meridional movement of the jet stream throughout the year. In the winter, the polar jet stream (henceforth PJS) is much further equatorward than the Arctic regions. Because of this, most intense cyclones are most intense close to 70°N and are a direct result of the favored regions of the Northern Hemisphere for intense cyclones to occur and move poleward (Wernli and Schwierz 2006). The summer is driven by more in-situ development of intense cyclones. The PJS is much weaker in the summer and the overall intensity of cyclones is diminished (see Fig. 3.3.1). With the loss of jet-stream dynamics, a great source of cyclogenesis is the Eurasian coastline of the Arctic Ocean. Interior sections of Eurasia have the greatest seasonal temperature swing from cold winters to hot summers on earth (Gorczynski 1922). These extreme seasonal temperature differences are due to the continentality of Eurasia, with interior regions being incredibly distant from the temperature-moderating oceans. After bitterly cold winters, interior regions of Eurasia heat up very quickly in the spring. The air mass over the Arctic Ocean stays much cooler due to the impact of the ice sheet. The temperature contrast produced results in a very pronounced baroclinic zone along the northern coast of Eurasia that cyclones feed off for development.
4.2 Composites

4.2.1 Geographic Relative Composites

One goal of the geographic relative composites was to identify patterns that lead to the development of intense cyclones in the Arctic. As early as 10 days prior to development, a positive phase of the AO is evident, as lower than normal 500-hPa geopotential heights appear across the Arctic, and lower than normal 500-hPa geopotential heights exist in the mid-latitudes. The progression of troughs and ridges leading up to cyclone development give forecasters a better understanding of what primes intense cyclones in the Arctic. This study helps to gain a better understanding of where these troughs and ridges form and how they propagate. This allows for recognition of which regions of the globe are important to watch in anticipation of future intense cyclones in the Arctic. For the winter months, the development of a deep trough across Greenland and the Canadian Archipelago was the largest determinant of whether or not an intense cyclone would form. Pronounced ridging extending from the eastern Seaboard through eastern Europe was apparent. While it is difficult to ascertain the importance of this ridging regarding its contribution to rapid cyclone development, the associated baroclinic zone in the North Atlantic between the intense trough and broad ridging must play a large role in cyclone formation. The transitional seasons had this same trough, but it was extended further east towards central Eurasia. This promoted an eastward shift of where the greatest thermal gradients occurred. As expected, the longitudes of where intense cyclones entered the Arctic also shifted eastward (Figs. 3.1.8a and 3.1.8c). The summer composites were perhaps the most interesting, as it was much more difficult to ascertain large-scale features associated with cyclogenesis. The case-study cyclone of August 2012 had a very deep trough over central-eastern Eurasia, but
when looking at the composites only a weak trough is present. This indicates that this deep trough is being washed out by the 14 other geopotential height patterns. Generally speaking, a trough of lower heights somewhere over the Arctic Ocean was required for intense cyclone genesis. Ridging over continental Eurasia appeared to contribute to summertime Arctic cyclone formation, but once again did not have a distinct signal in the composites. Once again, the intense ridging with the August 2012 cyclone did not appear in the composites. Overall, it was difficult to determine the main driving factors for the development of intense cyclones in the summer for the top 15 cyclones. Perhaps a more distinct pattern in the 500-hPa geopotential height composites would appear when looking at the top 10 or 5 intense summer cyclones, but for this study only vague observations can be made. Greater clarity regarding driving factors for summertime cyclone development was seen when studying the cyclone-relative composites.

