An analysis of the linkages between large-scale flow regime transitions on the spatiotemporal distribution of clustered extratropical cyclone events over the Northern Hemisphere

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An Analysis of the Linkages Between Large-Scale Flow Regime Transitions on the Spatiotemporal Distribution of Clustered Extratropical Cyclone Events over the Northern Hemisphere

by

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ABSTRACT

Northern Hemisphere (NH) cool season atmospheric predictability for sub-seasonal time scales (2–4 weeks) is greatly affected by the structure and evolution of the North Pacific jet stream (NPJ). The susceptibility of the NPJ to external perturbations is a function of the phase and amplitude of ENSO on interannual time scales, the MJO on intraseasonal time scales, and the frequency of transient tropical, midlatitude, and polar disturbances that interact with the NPJ on synoptic time scales. Perturbations to the NPJ can result in the formation of eastward-propagating Rossby wave trains (RWTs) and subsequent downstream wave amplification across North America. Downstream RWT propagation arising from perturbations to the NPJ may enhance the probability of occurrence of extreme weather events (EWE) over North America and Eurasia. The purpose of this M.S. thesis is to construct a climatological and composite analysis of clustered EWEs associated with sequential extratropical cyclone (EC) events over the NH. This thesis is motivated both by the predictability challenges of forecasting clustered EC events and the significant potential for societal disruption from clustered EWEs.

The results from a NH seasonal EC clustering climatology constructed for the years 1979–2014 will be illustrated. Clustered EC tracks were derived from the Hodges (1994) cyclone-tracking algorithm and another algorithm produced by the ECMWF (e.g., Wernli and Schwierz 2006; Sprenger et al. 2017) on the ERA-I reanalysis datasets (Dee et al. 2011). The Hodges (1994) cyclone-tracking algorithm was utilized for the study. Extratropical cyclone clusters (ECCs) were classified by identifying two or more ECs that transect within 500 km of any gridpoint in the ERA-I reanalysis dataset during a
seven-day period. An assessment of the sensitivity of the seasonal distribution of clustered EC events over the NH as a function of the tracking methodology was constructed. A NH climatology of clustered ECs shows that they occur preferentially over the Gulf of Alaska, central North Pacific, north-central Russia, the Mongolian Plateau, and from the Canadian Maritimes to northwestern Europe. Composite analyses of ECCs in the aforementioned regions reveals that ECC events occur preferentially during quasi-stationary wavenumber three and four wave patterns, and where the NH flow pattern is in a quasi-stationary highly amplified flow pattern. ECC events in the Gulf of Alaska typically form in the diffluent flow pattern of the left exit regions of an extended NPJ in conjunction with cyclonic wave breaking. ECC events in the central and western North Pacific typically form in diffluent flow in the right entrance regions of a retracted NPJ that develops on the upstream side of a Rex block. ECC events in the North Atlantic occur preferentially in the divergent right entrance and left exit regions of the North Atlantic jet. In particular, ECC events occur more frequently downstream of the left exit region of the North Atlantic jet in a broad diffluent flow pattern, which supports recurrent EC activity and ECC occurrence. The frequency and distribution of these ECC events associated with these large-scale flow patterns is contingent upon the orientation and position of the midlatitude jets over the Atlantic and Pacific, which in turn are sensitive to the external perturbations aforementioned above. Finally, case studies of two representative clustered EC events will be shown to illustrate the life cycles of these events and the dynamical linkages between clustered EC events and individual large-scale circulation regimes on subseasonal time scales.
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(knots), 925–850-hPa layer averaged cyclonic relative vorticity (black contours, every 0.5 x 10^{-4} s^{-1}), at 1200 UTC 13 March 2018. Plots created by Heather Archambault: (http://www.atmos.albany.edu/student/heathera/).

Fig. 4.10. (a) 250-hPa wind (m s^{-1}, color shading), precipitable water (mm, grey shading), SLP (solid contours every 4 hPa), and 1000–500 thickness (dashed contours every 6 dam), and (b) DT (1.5-PVU surface) potential temperature (K, shaded), wind barbs (knots), 925–850-hPa layer averaged cyclonic relative vorticity (black contours, every 0.5 x 10^{-4} s^{-1}), at 1200 UTC 20 March 2018. Plots created by Heather Archambault: (http://www.atmos.albany.edu/student/heathera/).

Fig. 4.11 ECC density distribution and frequency for the 2–12 February 1998 case study of the West Coast of the United States.

Fig. 4.12. (a) 300-hPa geopotential height anomalies (m) and (b) 300-hPa geopotential height mean field composites (m) for 2–12 February 1998.

Fig. 4.13. 1000–500-hPa thickness (dashed blue; every 6 dam, with the 540-dam contour boulded), 500-hPa absolute vorticity (shaded; beginning at 5 x 10^{-5} s^{-1} according to the color bar), and MSLP (solid black; every 5 hPa). The five-day period surrounding the onset of a section of the ECC event for the western United States, where (a) T−48 to T+0 (0000 UTC 2 February–0000 UTC 7 February 1998) and (b) T+24 h to T+72 h (0000 UTC 5 February–0000 7 February 1998).
1. Introduction

1.1 Motivation

Extratropical cyclone clusters (ECCs), also known as cyclone families, are characterized by multiple extratropical cyclones (ECs) that form in a narrowly defined window of space and time. ECCs often form when consecutive shortwave disturbances develop along a trailing cold front in the wake of a strong, occluding, parent low (Bjerknes and Solberg 1922; Mailier et al. 2006). ECCs can sequentially develop in the equatorward entrance regions and poleward exit regions of the midlatitude jet streams. ECCs can be responsible for increased severe weather conditions (e.g., damaging winds and heavy precipitation) (Mailier et al. 2006) over large areas of the Northern Hemisphere (NH) during certain large-scale flow patterns, which can have substantial socioeconomic impacts. NH atmospheric predictability on subseasonal time scales (one–four weeks) depends significantly on the structure, position, and evolution of the North Pacific jet stream (NPJ) waveguide (Torn and Hakim 2015; Winters et al. 2018). The NPJ waveguide can be perturbed by recurving and transitioning tropical cyclones, Arctic disturbances, and Madden–Julian Oscillation (MJO)-related convection. NPJ waveguide perturbations can result in the formation of downstream propagating Rossby wave trains, that may be associated with clustered cyclone events, some of which may lead to the occurrence of extreme weather event (EWE), (Torn and Hakim 2015) defined as high-impact weather events, over large geographical areas for extended periods of time. Some persistent large-scale flow regimes (i.e., teleconnection patterns), such as the negative phase of the North Atlantic Oscillation (NAO) and associated high-latitude blocking episodes, may be particularly conducive to ECC events (Rogers 1997; Gulev et al. 2001).
Likewise, the structure and evolution of the NPJ can vary on subseasonal, seasonal, and annual time scales. These variations may be directly related to, and modulated by, the large-scale teleconnection patterns on these time scales (Winters et al. 2018). This study will be investigating how the phases and amplitudes of NH teleconnection patterns can exhibit a strong relationship with the location and intensity of the NPJ, and consequently, where ECCs develop.

The opportunity to improve scientific understanding of the linkages between certain teleconnection patterns, ECC events, and EWEs is the motivation for this research. Although past studies have investigated the serial clustering of ECs in the North Atlantic (Mailier et al. 2006), they have not unequivocally explored ECs that cluster in a spatiotemporal distribution around the entire NH, nor have they analyzed those ECCs in a climatological sense to large-scale regime transitions. Therefore, a goal of this research is to better understand whether major teleconnection patterns such as the NAO and Pacific/North American pattern (PNA), can be related to the locations of preferred ECC regions. The NAO is defined as a north–south dipole of geopotential height and sea level pressure anomalies over the North Atlantic with ECs either forming near Iceland or the Azores depending on the NAOs phase and amplitude (Wallace and Gutzler 1981; Hurrell 1995; Archambault et al. 2010). The PNA is defined as having four geopotential height and sea level pressure anomalies of opposite sign that resemble a Rossby wave train, with ECs forming in either the Gulf of Alaska or north central Pacific (Wallace and Gutzler 1981; Barnston and Livezey 1987; Archambault et al. 2010). It is imperative to state that these teleconnection patterns do not cause ECC events, but may themselves be influenced by ECCs, further enhancing their evident
The ECCs have the potential to reinforce the teleconnection patterns by continually forming new cyclones in preferred regions in conjunction with multiple trailing upper-level disturbances.

1.2 Literature Review

1.2.1 Formation of cyclones and cyclone clusters

Large-scale flow patterns persist during time scales of roughly one–two weeks, and occur over 5,000–10,000-km length scales, with ECCs persisting over 1,000–5,000-km length scales, which may contribute to the development of EWEs over these regions. To understand how ECCs can contribute to the formation of EWEs, it is pertinent to recognize the structure and evolution of ECs. The Norwegian cyclone model (Fig. 1.1) developed by Bjerknes (1919) was one of the first to systematically introduce an idealized representation of an EC. Bjerknes’ groundbreaking schematic describes how ECs redistribute and transport air masses vertically (i.e., cold air sinking and warm air rising) and horizontally via fronts (e.g., warm and cold fronts) associated with propagating cyclones. Bjerknes and Soldberg (1922) elegantly describe their polar front theory of extratropical cyclogenesis and the evolution of an idealized EC (Fig. 1.2) as it progresses through its life cycle from a broad temperature gradient, to a mature cyclone via perturbing mechanisms (e.g., baroclinic instability and atmospheric vorticity), and, finally to an occluded cyclone. Bjerknes and Soldberg (1922) were the first to propose the idea of cyclone families (Fig. 1.3), which are synonymous with cyclone clusters in this paper. Bjerknes and Soldberg (1922) proposed that cyclones are continually forming off the trailing frontal boundary of a parent cyclone to form cyclone clusters. These trailing frontal boundaries, which already contain residual vorticity in the lower and
midlevels, can be reinforced when in conjunction by trailing upper level disturbances to form cyclone clusters.

1.2.2 Climatologies of cyclones

EC climatologies across the NH, especially over the North Pacific and North America (Table 1.1), have been extensively studied over the last two centuries (e.g., Bentley 2018). Until the advent of reanalysis datasets and cyclone tracking algorithms within the most recent two and a half decades (e.g., Hoskins and Hodges 2002; Hodges et al. 2011; Grise et al. 2013; Tilinina et al. 2013), EC climatologies were meticulously constructed from hand-drawn maps in which the ECs were, manually tracking and counting (e.g., Saucier 1949; Hurley 1954; Petterssen 1956; Klein 1957). Many initial EC climatologies (e.g., Reitan 1974; Sanders and Gyakum 1980; Zishka and Smith 1980; Whittaker and Horn 1981; Roebber 1984; Changnon et al. 1995; Eichler and Higgins 2006; Grise et al. 2013) focused on North America. Based on these climatologies, ECs form and track preferentially: 1) in the lee of mountain ranges, in conjunction with vortex tube stretching downstream of the barrier, and 2) along continental–ocean boundaries where semi-permanent baroclinic zones are created when cold continental air masses are juxtaposed next to relatively warmer ocean surfaces. These EC climatological studies established that ECs are more frequent in the NH winter season (October–March) as compared to the NH warm season (April–September) due to the increased presence of 1) stronger horizontal temperature gradients in winter, 2) baroclinic instability, and 3) stronger values of vorticity (Whittaker and Horn 1981; Changnon et al. 1995).

ECs that form along the east coasts of NH continents near warm ocean currents (e.g., the Kuroshio Current and Gulf Stream) have been found to exhibit rapidly
deepening characteristics as compared to their lee cyclone counterparts (Sanders and Gyakum 1980; Roebber 1984; Sanders 1988; Gyakum 1988; Wernli and Schwierz 2006). These rapidly deepening coastal ECs also tended to be associated with significant surface fluxes of heat and moisture. Furthermore, the aforementioned studies have shown that rapidly deepening ECs have favorable genesis regions and preferred pathways along warm ocean currents (e.g., The Gulf Stream and Kuroshio Current) that are found near the eastern coasts of the CONUS and Asia. Using European Reanalysis-Interim (ERA-Interim) reanalysis data from 1989–2009, Hodges et al. (2011) were able to create an EC climatology of both track density and genesis density for the entire NH (Fig. 1.4). Figure 1.4 nicely highlights the preferred genesis locations near the Kuroshio and Gulf Stream currents, as well as in the lee of prominent mountain ranges (e.g., The Rockies, Alps, and the Tibetan Plateau). Preferred EC paths are also highlighted across both the North Atlantic and Pacific, the Great Lakes region of the United States, North Central Russia, the Mediterranean, and Mongolia extending towards northern Japan. Although climatologies of ECs track locations and genesis locations have been created, a climatology of ECCs has yet to be conducted. The purpose of this study is to document and identify preferred ECC locations and establish the role that the large-scale flow patterns (and associated teleconnections) and the configurations of the principal NH jet streams play in enabling the occurrences of ECCs.

1.2.3 Cyclone tracks and the formation of a cyclone cluster

ECCs have previously been studied by Pinto et al. (2013), but only for a single case study over the British Isles. An example of an ECC from Pinto et al. (2013) is shown in Fig. 1.5 where 23 ECs of varying intensity intersect the 700-km radius around
(55°N, 5°W) during January 2007. Pinto et al. (2013) observed that ECCs occurred more frequently with an intensified quasi-stationary upper-level jet. This allowed for continuous strong vertical motion of air near the right-entrance and left-exit regions of the jet core and promoted the development of rapidly intensifying cyclones (e.g., Uccellini 1990; Rivière and Joly 2006a, 2006b). Pinto et al. (2013) and Priestly et al. (2017) determined that clustering events that occurred over northwestern Europe were associated with a zonally extended quasi-stationary North Atlantic jet that was influenced both by cyclonic Rossby wave breaking (RWB) on its northern flank and anticyclonic RWB on its southern flank (Fig. 1.6 and Fig. 1.7). Figure 1.8 exhibits a schematic revealing the positions of clustering cyclones with respect to the North Atlantic jet stream and the RWB activity. At issue is whether this large scale RWB pattern that lead to the quasi-stationary North Atlantic jet could also be applied to quasi-stationary North Pacific jet patterns that also lead to ECC events.

1.2.4 Impacts of large-scale regimes

Large-scale teleconnection patterns such as the NAO have both a positive and negative phase that are associated with different synoptic-scale patterns. The positive phase of the NAO is characterized by an anomalous low-over-high geopotential height pattern, with above-normal heights and pressure over the central North Atlantic, eastern United States, and western Europe (Benedict et al. 2004). These height anomaly patterns would result in a poleward shift of the North Atlantic jet and a poleward shift in the storm track. Above average temperatures in the eastern United States and across northern Europe would result whereas below average temperatures would be likely over Greenland in conjunction with anomalously low upper-level geopotential heights (e.g.,
Walker and Bliss 1932; van Loon and Rogers 1978; Rogers and van Loon 1979; Hurrell 1995). The negative phase of the NAO is characterized by a high-over-low geopotential height anomaly pattern, opposite to the height and pressure anomalies over the aforementioned regions during the positive phase of the NAO (Benedict et al. 2004). During the negative phase of the NAO the North Atlantic jet and the associated storm track is shifted equatorward.

Although the NAO has a large influence on the synoptic flow patterns in the North Atlantic (Lau 1988; Rogers 1990; Archambault et al. 2010), the PNA, which similarly has a positive and negative phase, has a dominant influence on the synoptic flow patterns of the North Pacific and the location of the strength of the NPJ (e.g., Wallace and Gutzler 1981; Barnston and Livesey 1987; Franzke and Feldstein 2005; Strong and Davis 2008; Athanasiadis et al. 2010; Franzke et al. 2011; Griffin and Martin 2017; Winters et al. 2017). Unlike the NAO, which measure the strength of the polar vortex, the PNA is associated with wave trains that originate in the central tropical Pacific that first propagate poleward and eastward toward western North America and then propagate equatorward and eastward toward southeastern North America and the Caribbean. The positive phase of the PNA is characterized by above-normal heights and pressure over Hawaii and the Intermountain West region of North America, and below-normal heights and pressures over the Aleutian Islands and the southeast United States. This phase of the PNA is associated with an extended NPJ and enhanced cyclone activity in the Gulf of Alaska. Below-normal heights and pressures over the southeastern United States are associated with enhanced cyclone activity across this region extending out of the Gulf of Mexico. The negative phase of the PNA is associated with a retracted jet and
an increase of cyclone activity over the central North Pacific in conjunction with the below-normal heights and pressures over eastern Asia and often a blocking pattern across the central North Pacific.

The Climate Prediction Center’s (CPC) Oceanic Niño Index (ONI) is a measure of the phase and amplitude of the El Niño Southern Oscillation (ENSO). The ONI is an index of the three-month running mean of ERSST.v5 sea surface temperature (SST) anomalies in the Niño 3.4 region (5°N–5°S, 120°–170°W). A positive ONI coincides with EL Nino conditions whereas a negative ONI is associated with La Nina conditions. Although the ONI index is not a measure of a synoptic-scale driven teleconnection pattern, the SST anomalies that influence the phase of this index can impact the synoptic evolution and variability of the midlatitude flow (e.g., Hoerling and Kumar 1997; Hoerling and Kumar 2000; Higgins et al. 2002; Eichler and Higgins 2006). The positive phase of ONI (i.e., El Niño) is characterized by an equatorward shift in the storm track in the North Pacific and an enhanced, and equatorward-shifted, United States East Coast storm track. The negative phase of ONI (i.e., La Niña) is characterized by a poleward shift in the storm track (comparative to El Niño) in the North Pacific and United States East Coast storm tracks. Since the different phases of ENSO affect the NH storm track, the different phases of ENSO may also affect the formation and frequency of ECCs.

1.3 Research goals and thesis structure

ECCs may play an important role in influencing certain large-scale synoptic regimes such as the NAO and other teleconnection patterns, and in the development of EWEs. The research in this thesis will expand upon previous work on spatiotemporal ECCs. This research will also attempt to link the occurrence of ECCs to large-scale
atmospheric teleconnection patterns to help determine whether a correlation may exist between ECCs and teleconnection indices and, if so, how and why. Therefore, the goals of this research are to improve the understanding of: 1) the frequency of ECC events and their geographical distribution; 2) the environments associated with the cyclones, both upstream and downstream of the clustered cyclone events to determine what enables these ECC events to occur, and, 3) the extent cyclone clustering is influenced by atmospheric blocking and the phases, and amplitudes, of the major teleconnection indices (e.g., NAO and PNA), and ENSO.

The thesis is organized as follows. Data and methodology are described in chapter 2. NH climatologies of ECCs, ENSO, and other teleconnection indices, and ECCs that are linked to chosen teleconnection indices are discussed in chapter 3. Case studies of ECCs are discussed in chapter 4. Research results, conclusions, and suggestions for future work are discussed in chapter 5.
a. Tables

TABLE 1.1. Examples of climatologies that document ECs over the Northern Hemisphere, North America, the United States, and regions comprising central and eastern North America (i.e., Rocky Mountains, central United States, and east coast of North America). Climatologies are separated into regions according to the genesis location of ECs included in the climatology. The period of study and applicable reference are given for each climatology. An asterisk next to a reference indicates that the climatology exclusively identifies rapidly deepening ECs. This table is adapted from table 1 of Bentley et al. (2018).

<table>
<thead>
<tr>
<th>Region</th>
<th>Period of study</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Northern Hemisphere</td>
<td>1899–1936</td>
<td>Petterssen (1956)</td>
</tr>
<tr>
<td></td>
<td>1899–1936</td>
<td>Klein (1957)</td>
</tr>
<tr>
<td></td>
<td>1979–2009</td>
<td>Hodges et al. (2011)</td>
</tr>
<tr>
<td></td>
<td>1979–2010</td>
<td>Tilinina et al. (2013)</td>
</tr>
<tr>
<td></td>
<td>1976–1979</td>
<td>Sanders and Gyakum (1980)*</td>
</tr>
<tr>
<td>United States</td>
<td>1892–1912</td>
<td>Bowie and Weightman (1914)</td>
</tr>
<tr>
<td></td>
<td>1905–1954</td>
<td>Hosler and Gamage (1956)</td>
</tr>
<tr>
<td>Rocky Mountains</td>
<td>1958</td>
<td>Chung et al. (1976)</td>
</tr>
<tr>
<td>Central United States</td>
<td>1899–1939</td>
<td>Saucier (1949)</td>
</tr>
<tr>
<td></td>
<td>1920–1929</td>
<td>Hurley (1954)</td>
</tr>
<tr>
<td>East coast of North America</td>
<td>1929–1939</td>
<td>Miller (1946)</td>
</tr>
<tr>
<td></td>
<td>1921–1962</td>
<td>Mather et al. (1964)</td>
</tr>
<tr>
<td></td>
<td>1979–2004</td>
<td>Colle et al. (2013)</td>
</tr>
</tbody>
</table>
Fig. 1.1. Schematic representation of an idealized EC. The blue and red lines indicate the positions of the cold and warm fronts, respectively. The thin black and white arrows indicate the movement of cold and warm air around the EC, respectively. The thicker black arrow represents the motion of the EC. The hatched areas represent the areas of precipitation around the EC. This figure is adapted from Fig. 1 of Bjerknes and Solberg (1919).
Fig. 1.2. Schematic representation of the structure of an idealized EC through its life cycle, where panels (a)–(h) increment in time of the ECs life. The dashed black lines represent the boundaries between cold and warm air masses. The black and white arrows represent the flow of cold and warm air around the EC, respectively. Gray shaded regions represent the regions where it is precipitating around the EC. This figure is adapted from Fig. 2 of Bjerknes and Soldberg (1922).
Fig. 1.3. Schematic representation of cyclone families, where a secondary cyclone is formed on the trailing wavy cold front of a primary parent cyclone. This figure is adapted from Fig. 5 of Bjerknes and Soldberg (1922).

Fig. 1.4. Extratropical cyclone climatology based on the ERA-Interim reanalysis (1989–2009) using vorticity at 850 hPa. (a) NH DJF track density; (b) NH DJF genesis density. Densities are in units of number density per month per unit area, where the unit area is equivalent to a 5° spherical cap (~10^6 km^2). This figure is adapted from Fig. 1 of Hodges et al. (2011).
Fig. 1.5. January 2007 cyclone tracks affecting the British Isles. Red lines are the 98\textsuperscript{th} percentile most intense cyclones, blue lines are the 95\textsuperscript{th} percentile most intense cyclones, and black lines are all other cyclones. The circle around the British Isles has a radius of 700 km. This figure is adapted from Fig. 1 of Pinto et al. (2013).
Fig. 1.6. Composite images of North Atlantic RWB and jet anomalies on days lagged to clustered days at 55°N. The red shading is the percentage increase in the $B$ index compared to climatology. The $B$ index is a field that identifies where the meridional gradient of $\theta$ is reversed with positive values indicating the presence of RWB. Black contours are the 250 hPa wind speed anomalies ($\text{m s}^{-1}$). Shown is from lag $-8$ days to lag $+8$ days at 2-day intervals. This figure is adapted from Fig. 3 of Priestly et al. (2017).
Fig. 1.7. 6–20 January 2007. Red/blue shadings: θ on the 2 PVU surface in K (00 UTC). Hatched fields: daily RWB occurrence. Dashed contours: wind intensity at 250 hPa (m s\(^{-1}\), 00 UTC), contours drawn from 40 m s\(^{-1}\) with 10 m s\(^{-1}\) contour intervals. Solid contour lines: Full p95 cyclone trajectories until 18 UTC of each day. Large filled black dots: Cyclone positions at 00 UTC. Small circles: three forthcoming cyclone positions on the same date. Large open white dots: Positions (00 UTC) of named historical storms crossing the detection area ((55°N, 5°W), r = 700 km) on that day. This figure is adapted from Fig. 3 of Pinto et al. (2013).
Fig. 1.8. (a, c, and e) RWB occurrence (B > 0; hatched), wind intensity at 250 hPa (m s$^{-1}$; dashed contours drawn from 40 m s$^{-1}$ with 10 m s$^{-1}$ contour intervals), cyclone surface centers and fronts (from UK Met Office weather charts; red/blue/purple solid lines for warm/cold/occluded fronts) for 00 UTC on example dates 11, 13, and 19 January 2007. (b, d) Weather charts (00 UTC) on 11 and 13 January 2007. (f) Schematic summary showing relative positions of clustering cyclones with respect to jet streak location and location of RWB. This figure is adapted from Fig. 7 of Pinto et al. (2013).
2. Data and Methodology

2.1 Cyclone tracks and cyclone tracking

ECs and their tracks were identified using Hodges’ (1994, 1995) tracking algorithm, which utilizes a feature-tracking algorithm developed by Sethi and Jain (1987) and Salari and Sethi (1990). This tracking algorithm utilizes feature point identification from images, which Hodges (1994, 1995) applied to pressure and relative vorticity fields to find local minima and maxima, respectively. For the purpose of this study, only pressure centroids of ECs were utilized for tracking purposes.

The cyclone-tracking algorithm developed by Hodges (1994, and 1995), which utilizes a feature point tracking identification scheme developed by Sethi and Jain (1987) and Salari and Sethi (1990), was used on the mean sea-level pressure (MSLP) and 850-hPa relative vorticity fields to find relative minima and maxima values, respectively. Pressure centroids at six-hour intervals were utilized to track cyclones. This cyclone tracking algorithm was applied to the ~0.7° horizontal resolution ERA-Interim (ERA-I) reanalysis dataset (Dee et al. 2011). All cyclones identified by the tracking algorithm in the NH between the years 1979–2014 were considered.

2.2 Cyclone Clustering Algorithm

Pinto et al. (2013) analyzed one ECC event during a one-month period, but for this study to develop a better representation of cyclone clusters, a seven-day overlapping window of time that progress one day at a time starting on 1 January every year and ending 31 December was chosen as the temporal period for cyclone clustering. This
temporal period was subjectively chosen to get a better idea of individual cyclone cluster events and to effectively calculate a more efficient average of ECC events in given regions. Longer time periods were not investigated, because the purpose of this study was to identify single ECC events of two–three cyclones. To effectively determine the distribution of NH ECC events, a 0.7° X 0.7° grid was created using the ERA-I native horizontal resolution. A 500-km radius was created around each grid point, which was subjectively chosen based on the general representation, intensity, and size of an average EC found in Fig. 2.1 (Rudeva and Gulev 2007). This radius value is slightly smaller than the 700-km radius chosen by Pinto et al. (2013), because in this study we investigate ECCs across all meteorological seasons, and not just the winter season when ECs are characteristically deeper, more expansive, and move faster. In order for an ECC to be considered two or more cyclones had to have at least three six-hour time steps within the 500-km radius of any given grid-point (Fig. 2.2). To create an effective representation of ECCs, these grid-points were color contoured based on the number of cyclones that occurred within the 500-km radius. Finally, the seven-day centered running mean cyclone cluster values were summed for the three-month meteorological seasons (i.e., December, January, February (DJF), March, April May (MAM), June, July, August (JJA), and September, October, November (SON)), as well as by year.

2.3 Teleconnection indices and ONI

The monthly ECC averages previously discussed were utilized to create seasonal averages of teleconnection patterns such as the NAO and the PNA along with the CPC’s ONI. Each teleconnection patterns values were averaged and filtered by meteorological season from 1979–2014. These averages of the teleconnection indices phase and
amplitude were done by the meteorological seasons to remove the need for considering lag days of each teleconnection pattern. Furthermore, by using averaged seasonal values of ECCs, we will be able to create a better representation of the correlations between these large-scale teleconnection patterns and preferential ECC corridors. All of the meteorological seasons were then identified and selected based off of whether the averaged season had a value of >=1 and <=−1 to indicate whether the season had a positive or negative value of those indices respectively.

2.4 Teleconnection Cyclone Density Cluster Averages

The ECC density values that corresponded with the seasons that met the aforementioned criteria of >=1 and <=−1 for each teleconnection pattern were averaged together to create seasonal ECC patterns for each large-scale teleconnection pattern. This averaging of ECC values during periods of the selected teleconnection patterns will help establish preferred and preferential ECC locations. Knowing the preferential synoptic pattern (e.g., jet configuration) that corresponds with each large-scale regime associated with these teleconnection patterns and ECC events will help to determine whether there is a linkage and correlation between these large-scale regimes and ECCs. The mean and anomaly 300-hPa geopotential height from the 2.5° NCEP–NCAR reanalysis dataset (Kalnay et al. 1996) was utilized for the same temporal periods as the aforementioned averages of the teleconnection patterns selected to derive the synoptic flow patterns and jet locations. These averages of ECCs for both the positive and negative values for each teleconnection pattern by meteorological season were finally subtracted (i.e., positive−negative) to determine whether differences in ECC activity were evident between the positive and negative values of teleconnection patterns.
2.5 Composite analysis of cluster events

To better stratify whether clustering is associated with regime onset or demise, composites of mean and anomaly (e.g., precipitable water, 300 hPa geopotential heights, 500 hPa geopotential heights, 850 hPa temperature, and mean sea level pressure) were created using the 2.5° NCEP–NCAR reanalysis dataset (Kalnay et al. 1996). The weekly ECC events for all the meteorological seasons that were previously identified at the (>=1), and (<=-1) thresholds for specific teleconnection patterns were examined, and key regions where ECC events occurred were identified. The weekly ECC events were then selected during these large-scale patterns where the threshold of the algorithm for ECC was greater than or equal to three ECs for that region. The composites of the aforementioned variables were then created three days prior to the onset of the ECC event, during the ECC event, and 12 days after the ECC event.

2.6 Case Studies

Representative case studies of ECC events were chosen based on several key characteristics (e.g., societal impact, event duration, correlation/linkage between certain teleconnection patterns, and downstream influence/impacts). Two main case studies were chosen from the Atlantic and Pacific basins that met these specific characteristics. For the Atlantic basin, the March 2018 East Coast ECC event was chosen due to the societal impacts, including significant snowfall and rainfall totals, extreme wind events, and coastal flooding across the Mid-Atlantic and Northeast regions. Although this case falls outside our cyclone track dataset cyclone tracks were manually created from the MSLP field in the ~0.70° horizontal resolution ERA–I reanalysis dataset, and then clustered using the aforementioned algorithm. For the Pacific basin, the February 1998
West Coast ECC event was chosen due to the significant societal impacts, including heavy rainfall totals leading to mudslides, extreme wind events, and coastal flooding. This case also had a strong linkage to the positive phase of ENSO.
b. Figures

Fig. 2.1. (a) Winter and (b) summer climatological distribution of the cyclone effective radius (km). This figure is adapted from Fig. 3 of Reduva and Gulev (2007).
Fig. 2.2. Cyclone tracks from 14–22 January 2010. Black Xs represent a cyclone’s starting position, blue dots represent any 0000 UTC time-step along a cyclone’s track, and red dots represent every other six-hour time-step. Green dots represent track points along a cyclone’s track that fall within the 500-km radius of the purple dot that represents the center grid point of the 500-km green radius.
3. Climatologies

3.1 Climatology of clustered cyclones

The purpose of this chapter is to present a climatological analysis of ECCs across the NH. This NH ECC climatology will be illustrated seasonally and annually. The ECC distribution for selected years will also be shown to illustrate year-to-year variability in the location and frequency of ECC maxima. The NH ECC climatology will be stratified as a function of the CPC Oceanic Nino Index (ONI), the North Atlantic Oscillation (NAO), and the Pacific–North American (PNA) pattern teleconnection indices.

a) Seasonal ECC distribution

(i) DJF

Averaged ECC density plots using the ECC algorithm previously discussed were created for all meteorological seasons (Fig. 3.1). During DJF, ECCs were prevalent across the North Pacific and the Gulf of Alaska, the northwest Atlantic near Newfoundland, east of Greenland, and north of the British Isles toward Scandinavia (Fig. 3.1a). Other ECC maxima occur across northeast Russia in the lee of the Ural Mountains over Siberia north of Lake Baikal, south of the Hudson Bay, and close to the North Pole near the 150° meridian. To better understand the large-scale flow pattern, the 2.5° NCEP–NCAR reanalysis dataset (Kalnay et al. 1996) was used to construct composite maps of seasonal 300-hPa geopotential height to facilitate comparison of the observed NH ECC distribution with the larger-scale flow pattern (Fig. 3.2a). A quasi wavenumber-three pattern is observed in the DJF 300-hPa composite map with a trough over eastern North America, a broad trough over the Kamchatka Peninsula, and a trough over eastern Europe (Fig. 3.2a). A diffluent height pattern is observed over the north-central Pacific
extending into the Gulf of Alaska, indicative of an extended NPJ that would allow ECs to
develop in the left-exit region of the NPJ, consistent with the high density of ECC events
in this region. ECC occurrences also maximize in the difffluent flow region downstream
of an eastern North American trough over the western North Atlantic and downstream of
the broad trough located over eastern Europe and western Russia.

(ii) MAM

In MAM, the North Pacific exhibits a broader region of preferred ECC occurrence
as compared to DJF with maxima extending from the coast of Japan across the North
Pacific to the Gulf of Alaska. The North Atlantic maximum observed in DJF shifts from
the northwest Atlantic to off the coast of Newfoundland in MAM. Separate, smaller ECC
maxima exist off the east coast of Greenland and north of the British Isles. ECC maxima
are also evident in the lee of the Rockies, Himalayas, and Urals. At 300-hPa, a broad
trough extends from eastern Asia eastward to the Gulf of Alaska (Fig. 3.2b). A quasi
wavenumber-four pattern is observed in the MAM 300-hPa composite map with troughs
over eastern North America, Japan, central Europe, and the Gulf of Alaska (Fig. 3.2b).
Other 300-hPa troughs are evident over eastern North America, central Europe, and over
the western North Pacific (Fig. 3.2b). The 300-hPa height pattern is slightly difffluent
over the Gulf of Alaska where a large ECC maximum is observed where a left-exit region
of the NPJ would be expected. Similarly, the ECC maxima in the northwest Atlantic and
over north-central Russia occur in difffluent flow regions downstream of 300-hPa troughs
(Fig. 3.2b).