4.2.2 Cyclone Relative Composites

500-hPa geopotential height anomalies for summertime cyclones did not display a clear pattern (Fig. 3.2.5) in the geographic-relative composites that would help explain why and where these intense cyclones had formed. This could be because of the fact that there was greater spread regarding location, but regardless there are less concrete results that explain cyclogenesis in the summer versus all other seasons in the geographic-relative 500-hPa height composites. To help gain a better understanding of cyclogenesis in summer cyclones, two groups were formed. Cyclones that developed close to the Arctic coast were grouped together as one group, and cyclones that developed further out to sea were grouped together as another group. The main result of this analysis was that, especially at low-levels, the coastal cyclones are enhanced by a
more baroclinic environment than that of oceanic cyclones. The oceanic cyclones appear to be
produced and supported by a favorable environment in the higher levels of the atmosphere (i.e.
300-hPa jet stream and 500-hPa height anomalies). The favorable environment was due to a
stronger jet at 300 hPa, which appears to be coupled. The jet stream also encompasses more of
the cyclone vs. that of the coastal group. 500-hPa heights are 16 dam lower in the composites for
the oceanic group vs. the coastal group, adding to the assertion that the oceanic group had better
upper-level support for cyclogenesis. Relative to climatology, this group was more anomalous at
500-hPa as geopotential height anomalies were upwards of 3 sigma versus only 2.5 sigma for
that of the coastal group. Moving down to the 700 hPa level, support now favors the coastal
group. Q-vectors at 700-hPa showed that more forcing was evident in the coastal cyclone group
than the oceanic group. 700-hPa Q-vector convergence was much more evident downstream of
the coastal cyclone group, while divergence of 700-hPa Q-vectors was prominent upstream. The
oceanic group had a more sporadic pattern, with a lesser extent of 700-hPa Q-vector divergence
or convergence. Finally, looking towards the lower levels at 850-hPa, temperatures and
temperature anomalies were plotted. The low-level thermal gradient of the coastal group was
much greater than that of the oceanic group. The cold temperature anomalies to the south of the
cyclone, and warm temperature anomalies to the north of the cyclone, indicate that these
cyclones had much greater temperature advection patterns and stronger frontal boundaries.
Stronger temperature advection and Q-vector forcing at the lower levels indicate greater support
for intensification at the lower-levels for the coastal group. A stronger, coupled jet-stream along
with lower geopotential heights at 500-hPa indicate better support for intensification at the
upper-levels for the oceanic group. When looking at the big picture, the issue with climate
change is an important one to consider. In a warming climate, the continent of Eurasia is
expected to warm up fairly rapidly (Schubert et al. 2014; Bekryaev et al. 2010). However, the Arctic Ocean will remain relatively cool, especially while an extensive sea ice cover still exists. The reason the Arctic will remain cooler for longer is because as long as extensive ice cover exists in the spring, solar energy will go into melting the ice and latent heat release, as opposed to sensible heat release over Eurasia. No significant trend regarding cyclone intensity in the summertime was found, but this could change as the climate continues to warm.

4.3 Early August 2012 Cyclone Case Study

The intense cyclone of early August 2012 was chosen for analysis because it was the most intense cyclone to ever occur in the modern era (section 3.3.1; Simmonds and Rudeva 2012). Its impact on the loss of sea-ice is well documented (section 3.3.4) and led to increased awareness of the impacts that climate change is having on the globe. A minimum MSLP of 964 hPa in the August 2012 cyclone produced a greater than negative 4 standard deviation anomaly, was clearly an incredible event. The background state of a positive AO set the stage for the cyclone. Lower heights over the Arctic Ocean aligned perfectly with higher heights over the continent of Eurasia to contribute to the positive phase of the AO. The question needs to be asked whether the positioning of the enhanced baroclinic zone with the coastline of Eurasia was a coincidence. There is no easy way to answer this question, but it is fair to say that as the climate warms, the baroclinic zone along the Arctic Ocean and the continents will increase as the land warms faster than the increasing ice-free Arctic Ocean. The antecedent baroclinicity associated with the August 2012 cyclone was astounding. Prior to development during 4-6 August 2012, the 850-hPa anomalous temperature gradient was ~10°C/1000 km, which is clearly
a value well above the climatological norm. Since this baroclinic zone helped lead to the most intense Arctic summer cyclone on record, the case can be made that as this baroclinicity increases with time, so will the number of intense Arctic cyclones. It is yet to be seen if this increased baroclinicity will impact the future of intense cyclones in the Arctic, but it is a noble hypothesis to be made.