(iii) JJA
During JJA, the North Atlantic exhibits the highest frequency of ECCs in a broad region extending from North America towards northwestern Europe (Fig. 3.1c). A smaller maximum south of the Hudson Bay extends eastward from the lee of the Canadian Rockies. The North Pacific also has an ECC maximum extending from Japan into the Gulf of Alaska with a local maximum in the north central Pacific (Fig. 3.1c). The JJA 300-hPa mean geopotential height field shows a diffluent trough in the North Atlantic, a larger-scale trough in the North Pacific, and another trough across central Russia (Fig. 3.2c). A quasi wavenumber-five pattern is observed in the JJA 300-hPa composite map with troughs over eastern North America, central Europe, Siberia, eastern Asia, and the Gulf of Alaska (Fig. 3.2c). The broader trough and the associated 300-hPa diffluent height field over the eastern North Pacific supports the frequency and density of ECC events across the North Pacific.

(iv) SON

Although the SON and MAM seasonally averaged ECC density maps have many similarities, there are differences in location and frequency of ECC events (Fig. 3.1d). The North Atlantic maximum is stronger and shifted farther eastward over Iceland and towards the Scandinavian Peninsula. Likewise, the North Pacific ECC maximum is shifted poleward and westward, extending from northeast of Japan to the Gulf of Alaska. Another ECC maximum is situated in the lee of the Ural Mountains and extends east towards central Russia and north of Lake Baikal (Fig. 3.1d). The largest spatial distribution of ECC density maximum is located in the lee of the Himalayas over
Mongolia and south of Lake Baikal. At 300-hPa, a quasi wavenumber-four pattern is observed with troughs over eastern North America, central North Pacific, and over the Korean Peninsula (Fig. 3.2d). The north-central Pacific ECC maximum lies downstream of a weakly diffluent 300-hPa trough and supports ECC occurrences across the North Pacific basin. The ECC maximum over north-central Russia is also associated with a very broad, long wavelength trough over eastern Russia. Although the ECC maximum over Mongolia does not appear to be associated with any climatological trough, its location in the lee of the Himalayas suggests that it can be attributed to lee cyclogenesis (Figs. 3.1d and 3.2d).

b) Climatological ECC distribution

A 35-year average ECC density climatology was computed for 1979–2014 (Fig. 3.3). ECC maxima are evident along the North Atlantic and North Pacific storm tracks (Fig. 3.3). ECC maxima are also evident just south of the Hudson Bay, in central Russia in the lee of the Urals, and north, and east, of the Himalayas in eastern Mongolia (Fig. 3.3a). The 300-hPa geopotential height field composite for 1979–2014 (Fig. 3.3b) exhibits a quasi-wavenumber-three pattern with climatological trough axes over eastern North America and Japan, and with much broader troughs across eastern Russia and the North Pacific into the Gulf of Alaska. The eastern North American and eastern Asian troughs support the observed ECC maxima downstream of these troughs while the eastern Russia trough supports the ECC maxima over Siberia and the Arctic. The ECC maxima over Mongolia and south of the Hudson Bay are likely associated with lee
cyclogenesis events.

The above ECC climatology compares favorably with previous EC climatologies discussed in chapter 1 (i.e., Hodges 2011; Roebber 1984; Sanders and Gyakum 1980; Wernli and Schierz 2006). Both the North Atlantic and North Pacific EC maxima are evident in all of the datasets. The north-central Russia ECC maximum is slightly less evident among the earlier EC climatologies except for Hodges (2011), which clearly show this maximum. These climatological differences may arise from dataset and resolution differences between the various studies, as well as differences in the methods used to track the ECs.

3.2 ECC maxima interannual variability

An analysis of ECC maxima variability for 1984, 2000, and 2010 is next presented. These years were chosen because they either exhibited significant anomalous behavior in terms of ECC frequency of occurrence or location, or they had a significant societal impact.

a) 1984

The year 1984 had an anomalously high frequency of ECC events over the Mongolian Plateau (Fig. 3.4). The shoulder seasons, MAM and SON, were very active and had the highest ECC frequency (Fig. 3.5b and Fig. 3.5d). The anomalous 300-hPa height field exhibits anomalously low heights over the Mongolian Plateau (Fig. 3.6a). A short-wave trough just east of the Mongolian Plateau in the mean 300-hPa height is also observed which would help explain the anomalous ECC frequency over this region (Fig.
3.6b).

b) 2000

The year 2000 had an anomalous occurrence of high ECC frequency over the north-central Pacific (Fig. 3.7). This year exhibited a strong negative signal of the ONI, which supports an anomalously strong La Nina event for the year 2000. This pattern had the highest frequency of ECC events during DJF (Fig. 3.8). Anomalously low 300-hPa geopotential heights are observed over the north-central Pacific (Fig. 3.9a). A trough axis and associated diffluent pattern is observed downstream of the north-central Pacific in the mean 300-hPa height field, with many of the ECC events occurring in the base of the trough axis (Fig. 3.9b). This distribution supports the higher frequency of ECC events in this region, and also provides a pattern for regions that feature preferential clustering of cyclones during certain phases of ENSO (e.g., Hoerling and Kumar 1997; Hoerling and Kumar 2000; Higgins et al. 2002; Eichler and Higgins 2006).

c) 2010

The year 2010 was chosen due to its anomalous ECC activity over the Gulf of Alaska, off the coast of the Canadian Maritimes, west of the Iberian Peninsula, Eastern Europe, and north central Siberia (see Fig. 3.1). This year was characterized by a strong negative phase of the NAO with DJF having the largest negative value of the NAO in the dataset. This large negative NAO value supports the anomalous ECC activity compared to (Fig. 3.1) is observed off the coast of the eastern Canadian Maritimes and to the west
of the Iberian Peninsula (Fig. 3.10a and Fig. 3.10b). The climatologically anomalous ECC activity compared to (Fig. 3.1) in Eastern Europe can be attributed to the strong negative NAO pattern that was persistent throughout the year. Furthermore, climatologically above average ECC activity compared to (Fig. 3.1) was prevalent in Siberia during the summer season (Fig. 3.10c). A wavenumber-four pattern across the NH was observed in the 300-hPa mean height field while negative height anomalies observed across central Siberia in the 300-hPa anomalous height field (Figs. 3.11a and 3.11b). There is anomalous troughing in the Gulf of Alaska, over eastern Europe, off the coast of North America, and in central Siberia. These height anomalies directly link to the position of the North Atlantic jet, which is extended equatorward across the Atlantic extending from the mid-Atlantic region of the United States to the Iberian Peninsula due to the anomalously high heights over Greenland. In the mean 300-hPa heights, there is a diffluent pattern extending from North America to western Europe (Fig. 3.11b). This diffluent pattern supports the anomalous ECC activity observed across those regions. It is evident that the negative NAO pattern had a significant role in the large-scale synoptic pattern and ECC activity during 2010.

These few cases have provided evidence that ECCs have preferred corridors in the NH especially in the North Atlantic and North Pacific, with specific regions receiving more ECC events during certain large-scale flow regimes (e.g., during the negative phase of NAO, clustered cyclones preferentially occur across southern parts of the North Atlantic).
3.3 Climatology of ECC events during the positive and negative phases of the ONI

3.3.1 Climatology of ECC events during the positive phase of the ONI

A climatology of ECC location and frequency during seasons of \(>=1\) ONI is presented. Dynamic and thermodynamic variables will be composited to determine preferred large-scale flow patterns that support the exhibited ECC frequency and location.

Figure 3.12 illustrates the seasonal variability of the phase and amplitude of ENSO from 1979–2014. Certain years stand out as representative strong El Niño years (e.g., 1983, 1992, 1998, and 2010). Storm tracks during this phase of ENSO can be characterized by an equatorward shift in the storm track in the North Pacific and an enhanced, and equatorward-shifted, North American East Coast storm track (e.g., Hoerling and Kumar 1997; Hoerling and Kumar 2000; Higgins et al. 2002; Eichler and Higgins 2006). There are several years that stand out as representative strong La Niña years (i.e., 1988, 1999, 2000, 2008, and 2011). Storm tracks during this phase of ENSO can be characterized by a poleward shift in both the North Pacific and North American East Coast storm tracks.

a) DJF

Figure 3.13a reveals a composite of DJF ECC distribution for 1983, 1987, 1992, 1995, 1998, and 2010, which all observed a value of ONI \(>=1\). ECC events occur preferentially in the Gulf of Alaska, north central Russia, and along a more southern
storm track across the North Atlantic compared to (Fig. 3.1a). Figure 3.14a reveals that there are anomalously low 300-hPa heights across the eastern North Pacific into the Gulf of Alaska, extending across the southern United States, and into the North Atlantic. There are also positive height anomalies over the Hudson Bay. This patterns highlights both the midlatitude and subtropical jets across the interior United States, the North Pacific, and central North Atlantic, respectively. The mean 300-hPa geopotential height field exhibits deep troughs over eastern North America and over eastern Russia, with a diffluent pattern over the Gulf of Alaska and the central North Atlantic (Fig. 3.15a). This large-scale pattern supports the distribution and frequency of DJF ECC events during El Niño years.

b) MAM

Figure 3.13b reveals a composite of MAM ECC distribution for 1983, 1992, and 1998, which all observed a value of ONI >= 1. During MAM, the overall ECC distribution is very similar to the pattern observed during DJF; however, the ECC frequency observed is different in many respects (Fig. 3.13b). ECCs preferentially occur in an equatorward-shifted storm track compared to (Fig. 3.1b) across the Pacific and Atlantic basins (Fig. 3.13b). The North Pacific now has two tracks across the Pacific (e.g., the central Pacific and the Gulf of Alaska). A maximum in ECC frequency is also observed over north-central Russia albeit shifted slightly eastward. At 300-hPa, negative height anomalies extend from the central North Pacific through the United States and into the North Atlantic (Fig. 3.14b). The positive height anomalies observed over the Hudson
Bay during DJF have shifted east towards the Canadian Maritimes. There are negative height anomalies over north-central Russia where a local ECC frequency maximum occurred. A similar result is observed in the 300-hPa mean height field for MAM as compared to DJF (Fig. 3.15b). A sharp trough is observed over eastern North America with a diffluent pattern over both the Gulf of Alaska and the central North Atlantic, which support the distribution of ECC frequency in the aforementioned regions.

c) JJA

Figure 3.13c reveals a composite of JJA ECC distribution for 1987 and 1997, which both observed a value of ONI >= 1. Although the observed ECC distribution and frequency shift slightly poleward during JJA, there are still maxima in ECC frequency in the Gulf of Alaska, north-central Russia, and the north-central Atlantic (Fig. 3.13c). The locations of the 300-hPa height anomalies also shift. Anomalously negative heights across southern portions of the United States no longer exist; however, the anomalous negative heights across the North Pacific are still present (Fig. 3.14c). Anomalous negative heights in north-central Russia are more apparent during JJA compared to DJF and MAM. A similar pattern is observed in the mean 300-hPa height field as a wavenumber-three pattern is apparent, with troughs over eastern North America, eastern Russia, and over the Korean Peninsula (Fig. 3.15c). The locations of these troughs, and the diffluent patterns associated with the troughs supports the distribution of ECC events observed.

d) SON:
Figure 3.13d reveals a composite of SON ECC distribution for 1982, 1987, 1997, 2002, and 2009, which all observed a value of ONI >= 1. During SON, the ECC maximum is located in the eastern North Atlantic (Fig. 3.13d). Figure 3.13d exhibits a maximum of ECC frequency in north-central Russia, as well as an equatorward shift in both the North Atlantic and Pacific compared to (Fig. 3.1d). Negative 300-hPa height anomalies are observed over the Gulf of Alaska, as well as extending into southern regions of the United States; however, the negative height anomalies over the North Atlantic have been shifted farther poleward towards the southern tip of Greenland compared to the previous seasonal composites (Fig. 3.14d). Negative height anomalies are also observed in west-central Russia, which is consistent with the frequency of ECC events. The mean 300-hPa heights exhibit a trough over eastern North America, with an associated diffluent flow pattern extending across the North Atlantic into Western Europe (Fig. 3.15d). Two broad troughs are also observed in eastern and western Russia (Fig. 3.15d). A trough is also observed over the Aleutian Islands in the North Pacific, with a diffluent flow pattern extending from the trough axis into the Gulf of Alaska where a higher frequency of ECC events occurred. These troughs, along with the associated diffluent flow patterns, support the distribution of the ECC events observed in the mean.

3.3.2 Positive ONI case composites

Individual cases of ECC events were composited during the DJF seasons with values of ONI that were >=1. To get a better understanding of the dynamic and thermodynamic processes that are associated with the ONI, the following mean and
anomaly values of precipitable water, 300-hPa geopotential heights, and 850-hPa temperature variables were composited for three days before, during, and 12 days after each ECC event. 12 days after the event was chosen to be able show that the large-scale flow pattern changes over the 1–2 week time frame. The composites for positive ONI were composed of 14 separate ECC events (i.e., 14 February 1983, 14 January 1983, 12 December 1982, 31 January 1987, 27 December 1986, 13 January 1992, 23 February 1992, 31 January 1995, 17 February 1995, 19 December 1994, 16 January 1998, 13 February 1998, 23 December 1997, and 14 February 2010) that occurred within a similar region.

a) Three days before the events

The Gulf of Alaska where ECCs preferentially occurred, was chosen for the positive ONI ECC region (see Fig. 3.13a). Anomalous 300-hPa heights are observed west of the Gulf of Alaska, suggestive of an extended NPJ, as can be seen in the mean 300-hPa geopotential height pattern (Fig. 3.16a and Fig. 3.16b). The NPJ (not shown) extends across the Pacific, but progressively recurves once it reaches north of Hawaii as ridge building develops over the Rockies with an associated diffluent pattern. This ridge building is associated with the numerous EC events that repeatedly develop in the left-exit region of the NPJ (not shown). There is an anomalously high PW field associated with these ECC events that extends from Hawaii towards the Pacific Northwest and into coastal Alaskan regions (Fig. 3.16c). In the composite mean PW field, there is a tongue of higher-PW air that overlaps the aforementioned PW anomalies (Fig. 3.16d). This
tongue of higher-PW air resembles an atmospheric river (AR). Future work could examine ECCs associated with AR events. The 850-hPa temperature anomaly composite exhibits positive anomalies as the repetitive EC events help build the ridge over the Rockies east of the ECC events (Fig. 3.16e and Fig. 3.16f).

b) During the events

The anomalously low 300-hPa geopotential heights in the Gulf of Alaska have shifted east and the positive height anomalies have intensified as ridge building progresses throughout the event (Fig. 3.17a). In the mean 300-hPa height field, the ridge has amplified, and the trough has deepened compared to (Fig. 3.16b), consistent with ECC events that help to advect heat and moisture poleward and cool dry air equatorward (Fig. 3.17b). This pattern indicates that the NPJ has potentially fractured into a dual jet formation which would support stronger and more long-lived ECs. The anomalous PW field, has now progressed further east and poleward as the ridge strengthens, and is continually reinforced by these ECC events (Fig. 3.17c). The mean PW values have decreased as the aforementioned AR is losing its support from the ECs as the trough and cool dry air starts to propagate equatorward (Figs. 3.17b and 3.17d). The 850-hPa temperature anomalies shown in figure 3.17e have intensified with positive anomalies under the developing ridge, and negative anomalies associated with the trough advecting equatorward shown in figure 3.17b. The thermal gradient is intensifying, which is strengthening the baroclinic zone leading to a stronger jet, and, hence, stronger cyclones (not shown). Furthermore, in the mean 850-hPa temperature field, it is once again
evident that the ridge has amplified as the thermal field along the west coast of North America advects poleward (Fig. 3.17f).

c) Twelve days after the events’ onset

The 300-hPa height field exhibits anomalously negative heights over the Gulf of Alaska, which have expanded both east and west (Fig. 3.18a). The mean 300-hPa height field exhibits a new trough propagating eastward from Asia across the North Pacific, which suggests the large-scale flow pattern is evolving out of the pattern observed in this section (Fig. 3.18b). Figure 3.18a reveals that the positive height anomalies observed over the western United States have started to propagate eastward across eastern Canada. The anomalous PW maximum has dissipated across the region, as there are fewer ECs transporting PW poleward at this time (Fig. 3.18c). The mean PW field reveals the higher-PW tongue has weakened further across the west coast of North America, suggesting a loss of ECs that were assisting in advecting this higher PW air poleward (Fig. 3.18d). The positive 850-hPa temperature anomalies are now south of the Hudson Bay (Fig. 3.18e). Negative temperature anomalies are starting to develop across the western Pacific. Furthermore, the mean 850-hPa temperature field reveals the thermal ridge along the west coast of North America has de-amplified, as the forcing from the ECs is no longer present (Fig. 3.18f).

3.3.3 Climatology of ECC events during the negative phase of ONI

We next present a climatology of ECC location and frequency during seasons of
<=-1 of CPC’s ONI. Dynamic and thermodynamic variables will be composited to determine preferred large-scale flow patterns that support the exhibited ECC frequency and location. Due to the small sample size of only one season for MAM that exhibited an ONI of <=-1, the seasonal averages of ECC will not be discussed.

a) DJF

Figure 3.19a reveals a composite of DJF ECC distribution for 1985, 1989, 1999, 2000, 2008, and 2011, which all observed a value of ONI <=-1. A negative value of ONI corresponds with a La Niña phase of ENSO, which is characterized by a poleward shift in the storm tracks in both the North Pacific and North American East Coast (e.g., Hoerling and Kumar 1997; Hoerling and Kumar 2000; Higgins et al. 2002; Eichler and Higgins 2006). During DJF, there are three main regions of ECC distribution across the NH: over the north-central Pacific, east of Greenland, and in eastern Russia (Fig. 3.19a). Figure 3.20a reveals anomalous, positive 300-hPa heights south of the Aleutian Islands extending into the Gulf of Alaska. Poleward of this positive height anomaly in the Pacific is a negative height anomaly, which extends into the Arctic and south across the Yukon regions of Alaska and Canada. This pattern would support a poleward shifted NPJ and an increase in ECC events in the Bering Sea. The mean 300-hPa height field reveals a diffluent pattern over Alaska and Canada, suggesting the ECC events in the Bering Sea are occurring in the left exit region of the NJP (Fig. 3.21a). There are negative 300-hPa height anomalies east of Greenland, and positive height anomalies over western Europe, which would support a poleward-shifted North Atlantic jet with the ECC
events occurring in the right entrance and left-exit regions of the jet. A 300-hPa trough axis over eastern North America is associated with a subtle diffluent pattern east of Greenland (Fig. 3.21a).

b) JJA:

Figure 3.19b reveals a composite of JJA ECC distribution for 1988, 1999, and 2010, which all observed a value of ONI <=−1. During JJA, the North Pacific has similar ECC frequency and distribution compared to the DJF composites; however, there is a new maximum in the Gulf of Alaska (Fig. 3.19b). There are two ECC maxima in the North Atlantic, one east of Newfoundland, and the second poleward of the British Isles. The maximum in north-central Russia has shifted east and north compared to DJF. The 300-hPa anomalous height field exhibits two locations of negative heights, over north-central Russia and poleward of the Aleutian Islands (Fig. 3.20b). The North Atlantic has positive height anomalies just south of the ECC maximum. Negative height anomalies are observed in east Russia, which supports the ECC events that occurred in this region. The mean 300-hPa geopotential height field exhibits a wavenumber-five pattern (Fig. 3.21b). One trough is located poleward of the Aleutian Islands, which is associated with a diffluent pattern over the Gulf of Alaska where an ECC maximum occurred, which suggests ECC events occurred in the left-exit region of the NPJ. A second trough is located over north-central Russia where an ECC maximum occurred. The final two troughs were observed over eastern North America and over the northwestern Europe. In between these two troughs is a diffluent pattern, which support the frequency of ECC
events observed.

c) SON

Figure 3.19c reveals a composite of SON ECC distribution for 1988, 1995, 1998, 1999, 2007, 2010, and 2011, which all observed a value of ONI $\leq -1$. During SON, there were ECC maxima east of Greenland, south of the Hudson Bay, over north-central Russia, and in the central North Pacific and Gulf of Alaska (Fig. 3.19c). The 300-hPa height field exhibits negative height anomalies over the Gulf of Alaska that extend west towards the Bering Straight, as well as positive height anomalies equatorward of these negative height anomalies that extend across the North Pacific (Fig. 3.20c). The NPJ would be expected to be in between these negative and positive height anomalies, and ECC events would form in the left-exit region of the NPJ. The mean 300-hPa height field exhibits a trough in the north-central Pacific with a diffluent pattern over the Bering Straight extending into the Gulf of Alaska (Fig. 3.21c). This height pattern supports the frequency of ECC events observed across the North Pacific.

3.3.4 Negative ONI case composites

Individual cases of ECC events were composited during the DJF seasons with values of ONI that were $\leq -1$. To get a better understanding of the dynamic and thermodynamic processes that are associated with this ONI regime, mean and anomaly composite analyses were constructed for three days before, during, and 12 days after each ECC event for precipitable water, 300-hPa geopotential heights, and 850-hPa
temperatures. 12 days after the event was chosen to be able show that the large-scale flow pattern changes over the 1–2 week time frame. The composites for negative ONI were composed of 11 separate ECC events (i.e., 27 December 1984, 31 January 1985, 26 February 1989, 06 February 1989, 05 January 1989, 14 February 1999, 06 February 2000, 20 January 2000, 17 January 2008, 19 February 2011, and 07 January 2011) that occurred within a similar region.

a) Three days before the event:

The central North Pacific near the Kamchatka Peninsula is the region selected for the negative ONI cases (See Fig. 3.19a). The 300-hPa height field exhibits anomalous positive heights in the Gulf of Alaska, and anomalous negative heights across the Kamchatka Peninsula (Fig. 3.22a). The NPJ would be observed between these two height anomaly centers, suggesting a retracted phase of the NPJ. The mean 300-hPa height field exhibits a trough over the western Pacific, with ridging and a diffuent pattern in the height field across the Gulf of Alaska and the Bering Straights, respectively (Fig. 3.22b). The PW field exhibits positive PW anomalies associated with the ECC events extending from west of Hawaii poleward towards the Bering Sea (Fig. 3.22c). East of the positive PW anomalies are negative PW anomalies off the West Coast of the United States. The 850-hPa temperatures exhibit positive temperature anomalies over the Aleutian Islands, which suggests that ridge building is occurring over this location (Fig. 3.22e). West of these positive temperature anomalies are negative temperature anomalies over the Kamchatka Peninsula. The mean 850-hPa temperature field and 300-hPa height
field exhibits evidence of ridge building across the Bering Sea (Figs. 3.22a and 3.22f). The tightening of the thermal gradient over the Kamchatka Peninsula and developing baroclinic zone allows ECs to develop along the thermal boundary and support the ECC events about to occur.

b) During the event

Negative 300-hPa height anomalies are evident across western North America, as ridging across Alaska forces cool dry air to equatorward across this region (Fig. 3.23a). The mean 300-hPa height pattern exhibits a quasi wavenumber-three pattern across the NH, with troughs over eastern North America, eastern Europe, and Japan (fig. 3.23b). The large-scale flow pattern is becoming more amplified across the NH, because of this perturbation to the NPJ waveguide. Positive PW anomalies continue to advect poleward during the ECC events as the ECs pull moisture from the tropical Pacific into the midlatitudes (Fig. 3.23c). The negative PW anomalies have shifted eastward and equatorward across western North America as the trough starts to dig south across this region (See Figs. 3.23a and 3.23b). The mean PW values stay constant; however, the higher values propagate east with the large-scale flow pattern (Fig. 3.23d). The 850-hPa temperature anomaly field reveals that positive temperature anomalies over the Bering Sea have advanced poleward (Fig. 3.23e). The negative 850-hPa temperature anomalies over western North America have advanced equatorward, as well as expanded as northerly flow on the east flank of the ridge forces down cold air from the Arctic. The mean 850-hPa temperature field reveals lower values are penetrating equatorward from
the Arctic into western and central North America (Fig. 3.23f). A temperature gradient has developed across western and central North America as a thermal ridge develops across the Gulf of Alaska and into the Arctic, a situation favorable for a subsequent cold-air outbreak across North America.

c) Twelve days after the events

The 300-hPa height field pattern indicates that positive anomalies have formed over the Bering Sea, while negative height anomalies have developed in the Gulf of Alaska (Fig. 3.24a). The mean 300-hPa height field reveals that a ridge still exists over the Bering Sea, although it has weakened considerably (Figs. 3.23b and 3.24b). Anomalous positive PW values are found across the Bering Sea in response to the presence of an AR equatorward of this region (Fig. 3.24c). Higher mean PW values are advancing poleward north of Hawaii (Figs. 3.24a and 3.24d). Negative 850-hPa temperatures anomalies are observed in the Gulf of Alaska and the Yukon regions of Canada as the ridge breaks down in the central North Pacific and lower heights migrate into the area (Fig. 3.24e). A new thermal ridge in the 850-hPa temperature field is developing poleward into the Bering Sea as the 850-hPa thermal gradient tightens (Fig. 3.24f). This developing thermal ridge would cause a trough to develop downstream in the Gulf of Alaska where positive 850-hPa temperature anomalies were previously observed.

3.3.5 Difference in ECC distribution between positive and negative ONI phases
The differences in ECC distribution in the NH discussed in the previous two sections will be presented. The previously shown ECC density maps for positive ONI and negative ONI were subtracted from one another (positive ONI − negative ONI) to determine the difference in ECC distribution and frequency.

a) DJF

During the positive phase of the ONI, the Gulf of Alaska has the dominant signal for increased ECC frequency (Fig. 3.25a). During a negative phase of the ONI, this maximum is shifted westward towards the central North Pacific, and extends southwest towards Japan. In the North Atlantic, during the negative phase of the ONI, the ECC distribution is shifted farther north from the Canadian Maritimes towards the British Isles. During the positive phase of the ONI, the ECC distribution is shifted south across the North Atlantic extending toward western Europe.

b) JJA

An ECC maximum across the central Pacific and a poleward shift in the storm track is observed during the negative phase of the ONI, whereas an equatorward-shift in the storm track and an ECC maximum across the Gulf of Alaska is observed during the positive phase of the ONI (Fig. 3.25b). In the North Atlantic, there is a poleward shift in ECC frequency over the Canadian Maritimes, British Isles, and off the coast of Greenland during the negative ONI. During the positive ONI, the maximum in ECC frequency is shifted slightly equatorward across the North Atlantic and extends into
Western Europe. During JJA, there seems to be more disparity among ECC distribution in north-central Russia, with a maximum in ECC frequency during positive ONI seasons just east of the Ural Mountain range extending from the Arctic Circle to the Kazakhstan border, as well as extending east into the west Siberian Plains. During the negative ONI seasons, the ECC maximum is shifted east over the Kamchatka Peninsula and poleward into the Arctic Circle.

c) SON

During the negative phase of the ONI, there is a maximum of ECC events south of the Hudson Bay, and across the North Pacific from the Kamchatka Peninsula into the Gulf of Alaska (Fig. 3.25c). This maximum is shifted poleward compared to the positive ONI. The North Atlantic exhibits ECC maxima south of the Hudson Bay and northeastern Europe during the negative phase of the ONI. During the positive phase of the ONI, an ECC maximum is observed in an equatorward shift across the North Atlantic, and has a maximum across the British Isles compared to the negative phase of the ONI.

3.4 Climatology of ECCs during the positive and negative phases of the NAO

A climatology of ECC location and frequency during seasons of >=1 for NAO is now presented. Dynamic and thermodynamic variables will be composited to determine preferred large-scale flow patterns that support the exhibited ECC frequency and location.

The NAO is defined as a north–south dipole of geopotential height and sea-level
pressure anomalies over the North Atlantic (Wallace and Gutzler 1981; Hurrell 1995; Archambault et al. 2010). The positive phase of the NAO is typically characterized by a strong Icelandic low and Azores high, or negative height anomalies near Iceland and positive height anomalies near the Azores. This phase is established with an anomalously strong southwest–northeast-positioned jet stream and storm track that extends from near the New England coast to northern Europe. Comparatively, the negative phase of the NAO is typically characterized by a weak Icelandic low and Azores high with the corresponding positive height anomalies and negative height anomalies, respectively. This phase is established with a weak west–east-positioned jet stream and storm track that extends from the Mid-Atlantic coastline to central Europe (Lau 1988; Rogers 1990; Archambault et al. 2010). Figure 3.26 illustrates the seasonal variability of the phase and amplitude of the NAO from 1979–2014. Certain years stand out as representative strong positive NAO years (i.e., 1979, 1983, 1992, 1995, 2000, and 2012). Conversely, certain years also stand out as representative strong negative NAO years (i.e., 1979, 1985, 1993, 2005, 2010, 2011, and 2012).

3.4.1 Climatology of ECC events during the positive phase of the NAO

a) DJF

Figure 3.27a reveals a composite of DJF ECC distribution for 1989, 1995, 2000, and 2012, which all observed a value of NAO $\geq 1$. A poleward shift is observed in the North Atlantic storm track as well as ECC maxima over Iceland and east of Greenland in comparison to a negative NAO phase which will be shown in later sections (Fig. 3.27a).
A high ECC frequency is indicated across the central North Pacific where an equatorward shift in the storm track is observed. Negative 300-hPa height anomalies over Greenland and Iceland, and positive 300-hPa height anomalies equatorward of the negative height anomalies over the Azores would support a west–east-orientated jet stream would be located between these height anomalies (Fig. 3.28a). The mean 300-hPa height field exhibits a diffuent pattern over western Europe and off the coast of Greenland (Fig. 3.29a). These diffuent patterns support the distribution of ECC events observed over these regions.

b) MAM

Figure 3.27b reveals a composite of MAM ECC distribution for 1989, and 1992, which both observed a value of NAO $\geq 1$. A maximum of ECC frequency extends from eastern Canada eastward towards Greenland and into northwestern Europe (Fig. 3.27b). An equatorward shifted storm track and ECC distribution in comparison to a negative NAO which will be shown in later sections is observed across the North Pacific from Japan to the Gulf of Alaska (Fig. 3.27b). Negative height anomalies over Greenland and Iceland, as well as positive height anomalies equatorward over the Azores support a west–east-orientated North Atlantic jet from the North East coast line towards northwestern Europe, that would be positioned in between these two height anomalies (Fig. 3.28b). This position of the North Atlantic jet supports the distribution of ECC events east of Greenland and over northwestern Europe as many ECs would be forming in the left-exit, and right-entrance regions of this jet (Not shown). The mean 300-hPa
height field exhibits a diffluent pattern in the height field over western and northwestern Europe. An upper-level trough is observed in the mean 300-hPa height field over eastern North America, which would support the distribution of ECC events observed over Greenland and Iceland.

c) JJA

Figure 3.27c reveals a composite of JJA ECC distribution for 1979, 1983, and 1994, which all observed a value of NAO >= 1. A poleward shift in storm track and ECC distribution is observed extending from the Canadian Maritimes towards Iceland (Fig. 3.27c). An equatorward shifted storm track and ECC distribution in comparison to a negative NAO phase shown in later sections is observed across the North Pacific from Japan to the Gulf of Alaska, as well as an ECC maximum in frequency east of the Ural Mountain range in north-central Russia (Fig. 3.27c). Positive 300-hPa height anomalies are observed extending from Newfoundland to the British Isles across the North Atlantic and negative height anomalies are identified equatorward over the Azores, which would support a southwest–northeast-orientated North Atlantic jet in between these two height anomalies (Fig. 3.28c). This position of the North Atlantic jet would support the distribution of ECC events and storm tracks across the North Atlantic. A diffluent height pattern in the mean 300-hPa height field is identified across southern portions of Greenland and over Iceland (Fig. 3.29c). This diffluent pattern would support the ECC distribution observed across this region.

d) SON
Figure 3.27d reveals a composite of SON ECC distribution for 1982, and 1986, which both observed a value of NAO $\geq 1$. A poleward shift in the storm track compared to a negative NAO phase shown in later sections is identified across the North Atlantic that extends from the Canadian Maritimes towards northwestern Europe, as well as an observed ECC maximum that extends from eastern Russia towards central Russia (Fig. 3.27d). Negative height anomalies extending from Canada northeastward towards Iceland, would support a southwest–northeast-orientated North Atlantic jet equatorward of these negative height anomalies, and ECs would form in the right-entrance and left-exit regions of this jet (Not shown). The mean 300-hPa geopotential height fields reveal a weak diffluent pattern over eastern Greenland and Iceland (Fig. 3.29d), which are also located downstream of a trough axis that is located over eastern North America, and thus, they are located under an area of persistent divergence, which would support the ECC frequency and distribution across this region.