The intense cyclone of August 2012 was the result of a merging of two cyclones, something fairly typical with summertime Arctic cyclones. A 990-hPa low pressure was located over the open Arctic Ocean prior to 04 August 2012. A coupled jet stream and intense baroclinic zone aided in atmospheric ascent and eventual development of a secondary low pressure on 04 August 2012. The merger of this secondary cyclone with the parent cyclone produced the very intense cyclone of August 2012 that reached maximum intensity on 1800 UTC 06 August 2012. One unusual aspect of the cyclone relative to climatology is the longitude at which the storm crossed the 70°N latitude circle. While the intensity of this cyclone was not a major outlier relative to all other summer <2-sigma strength cyclones, only 3 other cyclone tracks crossed 70°N between 135°E-150°E between 1979 and 2013. The vast majority (>90%) of cyclones crossed 70°N further west than the August 2012 cyclone. The unusual eastern track entering the Arctic could have aided in its intense cyclogenesis. The developing cyclone was able to use the anomalously warm airmass over northeastern Asia as a source for warm, moisture-laden air. While difficult to definitively prove, synoptic analysis appears to show increased moisture and heat advecting into northeastern Asia from southern Europe and India. There also appear to have been moisture contributions from three landfalling typhoons in the western Pacific. Between the dates of 28 July and 09 August 2012, typhoons Saola, Damrey, and Haikui all made landfall in eastern China. Associated moisture surges into northeastern Asia followed all three landfalls,
analyzed using total-columnar precipitable water. Cold-air advection on the western side of the developing cyclone clashed with the very warm, unstable air mass over northeastern Russia. The eastward track maximized the baroclinic zone, as the ice and oceanic-cooled air was able to interact more prominently with the aforementioned warm, unstable air mass. After reaching maximum intensity, the vortex at 500 hPa witnessed upscale growth, encompassing the entirety of the Arctic basin. While open for discussion, baroclinic influences on the coast of the Arctic Ocean would have contributed to upscale growth of the cyclone. Temperatures at 850 hPa cooled from 3°C prior to cyclogenesis to -6°C after maturation of the cyclone. The pool of cool air at 850 hPa covered the entirety of the Arctic Ocean, while warm continental air surrounded the Arctic Ocean. Owing to the natural orientation of the landmass surrounding the Arctic Ocean, the greatest baroclinicity was located along the coastline. It seems to be more than a coincidence that the cyclone at 500 hPa undergoes upscale growth in concert with the enhancement of the coastal baroclinic zone. The remnant vortex then moved through the Canadian Archipelago from 10-16 August 2012. The lower heights interacted with the large-scale flow and drove each successive shortwave further southwards. The passing of each shortwave helped to reorient the baroclinic zone further south than before its passing. On 15 August 2012, an intense shortwave dug into the northern Plains from northwestern Canada. Thanks to the preceding shortwave passages moving the baroclinic zone southwards, this intense shortwave helped form a large trough across the central United States. This trough brought relief in the form of rain and cooler temperatures to areas that had been under an intense ridge all summer.
4.4 Snowcover and the AO

Snowcover is a very efficient driver of cold temperatures. Not only does it reflect almost all of the incoming shortwave radiation back to space, but also it cools the air above it. This cold air is dense, stubborn to move, and enhances the thermal gradient at the edge of the snowfield. The large mass of cold, dense air enhances the baroclinic zone and acts as a block to the large-scale flow. Development of this cold pool can act to keep the jet stream further to the south of this snowfield. Being on the poleward side of a jet stream during the cool season can produce reinforcing snowfall that acts as a positive feedback and expands the pool of cold air (Leathers et. al 2002, Cohen et. al 2007; Francis et. al 2012). The correlation between a vast, rapidly expanding snowpack and a negative AO has been discussed in Section 3.4.1. Using this correlation as a forecast tool could have great value in terms of long-term forecasting. If the polarity of the Arctic Oscillation could be predicted perhaps as early as the end of October, it could be used to make seasonal forecasts in the fall for the following winter season. A negative (positive) AO does not guarantee cold (warm) air for the entirety of the northern hemisphere. However, if the AO is negative (positive), a forecaster can increase confidence in seasonal forecasts for increased risk of cold (warm) air outbreaks. As has been stated earlier, a negative AO is indicative of anomalously high heights over the pole and anomalously low heights over the midlatitude regions. The opposite is true for a positive AO. Since lower geopotential heights are associated with cooler atmospheric temperatures, prediction of a warmer or colder winter can be made using the AO and expected geopotential height anomalies. Financial gains, as well as preparedness for what to expect the upcoming winter, could prove very useful to humanity. It also was shown that snowcover frequency has been increasing in Eurasia over the past few
decades (Figs. 3.4.5a, 3.4.5b, and 3.4.5c). Comparing the past 10 years of snowcover data to the 1980’s in Eurasia, vast areas of real estate are 30-50% more likely to experience snowcover for weeks 42-44. This drastic increase in snowcover frequency could be a direct consequence of melting sea-ice. As much more open water is exposed to the air and the associated latent heat flux from the sea surface, the increased atmospheric moisture is available to precipitate as snow across northern Asia after the autumnal equinox.

5. Conclusions

5.1 Overview

The objectives of this research were to: (1) Create a comprehensive list of the most intense Arctic cyclones using MSLP data; (2) Classify these cyclones in the manner of which they formed or propagated into the Arctic; (3) Determine the frequency of intense cyclones in the Arctic; (4) Attempt to create composites of cyclones of a similar genesis; (5) Conduct a case study of the early August 2012 intense Arctic cyclone; (6) Examine long-term climatic trends within the Arctic that could be associated with future cyclogenesis, as well as identify trends of the Arctic within our warming world; (7) Examine the relationship between October Eurasian snowcover and the Arctic Oscillation; and (8) Examine changes in snowcover extent and frequency.