3.4.2 Positive NAO case composites

Individual cases of ECC events were composited during DJF with NAO values that were $\geq 1$. To get a better understanding of the dynamic and thermodynamic processes that are associated with the positive phase of the NAO, mean and anomaly precipitable water, 300-hPa geopotential heights, and 850-hPa temperatures were composited for three days before, during, and twelve days after each ECC event. 12 days after the event was chosen to be able show that the large-scale flow pattern changes over the 1–2 week time frame. The composites for the positive NAO cases were composed of
six separate ECC events (i.e., 01 February 1989, 05 December 1994, 15 February 1995, 25 February 2000, 01 February 2012, and 22 December 2011) that occurred within a similar region.

a) Three days before the events

The region selected for positive NAO cases is located over Greenland and Iceland (Fig. 3.27a). The 300-hPa height field exhibits anomalous negative heights off the east coast of Greenland and over Iceland (Fig. 3.30a). Anomalous positive heights over the eastern half of the United States are suggestive of an anomalous ridging event. Anomalous positive heights over Europe combined with the anomalous positive heights over the eastern half of the United States signify a ridge-trough-ridge pattern in the jet stream across the North Atlantic. The mean 300-hPa height field reveals a trough axis south of Greenland, and ridging over Europe and eastern North America (Fig. 3.30b). A diffluent height pattern off the east coast of Greenland supports the ECC frequency in this region, as this region would support the left-exit region of the North Atlantic jet stream (Not shown). The PW field reveals anomalous positive values over the eastern half of North America, and anomalous negative values of PW off the coast of Newfoundland and extending poleward towards the east coast of Greenland (Fig. 3.30c). Anomalous positive values of PW over the Azores are identified; this moisture may be transported poleward by the ECs during the ECC events. The mean PW field reveals higher values of PW advancing poleward across the east North Atlantic (Fig. 3.30d). Over the central North Atlantic, there are lower values of PW; however, these values progressively
increase over eastern North America. The 850-hPa temperatures reveal anomalous positive 850-hPa temperatures across eastern North America, as well as anomalous negative 850-hPa temperatures across the central North Atlantic and Greenland (Fig. 3.30e). This pattern suggests that ridge building is occurring over eastern North America, which will help develop and advect the anomalously cold 850-hPa temperatures equatorward that will deepen the trough previously identified. The mean 850-hPa temperature field reveals ridging in both eastern North America and Europe, and also reveals a thermal gradient along Greenland and Iceland, which will subsequently be able to support propagating ECs along this boundary (Fig. 3.30f).

b) During the events

The 300-hPa height field reveals positive height anomalies previously identified over eastern North America have expanded poleward, but have not progressed eastward (Fig. 3.31a). This expansion poleward has shifted the anomalous negative heights observed over Greenland and western Iceland towards northwest Europe. Positive height anomalies are observed equatorward over the Azores, which would support the North Atlantic jet in between the aforementioned negative height anomalies, and the positive anomalies over the Azores. The mean 300-hPa geopotential height pattern reveals the trough previously discussed has now become elongated as a secondary trough is propagating into eastern North America (Fig. 3.31b). This elongation supports a southwest–northeast-orientated jet pattern (Not shown), as well as a slight diffluent pattern over Iceland supporting the ECC events occurring over this region. These ECs
would likely be forming in the left-exit region of the North Atlantic jet stream. The positive PW anomalies previously identified over the eastern North Atlantic have expanded and now extend from the central North Atlantic towards the British Isles and northwestern Europe, which suggests ECs are aiding in the polward transport of moisture (Fig. 3.31c). The mean PW composite reveals higher PW values are advancing poleward as the ECs transport moisture poleward through their warm sectors (Fig. 3.31e). The 850-hPa temperature anomaly composite reveals the poleward expansion of warm air across southern Canada and eastern North America (Fig. 3.31e). The mean 850-hPa temperature field reveals the thermal gradient previously observed across eastern North America has propagated over the central North Atlantic (Fig. 3.31f). This thermal gradient allows for the formation of ECs and ECC events to occur as the thermal gradient is perturbed along this boundary (Fig. 3.31f).

c) Twelve days after the events

Positive 300-hPa height anomalies over the central North Atlantic have expanded both eastward and westward towards eastern North America and western Europe (Fig. 3.32a). Negative 300-hPa height anomalies from southern Greenland and Iceland have propagated to central Greenland and parts of the Arctic. This pattern effectively has shifted the southwest–northeast-orientated jet poleward across the North Atlantic (Not shown). The mean 300-hPa height field exhibits ridging across the North Atlantic as the ECC events have reinforced the ridge by transporting warm moist air poleward (Fig. 3.32b). The positive PW anomalies have expanded poleward across the southern tip of
Greenland, which supports the transport of warm moist air poleward and ridge building across this region (Fig. 3.32c and Fig. 3.32d). The 850-hPa temperature field reveals that ridge building is occurring at this time as positive temperature anomalies are observed off the coast of Newfoundland and extend eastward towards the British Isles (Fig. 3.32e). The mean 850-hPa temperature field reveals the aforementioned thermal gradient has dissolved across the western Atlantic, subsequently removing the boundary necessary for EC formation to occur (Fig. 3.32f).

3.4.3 Climatology of ECCs during the negative phase of the NAO

a) DJF

Figure 3.33a reveals a composite of SON ECC distribution for 1979, 1985, 2010, and 2011, which all observed a value of NAO \( \leq -1 \). During DJF, the overall ECC distribution reveals an equatorward shift in the storm track compared to the positive phase of the NAO (see Fig. 3.27a) across the North Atlantic, with maxima in ECC activity off the coasts of Newfoundland, Portugal, and over the British Isles (Fig. 3.33a). In the North Pacific, maxima are identified north of the Kamchatka Peninsula and over the Gulf of Alaska. The 300-hPa height field reveals positive height anomalies across Greenland and Iceland, and negative height anomalies equatorward across the Azores (Fig. 3.34a). Across the subtropical regions of the North Atlantic are anomalous positive height anomalies, which would support an abnormally strong subtropical jet across the central North Atlantic, and the observed southern storm track. The mean 300-hPa height field exhibits a trough over eastern North America and a broad diffluent pattern in the
height field across the central North Atlantic (Fig. 3.35a). This pattern supports the distribution of ECC events observed across the North Atlantic.

b) MAM

Figure 3.33b reveals a composite of SON ECC distribution for 2005, 2008, and 2010, which all observed a value of NAO <=−1. During MAM, the ECC distribution reveals an equatorward extension of ECC events across central North America that extends eastward towards western Europe compared to the positive phase of the NAO (Figs. 3.27b and 3.33b). A maximum of ECC events across eastern Russia is observed; however, a minimum in ECC events is identified across central and western Europe, which is suggestive of a trough-ridge-trough-ridge pattern in the midlatitude jet stream from eastern North America to Russia. The 300-hPa height composites reveal positive height anomalies over Greenland and negative height anomalies over the Azores (Fig. 3.34b). Positive 300-hPa height anomalies extend from the Caribbean Sea across the subtropical North Atlantic, which in conjunction with the negative height anomalies, to the north would support a strong subtropical jet extending from southern North America towards western Europe. The mean 300-hPa height field reveals a trough over eastern North America and a diffluent pattern in the height field downstream of the trough axis across the North Atlantic, which would support the ECC distribution observed over North America and western Europe (Fig. 3.35b).

c) JJA
Figure 3.33c reveals a composite of JJA ECC distribution for 1980, 1993, 1998, 2008, 2009, 2011, and 2012, which all observed a value of NAO $\leq -1$. During JJA, a noticeable equatorward shift in the storm track and ECC distribution across the North Atlantic is identified compared to the positive NAO phase composites (Figs. 3.27c and 3.33c). The 300-hPa height field reveals anomalous positive heights over Greenland (Fig. 3.34c). Negative height anomalies are observed over the British Isles that extend poleward towards northwestern Europe. This pattern suggests that the North Atlantic jet is shifted equatorward compared to the positive NAO phase composites, which supports the equatorward shifted storm track observed in Figure 3.33c. The mean 300-hPa geopotential height field reveals a trough over eastern North America and a ridge over Greenland (Fig. 3.35c). A broad trough is identified over the British Isles, which supports the distribution of the ECC events observed in this region.

d) SON

Figure 3.33d reveals a composite of SON ECC distribution for 2006, 2010, and 2012, which all observed a value of NAO $\leq -1$. During SON, a noticeable equatorward shift in the storm track and ECC distribution across the North Atlantic is observed (Fig. 3.33d). Figure 3.33d reveals a maximum of ECC frequency north of the British Isles. Positive 300-hPa height anomalies are observed over Greenland and negative height anomalies are identified over the British Isles, which support an equatorward shifted North Atlantic jet and equatorward shifted ECC distribution (Figs. 3.33d and 3.34d). The mean 300-hPa height field reveals a trough over eastern North America and Western
Europe, in conjunction with broad diffuuent pattern on the downstream side of the trough over eastern North America, which supports the ECC distribution observed (Fig. 3.35d).

3.4.4 Negative NAO case composites

Individual cases of ECC events were composited during the DJF seasons with values of NAO that were $\leq -1$. To get a better understanding of the dynamic and thermodynamic processes that are associated with the negative phase of the NAO, mean and anomaly precipitable water, 300-hPa geopotential heights, and 850-hPa temperatures were composited for three days before, during, and Twelve days after each ECC event. Twelve days after the event was chosen to be able show that the large-scale flow pattern changes over the 1–2 week time frame. The composites for the negative NAO cases were composed of 14 separate ECC events (i.e., 05 February 1979, 21 February 1979, 11 January 1985, 08 February 1985, 18 January 1985, 02 December 2009, 08 December 2009, 05 January 2010, 10 February 2010, 16 February 2010, 08 December 2010, 12 January 2011, 26 January 2011, and 22 February 2011) that occurred within a similar region.

a) Three days before the events

The region selected for negative NAO cases is off the east coast of Newfoundland, extending eastward towards the Azores and the Iberian Peninsula (See Fig. 3.33a). The 300-hPa height field exhibits anomalous positive heights across Greenland, and anomalous negative heights to the east of Newfoundland that extend both
eastward and westward (Fig. 3.36a). This anomalous height field pattern supports an equatorward-shifted, weak North Atlantic jet compared to the positive phase of the NAO that is typically associated with a negative NAO (Figs. 3.28a and 3.36a). The mean 300-hPa height field reveals a diffuent pattern across the eastern and central North Atlantic, which supports the distribution of ECC events observed across the central North Atlantic (Fig. 3.36b). A trough is observed in the 300-hPa mean height field across eastern North America, and east of Newfoundland, which resembles a cyclonic wave break. The PW field reveals anomalous positive values of PW extending from the Gulf of Mexico across the sub-tropical North Atlantic towards the Iberian Peninsula (Fig. 3.36c). Poleward of these positive PW anomalies, are negative PW anomalies. The mean PW field reveals an AR that is advecting moist air from the sub-tropical regions of the central North Atlantic poleward towards the Azores and Iberian Peninsula (Fig. 3.36d). The 850-hPa temperature field reveals positive temperature anomalies across Greenland and northern parts of Canada where a blocking ridge is located (Fig. 3.36e). Equatorward of the positive temperature anomalies are negative temperature anomalies across the United States and off the coast of Newfoundland, as expected due to the anomalous trough observed across this region (Figs. 3.36a and 3.36e). The mean 850-hPa temperature field reveals a thermal gradient across eastern North America and off the coast of Greenland, which would provide a baroclinic zone for ECs to propagate along and create the ECC distribution observed (Fig. 3.36f).

b) During the events
The 300-hPa height field reveals that the positive height anomalies across the subtropical regions of the North Atlantic have expanded, suggesting that the sub-tropical jet is extending and potentially strengthening across the North Atlantic (Fig. 3.37a). The trough–ridge pattern observed three days before the event in the mean 300-hPa height field has become less amplified; however, the diffluent pattern observed across the North Atlantic is still present, which supports the distribution of ECC events observed in the equatorward shifted storm track compared to the positive NAO phase storm track (Figs. 3.27a and 3.37b). The PW anomalies have shifted equatorward compared to the onset of the events (Fig. 3.37c). The mean PW composite doesn’t reveal any significant differences across the North Atlantic compared to the onset of the event (Fig. 3.37d). The 850-hPa temperature anomalies reveal that the thermal gradient between southern Greenland and off the northeast coast of the United States has decreased during the event compared to the onset of the event; however, this gradient has also shifted equatorward, and closer to the northeast coast of the United States (Fig. 3.37e). The mean 850-hPa temperature field reveals the thermal gradient across the North Atlantic has considerably weakened; however, the block still remains in place and lower temperatures are still observed across eastern North America (See Figs. 3.37a and 3.37f).

c) Twelve days after the events

The negative 300-hPa geopotential height anomalies that were previously located in the central North Atlantic have shifted eastward towards the eastern North Atlantic (Fig. 3.38a). The shift of the negative height anomalies would likely supports that the
ECC events are traversing across the North Atlantic, which also suggests that the North Atlantic jet is in an extended phase (Not shown). The positive height anomalies over Greenland have considerably weakened since the onset of the event. The mean 300-hPa height field reveals the trough over eastern North America has propagated eastward towards the central North Atlantic, which has shifted the diffluent pattern associated with the trough eastward towards the eastern North Atlantic (Fig. 3.38b). The positive PW anomalies that were previously identified in the central North Atlantic have advanced eastward, and are now along the West Coast of Africa and the Iberian Peninsula (Fig. 3.38c). This pattern is supported by the eastward shift of ECs and ECC events. The mean PW composites reveal that there are lower values of PW in the eastern North Atlantic compared to during the event, which suggests that the higher values of PW are being fluxed north by the ECs at a lesser rate over this region (Fig. 3.38d). The 850-hPa temperature anomalies reveal that the positive temperature anomalies over northeast Canada have shifted eastward towards Greenland, which suggests that a negative NAO pattern may start to rebuild over the region (Fig. 3.38e). Negative temperature anomalies across the western and central North Atlantic are shifted eastward towards the eastern North Atlantic where the ECC events have also shifted (Fig. 3.38e). This pattern supports that ECs are traversing along the thermal gradient present across the North Atlantic and forming in the right-entrance and left-exit regions of the North Atlantic jet (Not shown) (Fig. 3.38f).

3.4.5 Differences in ECC distributions between positive and negative NAO phases
The differences in ECC distribution in the NH discussed in the previous two sections are presented. The previously shown ECC density maps for positive NAO and negative NAO were subtracted from one another (i.e., positive NAO − negative NAO) to determine the differences in ECC distribution and frequency.

a) DJF

Figure 3.39a reveals that negative NAO phases have an equatorward shifted storm track and an ECC distribution that extends from off the coast of Newfoundland towards Western Europe compared to the positive NAO phase which has a poleward shifted storm track and ECC distribution. Positive NAO phases have a poleward shifted storm track and ECC distribution that extends from south of Greenland towards Iceland and into northwestern Europe compared to the negative NAO phase storm track and ECC distribution.

b) MAM

Figure 3.39b highlights that the negative phase of the NAO during MAM has an equatorward shifted storm track and ECC distribution, which extends from the central and Mid-Atlantic regions of the United States across the central North Atlantic compared to the positive NAO phase which has a poleward shifted storm track and ECC distribution. The positive NAO has a poleward shifted storm track and ECC distribution that extends from northeast Canada towards Iceland and northwestern Europe compared to the negative NAO phase storm track and ECC distribution. A maximum of ECC
distribution is identified across the northern tier of the United States from Montana to Maine.

c) JJA

Figure 3.39c highlights that the negative phase of the NAO during JJA has an equatorward shifted storm track and distribution of ECC events that extends from south of Newfoundland towards the British Isles and the Scandinavian Peninsula compared to the positive NAO phase which has a poleward shifted storm track and ECC distribution. There was also an ECC maximum south of the Hudson Bay during the negative phase of the NAO. During the positive phase of NAO, the ECC distribution and storm track has shifted poleward from Newfoundland towards Greenland and Iceland in a southwest–northeast-orientated storm track compared to the negative NAO phase storm track and ECC distribution.

d) SON:

Figure 3.39d highlights that the negative phase of the NAO during SON has an equatorward extension to the storm track and ECC distribution in the central North Atlantic compared to the positive NAO phase which has a poleward shifted storm track and ECC distribution. Figure 3.39d reveals a poleward shift of ECC distribution across the North Atlantic during the positive phase of the NAO, which extends over Iceland and Greenland as expected compared to the negative NAO phase storm track and ECC distribution. The signal for the negative phase of the NAO is not as prevalent as
compared to the previous seasons; however, the negative phase of the NAO does have a faint signal of an equatorward shift in its ECC distribution.

3.5 Climatology of ECC events during the positive and negative phases of the PNA

We next present a climatology of ECC location and frequency during seasons of \( \geq 1 \) for the PNA. Dynamic and thermodynamic variables will be composited to determine preferred large-scale flow patterns that support the exhibited ECC frequency and location.