5.1.1 Climatology

Cyclones were classified in terms of intensity (2, 1.5, and 1 sigma) and where they reached maximum intensities (Oceanic, North Atlantic, Eurasian, and North American sectors). No
significant trends were witnessed in regards to cyclone intensity or frequency. However, a slight increase in cyclone intensity was observed, albeit slight. Great seasonality regarding cyclone tracks was observed. Winter cyclones were heavily dominated by North Atlantic cyclones. Spring and fall cyclones were also primarily North Atlantic cyclones, but had a lot more spread regarding where they moved into the Arctic. Many more cyclones entered into the Arctic from Eurasian longitudes and North American longitudes. Summertime cyclones had the least degree of clustering in the track data, but were focused primarily over central and eastern Eurasia.

5.1.2 Composites

Geographical composites helped to identify planetary and synoptic-scale patterns that led to development of intense cyclones. For all three seasons, anomalously low geopotential heights over the high latitudes combined with anomalously high geopotential heights over the midlatitudes to produce a very strong PJS, and subsequent cyclogenesis. Of all the seasons, summertime cyclones were most difficult to find convincing antecedent conditions for cyclogenesis in the geographic-relative composites. To gain clarity regarding these summer cyclones, cyclone-relative composites were created. Cyclone-relative composites also helped to reduce noise in the geographic-relative composites due to the spread in where the cyclones occurred. A coastal group was classified based on cyclones that reached maximum intensity close to the Eurasian coastline. These cyclones had much greater low-level baroclinicity at 850-hPa and Q-vector forcing at 700 hPa. An oceanic group was classified based on the cyclones that reached maximum intensity over the Arctic Ocean far from the coastline. These cyclones had a more pronounced and intense jet stream at 300 hPa and lower heights at 500 hPa. Thus the conclusion was made that cyclogenesis of the coastal group of cyclones was aided by greater
low-level baroclinicity, and cyclogenesis of the oceanic group was aided by greater upper-level jet stream support.

5.1.3 Case Study

A case study was produced for the intense cyclone of Early August 2012. Antecedent conditions were favorable for development in agreement with the composites (positive AO values). (1) Southwesterly flow from southeastern Europe and India, combined with moisture contributions from landfalling typhoons in China, produced a very warm, moist airmass over northeastern Asia. (2) Anomalously low geopotential heights over the Arctic produced cooler than normal air over the Arctic. (3) The enhanced baroclinic zone produced a powerful PJS across northern Eurasia. (4) A 500-hPa shortwave dug southward towards central Russia. (5) Enhanced by a coupled jet stream and a very buoyant airmass, cyclogenesis began to take place on 04 August 2012. The incredible pressure gradient due to the cyclone produced powerful 10-m winds over 35 kts. These winds churned the Arctic Ocean, breaking up the ice, and aided in the record-setting ice loss.

5.1.4 Snowcover and the AO

Eurasian snowcover was shown to be a useful predictor for the AO the following winter. Correlation between the SGI and the AO was shown to be over $R=0.84$, a value deemed to be statistically significant. Thus, the SGI can be used as a forecast tool in the fall to aide in seasonal forecasting for the winter. An increase in Eurasian snowcover in October (weeks 42-44) would partially explain the downturn of winter AO values over the past decade, as SGI values would be
higher with this pronounced snowpack. This increase in snowcover, while difficult to explain, appears to be partially influenced by a significant increase in precipitable water over the Arctic.

5.2 Suggestions for Future Work

This research produced a distinctive climatology of cyclones using MSLP data from 1979-2013. While only the August 2012 cyclone was studied in detail, more work needs to be performed to examine specific impacts from these other intense cyclones, especially the impact on Arctic sea ice. The summertime baroclinic zone along the coast of Siberia needs to be studied more closely, especially how it might increase or decrease related to climate change. The seasons of fall, winter, and spring all had easily distinguishable patterns leading to development; the summer did not. An entire study based solely on summertime cyclones in the Arctic could prove important, especially as more of mankind enters the Arctic for shipping, energy, and tourist purposes. A better understanding of fall snowcover and winter AO would be helpful to improve the forecast skill of the SGI. Looking at other variables in conjunction with snowcover could be helpful in improving the forecasting skill of the SGI. More work relating the increase in October snowcover with climate change needs to be performed. As more ice melts, even more moisture is available to produce precipitation. The exact feedbacks between these physical processes have to be investigated. The Arctic is a fragile place, and deserves a better focus on the impacts brought forth by climate change.
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