The PNA is characterized by four height anomalies of opposite sign that resemble a stationary Rossby wave train that extends from Hawaii in the east-central tropical Pacific to the southeastern United States (e.g., Wallace and Gutzler 1981; Barnston and Livezey 1987; Archambault et al. 2010). The positive phase of the PNA is characterized by a ridge–trough pattern over North America, where a ridge is observed over western North America and a trough is observed over eastern North America (e.g., Leather et al. 1991; Archambault et al. 2010). In the central Pacific, a pronounced trough is also observed during the positive phase of the PNA. This pattern would displace the storm track and ECC events equatorward compared to the climatological average over the eastern United States. Conversely, the negative phase of the PNA is characterized by the opposite ridge–trough pattern, with a trough observed over western North America and a ridge observed over eastern North America (e.g., Leather et al. 1991; Archambault et al. 2010). This pattern would displace the storm track and ECC events poleward compared to climatology over the eastern United States. Figure 3.40 illustrates the seasonal

3.5.1 Climatology of ECC events during the positive phase of the PNA

a) DJF

Figure 3.41a reveals a composite of DJF ECC distribution for 1983, and 2003, which all observed a value of PNA >=1. During DJF, Figure 3.41a reveals an equatorward shift of the storm track and ECC distribution across the southeastern United States extending from the Gulf of Mexico towards Greenland compared to the ECC distribution of the negative phase of the PNA shown in later sections. An ECC maximum is observed in the Gulf of Alaska, with a minimum of ECs and ECC events across the western United States. The 300-hPa geopotential height field exhibits anomalous negative heights across the Gulf of Alaska and southeastern United States (Fig. 3.42a). Positive 300-hPa height anomalies are identified across the central United States and Yukon regions of Canada. The mean 300-hPa geopotential height field reveals a deep trough across the eastern United States and another across the Gulf of Alaska (Fig. 3.43a). A diffluent pattern in the height field is identified in the Gulf of Alaska, which flows into a ridge across western North America. This pattern supports the distribution and frequency of ECC events across these regions.
b) MAM

Figure 3.41b reveals a composite of MAM ECC distribution for 1983, 1984, 1993, and 2005, which all observed a value of PNA $\geq 1$. During MAM, Figure 3.41b reveals a maximum in the ECC distribution across the Gulf of Alaska and an equatorward shift of the storm track across the eastern United States that extends from the northern Gulf of Mexico towards Newfoundland compared to the ECC distribution of the negative phase of the PNA shown in later sections. The 300-hPa height field reveals negative height anomalies across the central North Pacific, extending towards the Gulf of Alaska and over the southeastern United States (Fig. 3.42b). Conversely, positive height anomalies across the Yukon regions of Canada and Alaska are observed suggesting a trough–ridge–trough pattern in the midlatitude jet stream waveguide supporting the distribution of ECC events observed. The mean 300-hPa geopotential height field exhibits troughs across the central North Pacific, the Gulf of Alaska, and over the eastern United States (Fig. 3.43b). A diffluent pattern associated with the trough across the Gulf of Alaska supports the distribution of ECC events across this region.

c) JJA

Figure 3.41c reveals a composite of JJA ECC distribution for 1982, 1983, and 2007, which all observed a value of PNA $\geq 1$. For JJA, Figure 3.41c reveals a maximum of ECC events across the central North Pacific extending into the Gulf of Alaska and an equatorward shift of the storm track across the Northeast United States compared to the ECC distribution of the negative phase of the PNA shown in later sections, and extends
eastward towards the British Isles and France. The 300-hPa height field reveals anomalous negative heights observed across the North Pacific near the Kamchatka Peninsula, Gulf of Alaska, the Northeast United States, and France in the North Atlantic basin (Fig. 3.42c). Anomalous positive heights are observed over the north-central United States and south-central Canada, which supports the positive PNA pattern and the distribution of ECC events. The mean 300-hPa height field exhibits troughs over eastern North America and across the central North Pacific, with a ridge identified across western North America (Fig. 3.43c). A diffluent pattern in the height field eastward of the trough axis in the North Pacific supports the distribution of ECC events observed in the Gulf of Alaska.

d) SON

Figure 3.41d reveals a composite of SON ECC distribution for 1979, 1980, 2007 and 2008, which all observed a value of PNA >=1. During SON, Figure 3.41d reveals an equatorward shifted storm track across the eastern United States compared to the ECC distribution of the negative phase of the PNA shown in later sections and two separate ECC maximum that are observed across the central North Pacific and the Gulf of Alaska. The 300-hPa height field reveals anomalous negative heights across the central North Pacific, extending into the Gulf of Alaska, and across eastern North America (Fig. 3.42d). Anomalous positive heights are identified across Western North America supporting the distribution of ECC events observed. The midlatitude jet waveguide would be in an amplified state and form a trough–ridge–trough pattern across the northeast Pacific into
North America. The mean 300-hPa height field reveals the amplified wave pattern over the central North Pacific and North America (Fig. 3.43d). The trough over central North Pacific develops into a diffluent pattern in the height field, which extends into the Gulf of Alaska (Fig. 3.43d). This pattern in the height field supports distribution of ECC events identified as well as the overall flow pattern of the midlatitude jet during a positive phase of the PNA (Not shown).

3.5.2 Positive PNA case composites

Individual cases of ECC events were composited during the DJF seasons with values of PNA that were >=1. To get a better understanding of the dynamic and thermodynamic processes that are associated with the positive PNA, mean and anomaly precipitable water, 300-hPa geopotential heights, and 850-hPa temperatures were composited for three days before, during, and twelve days after each ECC event. 12 days after the event was chosen to be able show that the large-scale flow pattern changes over the 1–2 week time frame. The composites for the positive PNA cases were composed of five separate ECC events (i.e., 12 January 1981, 14 January 1983, 19 February 1983, 23 February 2003, and 15 December 2002) that occurred within a similar region.

a) Three days before the event

The region selected for positive PNA cases is in the Gulf of Alaska extending northwest towards the Aleutian Islands (See Fig. 3.41a). The 300-hPa geopotential height field reveals anomalous negative heights off the Japanese east coast extending
across the North Pacific towards the Gulf of Alaska (Fig. 3.44a). Anomalous positive heights are observed across western North America, while anomalous negative heights are observed across eastern North America. This anomalous height pattern supports a trough–ridge–trough pattern across the North Pacific and North America indicative of a highly amplified mid-latitude jet stream waveguide. The mean 300-hPa geopotential height field reveals a diffluent pattern across the Gulf of Alaska, which supports the distribution of ECC events that are observed (Fig. 3.44b). The PW field reveals anomalous positive values of PW extending from the sub-tropical North Pacific towards coastal regions of Alaska and Canada, which supports ridge building across this region from recurving ECs that are transporting moisture poleward (Fig. 3.44c). The mean PW field reveals an AR extending from Hawaii towards the Canadian and Alaskan coastlines that is being reinforced by the EC events occurring across the Gulf of Alaska (Fig. 3.44d). The 850-hPa temperature composite reveals negative temperature anomalies across western regions of the Gulf of Alaska and eastern North America, which transition into positive anomalies across western North America and off the coast of Newfoundland, supporting the anomalous values of heat-transported poleward (Fig. 3.44e). The mean 850-hPa temperature composite reveals thermal gradients developing across the Gulf of Alaska and eastern North America, which will allow ECs to develop and propagate along to form the ECC events observed (Fig. 3.44f).

b) During the event

The 300-hPa geopotential height field reveals that the negative heights observed
three days prior to the event across the North Pacific have expanded eastward towards southwestern regions of North America (Fig. 3.45a). The positive height anomalies have expanded poleward and intensified across western North America, and negative height anomalies are observed across eastern North America and equatorward into the northern Caribbean. The ridge and trough in the mean 300-hPa geopotential height field over western and eastern North America, respectively, have become amplified as the ECC events across the Gulf of Alaska has built the ridge, which have forced a trough to dig farther equatorward across eastern North America (Fig. 3.45b). The PW field reveals positive anomalies from Hawaii to the coast of Alaska and Canada that has intensified as ECs continually aid in transporting moisture poleward (Fig. 3.45c). These higher values of PW are evident in the mean PW field over the Gulf of Alaska and off the East Coast of the United States, which are collocated with the locations of the ECC events that are occurring and transporting of heat and moisture poleward (Fig. 3.45d). The 850-hPa temperature field reveals negative anomalies across the North Pacific and off the East Coast of North America, and positive anomalies across Alaska and northern Canada, which are a result of ridge building caused by the ECC events over the Gulf of Alaska (Fig. 3.45e). The mean 850-hPa temperature field reveals the continuing tightening of the thermal gradients across the Gulf of Alaska and eastern North America, which increases baroclinic instability and supports the development of ECs to aid in the formation of the ECC events observed (Fig. 3.45f).

c) Twelve days after the events
The 300-hPa height field reveals negative height anomalies over the Gulf of Alaska have propagated eastward into western North America, which has shifted the anomalous positive heights previously over western North America eastward towards central North America and shifted the anomalous negative heights over eastern North America to the western North Atlantic (Fig. 3.46a). The mean 300-hPa geopotential height field reveals a cyclonic wave break over the Gulf of Alaska, which suggests that the ECs recurved northwestward as they encountered the blocking ridge across western North America (Fig. 3.46b). The PW field reveals that the positive PW anomalies previously over western North America have shifted eastward into central North America (Fig. 3.46c). The mean PW field reveals that the higher PW values previously over the Gulf of Alaska have shifted equatorward and eastward towards western North America as the ridge has become less amplified over this region (Fig. 3.46d). The 850-hPa temperature field reveals that the positive height anomalies associated with the ridge over central North America have expanded poleward as it has propagated eastward across the Rocky Mountains (Figs. 3.46a and 3.46e). The mean 850-temperature field reveals the thermal gradients and associated baroclinic gradient over eastern North America and the Gulf of Alaska have weakened considerably, suggesting that ECs are less likely to develop and propagate across this region (Fig. 3.46f).

3.5.3 Climatology of ECC events during the negative phase of the PNA

A climatology of ECC location and frequency during seasons of \( \leq -1 \) for the PNA are presented. Dynamic and thermodynamic variables will be composited to
determine preferred large-scale flow patterns that support the exhibited ECC frequency and location.

a) DJF

Figure 3.47a reveals a composite of DJF ECC distribution for 1979, 1982, 1989, 1990, 2009, and 2011, which all observed a value of PNA <=−1. Figure 3.47a reveals a poleward shifted storm track and ECC distribution across eastern North America compared to the positive PNA ECC distribution, and an eastward shift in ECC distribution over western North America compared to the positive PNA ECC distribution, and a maximum in ECC frequency over the Kamchatka Peninsula. The 300-hPa height field reveals positive height anomalies across the central North Pacific (Fig. 3.48a). Negative height anomalies are observed across the central subtropical North Pacific and over central Canada, suggesting there is a weaker poleward shifted midlatitude jet across the North Pacific (Not shown). The negative height anomalies across Hawaii support possible cut-off lows that are flowing equatorward and westward of this high amplitude-blocking pattern across the North Pacific. The mean 300-hPa height field reveals ridging across the central North Pacific, which leads into a broad trough across central and eastern North America (Fig. 3.49a).

b) MAM

Figure 3.47b reveals a composite of MAM ECC distribution for 1982, 1985, and 2002, which all observed a value of PNA <=−1. For MAM, Figure 3.47b reveals a
poleward shift in the storm track and ECC distribution across eastern North America compared to the positive PNA ECC distribution, and an ECC maximum over the Kamchatka Peninsula and coastal regions of Alaska. The 300-hPa geopotential height field reveals positive height anomalies across the central North Pacific and southern regions of the United States (Fig. 3.48b). Anomalous negative heights are observed over the tropical central Pacific, western North America, and eastern North America (Fig. 3.48b). A ridge–trough–ridge pattern would be expected in the midlatitude jet across the North Pacific and North America regions. The North Pacific 300-hPa height field reveals positive height anomalies that resemble a Rex block pattern with potential cut-off lows flowing equatorward and westward of this block. The mean 300-hPa height field reveals ridging across the North Pacific, which transitions into a broad trough across western and eastern Canada (Fig. 3.49b).

c) JJA

Figure 3.47c reveals a composite of JJA ECC distribution for 1980, 1984, and 2000, which all observed a value of PNA <= -1. During JJA, Figure 3.47c reveals a maxima of ECC frequency across the central North Pacific, and central and eastern Canada, suggesting that there is a poleward shift in the storm track and ECC distribution across North America compared to the positive PNA ECC distribution. The 300-hPa height field reveals positive height anomalies across the central North Pacific, as well as negative height anomalies over Hawaii and western North America (Fig. 3.48c). This 300-hPa height field supports a weakly amplified ridge–trough–ridge pattern across the
North Pacific and North America. These height anomalies also support the ECC distribution observed, as the ECC events develop between these height anomalies, which is where the midlatitude jet would be observed. The mean 300-hPa height field reveals weak troughs and diffluent patterns downstream in both the central North Pacific and western North America, which support the ECC distribution identified (Fig. 3.49c).

d) SON

Figure 3.47d reveals a composite of SON ECC distribution for 1985, 1990, and 1994, which all observed a value of PNA <=−1. During SON, Figure 3.47d reveals that there are ECC maxima along the west coast of Canada and the central North Pacific. A poleward shift of the storm track and ECC distribution is observed across Eastern North America compared to the positive PNA ECC distribution for SON; however, there are a few ECC events that traverse the southeastern United States. The 300-hPa height field reveals positive height anomalies across the central North Pacific and eastern North America, and negative height anomalies across the subtropical central Pacific and eastern North America (Fig. 3.48d). This pattern supports a Rex block across the North Pacific, with a large ridge across the central North Pacific, a trough over western North America, and a ridge over eastern North America. This flow pattern is extremely amplified, and a retracted NPJ would be expected during this phase of the PNA (Not shown). The mean 300-hPa height field reveals a broad ridge across the central Pacific, and a broad trough and diffluent pattern across western and central North America, which supports the ECC distribution and pattern identified (Fig. 3.49d).
3.5.4 Negative PNA case composites

Individual cases of ECC events were composited during DJF with negative values of PNA that were $\leq -1$. To get a better understanding of the dynamic and thermodynamic processes that are associated with a negative PNA, mean and anomaly precipitable water, 300-hPa geopotential heights, and 850-hPa temperature were composited for three days before, during, and twelve days after each ECC event. 12 days after the event was chosen to be able show that the large-scale flow pattern changes over the 1–2 week time frame. The composites for the negative PNA cases were composed of eight separate ECC events (i.e., 21 February 1979, 08 December 1981, 30 December 1981, 07 February 1990, 15 December 2008, 01 February 2009, 14 December 2010, and 11 February 2011) that occurred within a similar region.

a) Three days before the events

The region selected for negative PNA cases is over the Pacific Northwest regions of the United States and western Canada (See Fig. 3.47d). The 300-hPa height field reveals anomalous positive heights across the central North Pacific, signifying an amplified ridge across the region (Fig. 3.50a). Equatorward of these positive height anomalies are negative height anomalies across the central subtropical North Pacific, which is associated with a blocking pattern across the North Pacific. Anomalous negative heights are observed across western Canada and the Pacific Northwest regions of the United States. This 300-hPa height pattern supports a ridge–trough large-scale flow pattern across the North Pacific and western North America. The mean 300-hPa
height field indicates a large-amplified ridge across the central North Pacific, with a diffluent pattern downstream of the ridge across Western North America (Fig. 3.50b). The PW field shows negative PW anomalies across western North America, where the ECC events will be occurring, suggesting that the Rex block is aiding in transporting cool dry air equatorward across the west coast of North America (Fig. 3.50c). The mean PW field reveals low values of PW across western North America, due to the equatorward transport of cool dry air from the Arctic (Figs. 3.50c and 3.50d). The 850-hPa temperature field reveals positive temperature anomalies across the western and central North Pacific (Fig. 3.50e). Negative 850-hPa temperature anomalies are identified across western North America on the downstream side of the blocking ridge. The mean 850-hPa temperatures reveal a strong thermal and baroclinic gradient across western North America, supporting the development and propagation of ECs across this region (Fig. 3.50f).

b) During the events

The 300-hPa height field reveals positive height anomalies across the central North Pacific, with anomalous negative height anomalies over the central subtropical regions of the North Pacific and western North America, signifying a blocking pattern across this region (Fig. 3.51a). The 300-hPa height field reveals a ridge–trough pattern across the North Pacific and North America supporting the distribution of ECC events observed. The trough across western North America will act as a blocking mechanism in the flow pattern for the large blocking ridge to propagate eastward. This pattern is due to
the continued EC formation on the downstream side of the Rex block. The mean 300-hPa height field reveals a diffluent pattern in the height field across western North America supporting the distribution of ECC events (Fig. 3.51b). The PW field reveals positive PW anomalies across the western North Pacific on the upstream side of the ridge, as well as negative PW anomalies across coastal western North America where the trough is located (Fig. 3.51c). The mean PW field reveals a similar pattern as was identified during the onset of the event with lower PW values across western North America and higher PW values across the central North Pacific (Fig. 3.51d). Negative 850-hPa temperature anomalies have advanced equatorward across western North America due to the increased ridge building across the central North Pacific, suggesting that blocking patterns aid in the formation of these types of ECC events to develop along western North America (Fig. 3.51e). The mean 850-hPa temperature field reveals cooler temperatures advancing westward across western North America creating a strong thermal gradient across this region aiding in the development of ECs and ECC events along the baroclinic gradient providing the gradient is perturbed (Fig. 3.51f).

c) Twelve days after the onset of the events

The composited 300-hPa height field reveals positive height anomalies across the central North Pacific, and negative height anomalies across the subtropical regions of the North Pacific and western North America (Fig. 3.52a). The negative height anomalies have shifted poleward toward the Bering Straights which has aided in propagating the positive height anomalies equatorward and westward. This pattern suggests that the
ECC events and blocking patterns are forcing the large-scale pattern to remain stationary as the upstream disturbances struggle to propagate the flow pattern out of this negative PNA phase. The mean 300-hPa height field reveals a weaker amplified ridge and trough across the Central North Pacific and western North America, respectively (Fig. 3.52b). The PW field composite reveals that the positive PW anomalies across the western and central North Pacific have shifted eastward and equatorward, signifying the transport of moisture poleward has been cut off by a lack of EC activity in the western North Pacific (Fig. 3.52c). The mean PW field reveals a similar pattern as was indicated during the event with lower PW values across Western North America as the influences of the Rex block are still present in aiding the transport of dry air equatorward over this region (Figs. 3.52c and 3.52d). As expected, the negative 850-hPa temperature anomalies have shifted poleward and westward as the blocking pattern breaks down and becomes less amplified (Fig. 3.52e). The mean 850-hPa temperatures reveal a tight thermal and baroclinic gradient across Western North America, which supports EC and ECC development along its boundaries providing the temperature field is being perturbed (Fig. 3.52f).

3.5.5 Difference in ECC distribution between positive and negative NAO phases

The differences in ECC distribution discussed in the previous two sections are now presented. The previously shown ECC density maps for positive PNA and negative PNA were subtracted from one another (i.e., positive PNA − negative PNA) to determine the differences in ECC distribution and frequency.

a) DJF
Figure 3.53a reveals the positive PNA phase has preferred distribution of ECCs across the Gulf of Alaska, and has an equatorward shift in the storm track and ECC distribution across the eastern United States compared to the negative phase of the PNA. The negative phase of the PNA has a higher distribution of ECC events across the Kamchatka Peninsula extending southwest towards Japan, and western North America.

b) MAM

Figure 3.53b reveals the positive phase of the PNA has an equatorward shifted, higher ECC distribution across the North Pacific from Japan to the Gulf of Alaska compared to the negative phase of the PNA; conversely, the negative phase has a higher ECC distribution from the Kamchatka Peninsula to Alaska and western North America. An equatorward shift in the storm track and ECC distribution across the eastern United States is indicated during a positive phase of the PNA compared to the negative phase of the PNA. The negative phase of the PNA also has an ECC maximum in the lee of the Rocky Mountains in the United States.

c) JJA

Figure 3.53c reveals that the positive phase of the PNA has an equatorward shifted ECC distribution from Japan to the Gulf of Alaska compared to the negative phase of the PNA; conversely, the negative phase of the PNA has a poleward shifted ECC distribution across the North Pacific. The ECC maximum over the Kamchatka Peninsula during the negative phase of the PNA is shifted poleward towards the Arctic
Circle. A maximum in ECC distribution is identified across western North America and a lee cyclone maximum over the northern Rockies in the United States is also observed during the negative phase of the PNA.

d) SON

Figure 3.53d reveals an equatorward shifted ECC distribution across the North Pacific during the positive phase of the PNA that extends from Japan to the Gulf of Alaska compared to the ECC distribution observed during the negative phase of the PNA. A poleward shift in the North Pacific during a negative phase of the PNA is indicated over the Kamchatka Peninsula and eastward into Alaska, and western North America compared to the ECC distribution observed during the positive phase of the PNA. The lee cyclone maximum over the Rockies is still apparent across the United States.
c. Figures

Fig. 3.1. (a) DJF, (b) MAM, (c) JJA, and (d) SON ECC density seasonal averages across the NH.
Fig. 3.2. (a) DJF, (b) MAM, (c) JJA, and (d) SON mean 300-hPa geopotential height (m) composites from 1979–2014.
Fig. 3.3. (a) The total ECC density average and distribution from 1979–2014 and (b) mean 300-hPa geopotential height (m) composite from 1979–2014.

Fig. 3.4. Total ECC density for 1984.
Fig. 3.5. (a) DJF, (b) MAM, (c) JJA, and (d) SON total ECC density averages for 1984.
Fig. 3.6. (a) Anomaly and (b) mean 300-hPa geopotential height (m) field for 1984.

Fig. 3.7. Total ECC density for 2000.
Fig. 3.8. Total ECC density for DJF of 2000.

Fig. 3.9. (a) Anomaly and (b) mean 300-hPa geopotential height (m) field for 2000.
Fig. 3.10. (a) DJF, (b) MAM, (c) JJA, and (d) SON total ECC density averages for 2010.
Fig. 3.11. (a) Anomaly and (b) mean 300-hPa geopotential height (m) for 2010.
Fig. 3.12. CPC Oceanic Nino Index (ONI) three-month running mean of El Nino Region Sea Surface Temperatures (ERSST V5) SST anomalies in the Nino 3.4 region (5°N–5°S, 120°–170°W) from 1979–2014, where blue is DJF, green is MAM, red is JJA, and black is SON.
Fig. 3.13. (a) DJF, (b) MAM, (c) JJA, and (d) SON total ECC density average for all $\geq 1$ seasonal values of CPC’s ONI.
Fig. 3.14. (a) DJF, (b) MAM, (c) JJA, and (d) SON anomalous 300-hPa geopotential height (m) for all >=1 seasonal values of CPC’s ONI.
Fig. 3.15. (a) DJF, (b) MAM, (c) JJA, and (d) SON mean 300-hPa geopotential height (m) for all \( \geq 1 \) seasonal values of CPC’s ONI.
Fig. 3.16. Mean and anomaly composites for 300-hPa geopotential heights (m) (a,b), precipitable water (kg/m$^2$) (c,d), 850-hPa temperature (k) (e,f), respectively, are for selected positive ONI cases. Composites are for three days prior to the event starting.
Fig. 3.17. Mean and anomaly composites for 300-hPa geopotential heights (m) (a,b), precipitable water (kg/m$^2$) (c,d), 850-hPa temperature (k) (e,f), respectively, are for selected positive ONI cases. Composites are for during the event.
Fig. 3.18. Mean and anomaly composites for 300-hPa geopotential heights (m) (a,b), precipitable water (kg/m$^2$) (c,d), 850-hPa temperature (k) (e,f), respectively, are for selected positive ONI cases. Composites are for twelve days after the event.
Fig. 3.19. (a) DJF, (b) JJA, and (c) SON total ECC density average for all $\leq -1$ seasonal values of CPC’s ONI.
Fig. 3.20. (a) DJF, (b) JJA, and (c) SON anomaly 300-hPa geopotential height (m) for all \(<=-1\) seasonal values of CPC’s ONI.
Fig. 3.21. (a) DJF, (b) JJA, and (c) SON mean 300-hPa geopotential height (m) for all \( \leq -1 \) seasonal values of CPC’s ONI.
Fig. 3.22. Mean and anomaly composites for 300-hPa geopotential heights (m) (a,b), precipitable water (kg/m$^2$) (c,d), 850-hPa temperature (k) (e,f), respectively, are for selected negative ONI cases. Composites are for three days prior to the event.
Fig. 3.23. Mean and anomaly composites for 300-hPa geopotential heights (m) (a,b), precipitable water (kg/m$^2$) (c,d), 850-hPa temperature (k) (e,f), respectively, are for selected negative ONI cases. Composites are for during the event.
Fig. 3.24. Mean and anomaly composites for 300-hPa geopotential heights (m) (a,b), precipitable water (kg/m\(^2\)) (c,d), 850-hPa temperature (k) (e,f), respectively, are for selected negative ONI cases. Composites are for twelve days after the event.
Fig. 3.25. (a) DJF, (b) JJA, and (c) SON ECC density for positive ONI – negative ONI.
Fig. 3.26. NAO seasonal values from 1979–2014, where blue is DJF, green is MAM, red is JJA, and black is SON.
Fig. 3.27. (a) DJF, (b) MAM, (c) JJA, and (d) SON total ECC density average for all \( \geq 1 \) seasonal values for the NAO.
Fig. 3.28. (a) DJF, (b) MAM, (c) JJA, and (d) SON 300-hPa geopotential height (m) anomalies for all >=1 seasonal values of the NAO.
Fig. 3.29. (a) DJF, (b) MAM, (c) JJA, and (d) SON mean 300-hPa geopotential heights (m) for all >=1 seasonal values of the NAO.
Fig. 3.30. Mean and anomaly composites for 300-hPa geopotential heights (m) (a,b), precipitable water (kg/m$^2$) (c,d), 850-hPa temperature (k) (e,f), respectively, are for selected positive NAO cases. Composites are for three days prior the event.
Fig. 3.31. Mean and anomaly composites for 300-hPa geopotential heights (m) (a,b), precipitable water (kg/m$^2$) (c,d), 850-hPa temperature (k) (e,f), respectively, are for selected positive NAO cases. Composites are for during the event.
Fig. 3.32. Mean and anomaly composites for 300-hPa geopotential heights (m) (a,b), precipitable water (kg/m²) (c,d), 850-hPa temperature (k) (e,f), respectively, are for selected positive NAO cases. Composites are for twelve days after the event.
Fig. 3.33. (a) DJF, (b) MAM, (c) JJA, and (d) SON total ECC density average for all $\leq -1$ seasonal values of the NAO.
Fig. 3.34. (a) DJF, (b) MAM, (c) JJA, and (d) SON 300-hPa geopotential height (m) anomalies for all $\leq -1$ seasonal values of the NAO.
Fig. 3.35. (a) DJF, (b) MAM, (c) JJA, and (d) SON mean 300-hPa geopotential height (m) for all $\leq -1$ seasonal values of the NAO.
Fig. 3.36. Mean and anomaly composites for 300-hPa geopotential heights (m) (a,b), precipitable water (kg/m$^2$) (c,d), 850-hPa temperature (k) (e,f), respectively, are for selected negative NAO cases. Composites are for three days prior to the event.
Fig. 3.37. Mean and anomaly composites for 300-hPa geopotential heights (m) (a,b), precipitable water (kg/m$^2$) (c,d), 850-hPa temperature (k) (e,f), respectively, are for selected negative NAO cases. Composites are for during the event.
Fig. 3.38. Mean and anomaly composites for 300-hPa geopotential heights (m) (a,b), precipitable water (kg/m$^2$) (c,d), 850-hPa temperature (k) (e,f), respectively, are for selected negative NAO cases. Composites are for twelve days after the event.
Fig. 3.39. (a) DJF, (b) MAM, (c) JJA, and (d) SON ECC density for positive NAO – negative NAO.
Fig. 3.40. PNA seasonal values from 1979–2014, where blue is DJF, green is MAM, red is JJA, and black is SON.
Fig. 3.41. (a) DJF, (b) MAM, (c) JJA, and (d) SON total ECC density average for all $\geq 1$ seasonal values of the PNA.
Fig. 3.42. (a) DJF, (b) MAM, (c) JJA, and (d) SON 300-hPa geopotential height (m) anomalies for all >=1 seasonal values of the PNA.
Fig. 3.43. (a) DJF, (b) MAM, (c) JJA, and (d) SON mean 300-hPa geopotential height (m) for all >=1 seasonal values of the PNA.
Fig. 3.44. Mean and anomaly composites for 300-hPa geopotential heights (m) (a,b), precipitable water (kg/m$^2$) (c,d), 850-hPa temperature (k) (e,f), respectively, are for selected positive PNA cases. Composites are for three days prior to the event.
Fig. 3.45. Mean and anomaly composites for 300-hPa geopotential heights (m) (a,b), precipitable water (kg/m$^2$) (c,d), 850-hPa temperature (k) (e,f), respectively, are for selected positive PNA cases. Composites are for during the event.
Fig. 3.46. Mean and anomaly composites for 300-hPa geopotential heights (m) (a,b), precipitable water (kg/m$^2$) (c,d), 850-hPa temperature (k) (e,f), respectively, are for selected positive PNA cases. Composites are for twelve days after the event.
Fig. 3.47. (a) DJF, (b) MAM, (c) JJA, and (d) SON total ECC density average for all \(\leq -1\) seasonal values of the PNA.
Fig. 3.48. (a) DJF, (b) MAM, (c) JJA, and (d) SON 300-hPa geopotential height (m) anomalies for all $\leq -1$ seasonal values of the PNA.
Fig. 3.49. (a) DJF, (b) MAM, (c) JJA, and (d) SON mean 300-hPa geopotential heights (m) for all $\leq -1$ seasonal values of the PNA.
Fig. 3.50. Mean and anomaly composites for 300-hPa geopotential heights (m) (a,b), precipitable water (kg/m$^2$) (c,d), 850-hPa temperature (k) (e,f), respectively, are for selected negative PNA cases. Composites are for three days prior to the event.
Fig. 3.51. Mean and anomaly composites for 300-hPa geopotential heights (m) (a,b), precipitable water (kg/m$^2$) (c,d), 850-hPa temperature (k) (e,f), respectively, are for selected negative PNA cases. Composites are for during the event.
Fig. 3.52. Mean and anomaly composites for 300-hPa geopotential heights (m) (a,b), precipitable water (kg/m$^2$) (c,d), 850-hPa temperature (k) (e,f), respectively, are for selected negative PNA cases. Composites are for twelve days after the event.
Fig. 3.53. (a) DJF, (b) MAM, (c) JJA, and (d) SON ECC density for positive PNA – negative PNA.
4. Case Studies

4.1 3–23 March 2018 ECC event

The purpose of this chapter is to present two case studies of ECC events for the NH, one over the eastern United States and the North Atlantic, and the other over the western United States and the North Pacific. Analyses of the mean and anomalous 300-hPa geopotential height will be presented, and the results will be compared to the geopotential height fields observed during the large-scale teleconnection patterns previously discussed. These comparisons will provide forecasters with information that can help them recognize large-scale flow patterns that are correlated with ECC events, which have the ability to cause EWEs and widespread societal impacts.

4.1.1 Case overview

During 3–23 March 2018, five cyclones developed and tracked along the East Coast of the United States, leading to significant socioeconomic impacts. According to the National Oceanic and Atmospheric Administration’s (NOAA’s) National Centers for Environmental Information (NCEI), the 1 March 2018 and 18 March 2018 storms contributed to a total estimated cost of 3.6 billion dollars and nine deaths (https://www.ncdc.noaa.gov/billions/events/US/2018). The March ECC events led to a potent severe storm outbreak with over 20 tornadoes across Alabama, Florida, and Texas, as well as a powerful Nor’easter that impacted several Mid-Atlantic and Northeast states with extended periods of severe high winds, heavy snow, and extreme coastal erosion and flooding. The cyclones of March 2018 qualify as an ECC event due to the clustering of
the five cyclones along the east coast of the United States over a 21-day period ending 23 March 2018 (Fig. 4.1). Given the longevity of this ECC event, and the associated severe weather and socioeconomic impacts associated with this ECC event, the structure and evolution of the synoptic-flow pattern associated with this event will be briefly examined.

4.1.2 Synoptic evolution and comparison of ECC event to climatology

Figure 4.1 shows the ECC density distribution and frequency for the aforementioned 21-day period over eastern North America. The anomalous 300-hPa geopotential heights during the event will also be shown to compare to the composite analyses of the previously discussed PNA and NAO teleconnection patterns. These analyses were chosen to understand the large-scale flow pattern and jet structure that leads to this ECC event, as well as to determine which teleconnection patterns and their phases were most dominant during ECC event.

The 300-hPa height field exhibits positive height anomalies and a Rex block pattern across the North Pacific, as well as positive height anomalies southwest of Greenland (Fig. 4.2a). The 300-hPa height field also reveals negative height anomalies across western Europe, and eastern and western North America, which signifies a negative phase of both the PNA and NAO (see Fig. 3.48b and Fig. 3.34b). The 300-hPa height field implies that the North Pacific jet was shifted poleward and in a retracted phase with the extreme ridge amplification over the central North Pacific. This ridge amplification, associated with the negative phase of the PNA, likely plays an important role in the downstream development of a Rossby wave train across North America. This
Rossby wave train aids in the equatorward transport of cool, dry air, the formation of upper-level disturbances over the Pacific Northwest, and ridge building across central North America with the help of quasi-continuous ECs that advect warm moist air poleward across this region. Additionally, this ridge building across central North America plays a role in the equatorward transport of cool, dry air and upper-level disturbances over eastern North America. Furthermore, ridge amplification downstream from eastern North America off the southwestern tip of Greenland is associated with the negative phase of the NAO that characterizes high-latitude blocking patterns. Under these flow conditions, negative height anomalies are “locked-in” across eastern North America and western Europe (Fig. 4.2a and Fig. 4.2b).

Due to the longevity of this ECC event, a five-day analysis (T−48 h to T+72 h; 0000 UTC 10 March–0000 UTC 15 March 2018) of the synoptic pattern present during this ECC event is produced for the case study. The five ECs that compromised the ECC event of March 2018 will be identified as L1–L5 for the remainder of sections 4.1.2 and 4.1.3. T+0 h is 0000 UTC 12 March 2018, which coincides with the initial development of L3 observed during the 21-day period. This EC produces a secondary EC that forms on the trailing cold front of the initial EC, and tracks poleward to cluster with the initial EC over northern Maine. The initial EC also produced copious amounts of heavy snowfall across the Mid-Atlantic and Northeast coastlines and resulted in downed trees, power outages, coastal flooding, and several deaths. The ~0.7° horizontal resolution ERA-Interim reanalysis dataset (Dee et al. 2011) was utilized to illustrate the synoptic-
scale features of this event.

At T−48 h, a trough-ridge-trough pattern is evident across the North Pacific while a zonal flow pattern prevailed across the United States. A highly-amplified ridge over the Bering Straight and a deep trough across the Gulf of Alaska signifies a retracted NPJ, a negative PNA pattern, and indicates possible downstream development across eastern North America (Fig. 4.3a). At the same time, a strip of cyclonic vorticity is present across central Canada at 500 hPa that is associated with a weak short-wave trough located over Saskatchewan.

By T−24 h, the Rossby wave train across the North Pacific has amplified downstream of a region of ridge building associated with a cyclone east of the Kamchatka Peninsula. As this Rossby wave train progresses downstream, additional ridge building occurs across the central North Pacific, a trough deepens across the Gulf of Alaska, and a ridge amplifies over western North America (Fig. 4.3a). The development of the ridge over western North America occurs in conjunction with warm-air advection in advance of a developing surface low across the Gulf of Alaska. The ridge amplification across western North America enables the aforementioned strip of vorticity across central Canada to propagate equatorward to just northwest of the Great Lakes region. At the same time, high pressure is building westward over the southern tip of Greenland, which is consistent with a negative NAO pattern.

At T+0 h, the Rossby wave train across the North Pacific has remained in place as new surface cyclones have developed on the upstream sides of both ridges, further
amplifying these features through warm-air advection (Fig. 4.3a). At the same time, the development EC_L4 off the southeastern United States coast occurs in conjunction with cyclonic vorticity advection by the 1000–500-hPa thermal wind over the Tennessee Valley associated with the strip of cyclonic vorticity previously located over central Canada. A second region of shear vorticity strengthens over the Gulf of Mexico in conjunction with a tightening of the 1000–500-hPa thickness gradient.

By T+24 h, the merging of the two aforementioned regions of cyclonic vorticity has resulted in a significant deepening of the EC off the Northeast coast. The deepening EC occurs near a region of pronounced cyclonic vorticity advection by the 1000–500-hPa thermal wind favorable for deep ascent (Fig. 4.3b). At the same time, the northern region of cyclonic vorticity has become meridionally elongated, and extends from the northwestern Great Lakes poleward towards Hudson Bay, providing the possibility of a secondary development.

By T+48 h, a blocking high over the Canadian Maritimes develops in the left entrance region of the associated North Atlantic jet developing likely in response to subsidence across this region. This blocking high which is associated with a negative NAO pattern, prevents the EC off the coast of the northeastern United States from propagating northeastward, and inducing a downstream thermal ridge (Fig. 4.3b). A second cyclonic vorticity maximum, which was previously over the Great Lakes, has advanced equatorward off the coast of Georgia. It quickly rounds the base of the trough that is associated with the occluding EC off the northeastern coast of the United States,
and quickly develops a weak EC $L_5$ on the trailing frontal boundary associated with the occluding EC.

By T+72 h, cyclonic vorticity advection by the 1000–500-hPa thermal wind has aided in the development of a second EC on the trailing cold front of the occluding EC over northern Maine via a process originally described by (Bjerknes and Soldberg 1922). EC $L_5$ developed from the previously aforementioned cyclonic vorticity maximum that advanced equatorward off the coast of Georgia, propagated poleward over Newfoundland, and resulted in ridge development across the western North Atlantic (Fig. 4.3b). This developing ridge near the coast of southern Greenland helped to reinforce the existing negative NAO pattern.

The synoptic evolution of ECs $L_4$ and $L_5$ during the 3–23 March ECC event off the northeastern coast of the United States is illustrative of a negative NAO and PNA ECC event. This brief case study closely resembles the synoptic evolution surrounding the composites of the negative NAO and PNA patterns (see Fig. 3.48b and Fig. 3.34b). This case study is relatable to a joint negative NAO and negative PNA patterns in the following respects: At T−48 h, higher 1000–500-hPa thickness values were observed over the central North Pacific associated with an amplified high-pressure system consistent with a negative PNA pattern; lower thickness values were observed in the Gulf of Alaska associated with a deep trough, which was continually being reinforced by the development of ECs; and, finally, higher thickness and surface pressure values were observed over northeast Canada, which is consistent with a negative NAO pattern.
4.1.3 Dynamical evolution of the ECC event

The dynamical evolution to follow will utilize 1° GFS reanalysis plots created by Heather Archambault (http://www.atmos.albany.edu/student/heathera/). The 3–23 March 2018 case study was preceded by an anomalous ridging event across the eastern half of the United States (Figs. 4.4a and 4.4b). This antecedent flow pattern on 1200 UTC 22 February 2018 was associated with an amplified ridge across the Gulf of Alaska, and an anomalously strong 250-hPa midlatitude jet extending from New Mexico northeastward towards the Canadian Maritimes. An EC developing in the left exit region of the anomalously strong midlatitude jet will aid in the formation of a negative NAO pattern by causing ridging across the central North Atlantic, and a reversal of the flow pattern.

By 1200 UTC 26 February 2018, ridging is occurring across the Gulf of Alaska, while trough formation is happening across the western United States (Figs. 4.5a and 4.5b). Anticyclonic wave breaking is developing in conjunction with the trough formation across the western United States, leading to an enhanced subtropical jet across the southeastern United States (Figs. 4.5a and 4.5b). A high-pressure system in place over the Canadian Maritimes and the southern tip of Greenland is indicative of a negative NAO phase and blocking pattern, as flow reversal is evident east of Greenland.

The EC L1 develops around 1200 UTC 02 March 2018 in response to amplified ridging occurring in the Gulf of Alaska transitions downstream into a trough over the western United States and Canada (Figs. 4.6a and 4.6b). The EC L1 developing off the
Northeast Coast is developing in conjunction with a cyclonic wave-breaking event. A strong negative NAO pattern is present as evidenced by a blocking 1052-hPa high-pressure system over Greenland and the Canadian Maritimes.

EC L_2 develops around 1200 UTC 08 March 2018 along the Northeast Coast is developing in conjunction with a cyclonic wave breaking episode (Figs. 4.7a and 4.7b). Conversely, an anticyclonic wave-breaking episode across the Gulf of Alaska is occurring, and aiding in the reinforcement of the negative PNA pattern observed in the 300-hPa composite anomalies (see Fig. 4.2a). A 1032-hPa high-pressure system is supporting the negative NAO pattern poleward of EC L_2 over southern Greenland and the Canadian Maritimes.

The large-scale flow pattern observed on 1200 UTC 08 March 2018 aids in blocking the propagation of EC L_2, allowing for the development of EC L_3 along the trailing frontal boundary of EC L_2 during 1200 UTC 09 March 2018 (Figs. 4.8a and 4.8b). The cyclonic wave breaking that occurred in conjunction with EC L_2 is now occurring in conjunction with EC L_3. The EC L_2 and EC L_3 events help to reinforce downstream ridging near southern Greenland and contribute to the maintenance of the observed negative NAO and PNA patterns.

EC L_4 develops off the Mid-Atlantic Coast around 1200 UTC 13 March 2018 during an extremely amplified Rossby wave train extending from the Gulf of Alaska towards eastern North America (Figs. 4.9a and 4.9b). Cyclonic wave breaking occurring off the West Coast of the United States is juxtaposed in between two anticyclonic wave
breaking episodes occurring over the Gulf of Alaska and western North America (Figs. 4.9a and 4.9b). EC L₄ is developing in conjunction with a cyclonic wave-breaking episode off the East Coast of the United States. A blocking anticyclone poleward of EC L₄ is indicative of a continuing negative NAO pattern as are the easterlies observed off the southern tip of Greenland. The observed joint negative NAO and PNA pattern is conducive to maintaining the formation of ECs off the northeastern coast of the United States.

EC L₅ develops along the North Carolina coastline around 1200 UTC 20 March 2018 in conjunction with a less amplified Rossby wave train that extends from the Gulf of Alaska towards eastern North America (Figs. 4.10a and 4.10b). The aforementioned negative PNA pattern is still dominating the large-scale flow pattern across the North Pacific as is manifest by an anticyclonic wave-breaking episode occurring off the West Coast of the United States. However, the negative NAO pattern present since the end of February has transitioned into a positive NAO pattern in conjunction with cyclogenesis over the Canadian Maritimes (Figs. 4.10a and 4.10b). The transition to a positive NAO pattern will allow EC L₅ to propagate eastward, resulting in the end of the persistent blocking flow pattern associated with a negative NAO pattern by 1200 UTC 20 March 2018. The noteworthy ECC event over the eastern U.S. in March 2018 was made possible by anticyclonic wave breaking episodes over the Gulf of Alaska and off the West Coast of the United States, as well as cyclonic wave breaking episodes downstream over the eastern United States. This pattern resulted in the occurrence of sequential ECs
near the East Coast that contributed to and reinforced downstream blocking near Greenland and was manifest by a joint negative PNA and NAO pattern.

4.2 2–12 February 1998 ECC event

4.2.1 Case overview

During 2–12 February 1998, three cyclones tracked from the Gulf of Alaska towards the West Coast of the United States, leading to significant socioeconomic impacts. According to NOAA NCEI, California reported 550 million dollars in damages and 17 storm-related deaths from flooding, mudslides, and agricultural disruptions (https://www.ncdc.noaa.gov/billions/events/US/1998). The cyclones of February 1998 qualify as an ECC event due to the clustering of ECs L₁–L₄ along the West Coast of the United States over a ten-day period (Fig. 4.11). Given the longevity of the ECC event, and the severe weather and socioeconomic impacts associated with the event, an overview of this ECC event is presented.

4.2.2 Synoptic evolution and comparison of ECC event to climatology

Figure 4.11 shows the ECC density distribution and frequency for the 11-day period ending 12 February 1998. Fig. 4.12a shows the composite anomaly 300-hPa heights for the 11-day period of this case study to better understand the large-scale flow pattern that lead to this event. Negative 300-hPa height anomalies are evident across the central North Pacific and Gulf of Alaska, extending eastward across the southern United States. Positive height anomalies are evident across the northern United States and southern Canada, consistent with the positive phase of the ONI, indicative of an El Niño
event, and a positive PNA (see Fig. 3.14a and Fig. 3.42a). The 300-hPa anomalous height field implies an extended, and equatorward-shifted, NPJ across the North Pacific that manifests itself as a stronger subtropical jet across the eastern North Pacific and southern United States. The mean 300-hPa height field exhibits a quasi-wave four pattern across the NH, with the dominant feature being a very broad trough that extends across the North Pacific, which transitions into a diffluent pattern and highly-amplified ridge across the western United States where a string of ECs are recurving poleward in the Gulf of Alaska from their initial propagation eastward (Fig. 4.12b).

Due to the longevity of the event, a five-day analysis (T−48 h to T+72 h; 0000 UTC 2 February–0000 UTC 7 February 1998) of the synoptic pattern during this ECC event is produced. T+0 h is 0000 UTC 4 February 1998, which coincides with an EC occluding off the West Coast of the United States and a new EC about to develop to its west. At T−48 h, a zonal flow pattern exists across much of the North Pacific, except in the central North Pacific where an EC is building a weak thermal ridge (Fig. 4.13a). West of southern California, an EC is starting to develop downstream of a deep trough and strong cyclonic vorticity advection by the 1000–500-hPa thermal wind in association with an occluding cyclone in the northern Gulf of Alaska. At the same time, the development a ridge over western North America occurs in conjunction with warm-air advection in advance of the occluding and developing ECs across the Gulf of Alaska, indicative of the positive PNA pattern in the 300-hPa anomalous height field.

By T−24 h, the EC across the central North Pacific has propagated eastward to a
position equatorward of the Aleutian Islands as it occludes (Fig. 4.13a). The developing EC L₁ across the southern Gulf of Alaska also propagates eastward as it deepens in response to cyclonic vorticity advection by the 1000–500-hPa thermal wind. EC L₁, through warm-air advection, contributes to continued ridge building across the western United States. At the same time, an EC develops in response to cyclonic vorticity advection by the 1000–500-hPa thermal wind over the Gulf of Mexico on the cyclonic shear side of the subtropical jet. This flow pattern across the North Pacific and the United States is characteristic of both a positive PNA and ONI pattern.

At T+0 h, a new EC is developing across the southern tip of the Kamchatka Peninsula as strong cyclonic vorticity advection by the 1000–500-hPa thermal wind rapidly deepens the cyclone. Warm-air advection associated with this cyclone acts to reinforce a new downstream ridge in the central North Pacific southeast of the Kamchatka Peninsula (Fig. 4.13a). At the same time, the occluding EC and trough south of the Aleutian Islands has propagated eastward, and a new EC is developing on its trailing cold front and associated cyclonic vorticity in the Gulf of Alaska. EC L₁, which was impacting the west coast of the United States earlier, has occluded and is dissipating over California. The region of cyclonic vorticity associated with the now-occluded EC L₁ over California has advanced into the southwest United States. The EC currently over the southeastern United States has deepened from a merging of the aforementioned two regions of cyclonic vorticity over the southeast United States.

By T+24 h, the EC over the Kamchatka Peninsula has occluded and propagated
eastward toward the central North Pacific (Fig. 4.13b). At the same time, the cyclonic vorticity associated with the new EC that developed at T+0 h in the Gulf of Alaska has split into two distinct of cyclonic vorticity lobes: one which moves poleward and westward and remains with the initial EC, while the other advances eastward with the trailing cold front off the coast of Washington and Oregon. EC L₂ develops off the coast of Washington state in the region of cyclonic vorticity advection associated with the eastern lobe of vorticity that advanced eastward with the aforementioned trailing cold front. A strip of cyclonic shear vorticity extends from the Aleutian Islands equatorward and around the base of a broad trough that exists to the northeast of Hawaii. This shear vorticity maximum will help to generate the next EC.

At T+48 h, the occluded EC southeast of the Kamchatka Peninsula has propagated further eastward and cyclonic vorticity advection by the 1000–500-hPa thermal wind is occurring downstream of the occluded EC, which will help develop more ECs further downstream (Fig. 4.13b). At the same time, EC L₃ has developed off the West Coast of the United States from the elongated cyclonic shear vorticity maximum that extended from the Aleutian Islands equatorward and around the base of the trough that was located to the northeast of Hawaii at T+24 h. EC L₃ off the West Coast of the United States rapidly deepened due to strong cyclonic vorticity advection by the 1000–500-hPa thermal wind over the cyclone center in conjunction with the previously mentioned cyclonic shear vorticity maximum. Warm-air advection downstream of this EC contributed to downstream ridge building across western North America.
By T+72 h, cyclonic vorticity advection by the 1000–500-hPa thermal wind north of the Hawaiian Islands in the southern Gulf of Alaska occurring downstream of an occluded EC southeast of the Kamchatka Peninsula has propagated eastward. This cyclonic vorticity advection by the 1000–500-hPa thermal wind north of the Hawaiian Islands results in the development of EC L₄ off the west coast of the United States (Fig. 4.13b). EC L₃ previously off the West Coast of the United States, has propagated poleward to a position off the coast of southern Alaska. Continued warm-air advection ahead of this cyclone will contributes to downstream ridge building across western North America, and will enables downstream vorticity maxima to be displaced equatorward across the southeastern United States where they may merge with the vorticity associated with the strong subtropical jet across the Gulf of Mexico.

This February 1998 case study closely resembles the synoptic evolution associated with the composite analyses of the positive ONI and PNA (see Fig. 3.14a and Fig. 3.42a) in the following respects: lower thickness values observed across the central and eastern North Pacific are consistent with a number of ECs traversing the North Pacific, and are also consistent with the positive phases of the PNA and ONI, which both support a strong extended NPJ. The continuous EC events across the Gulf of Alaska and West Coast of the United States have sequentially reinforced the ridge across western North America. The associated cyclonic vorticity with these ECs has propagated across the southern United States to merge with the vorticity associated with the strong subtropical jet across the Gulf of Mexico. These cyclonic vorticity merging events help
to facilitate EC development along the Gulf of Mexico and southeastern United States. The case study provides key details of the synoptic pattern that evolved to produce copious amounts of rain across the western states of the United States as well as coastal lows for the southeast United States. The low-amplitude quasi-wavenumber four-trough pattern across the NH, which remained quasi-stationary during the ECC event produced numerous EC events that developed in conjunction with a strong extended NPJ provides a characteristic synoptic pattern associated with the positive phases of the PNA and ONI indices.
c. Figures

Fig. 4.1. ECC density distribution and frequency for the 3–23 March 2018 case study of the Northeast United States.
Fig. 4.2. (a) 300-hPa geopotential height anomalies (m) and (b) 300-hPa geopotential height mean field (m) composites for 3–23 March 2018.
Fig. 4.3. 1000–500-hPa thickness (dashed blue; every 6 dam, with the 540-dam contour bolded), 500-hPa absolute vorticity (shaded; beginning at $5 \times 10^{-5}$ s$^{-1}$ according to the color bar), and MSLP (solid black; every 5 hPa). The five-day period surrounding the onset of a section of the ECC event for the northeast United States, where (a) T−48 to T+0 (0000 UTC 10 March–0000 UTC 12 March 2018) and (b) T+24 h to T+72 h (0000 UTC 13 March–0000 15 March 2018).
Fig. 4.4. (a) 250-hPa wind (m s$^{-1}$, color shading), precipitable water (mm, grey shading), SLP (solid contours every 4 hPa), and 1000–500 thickness (dashed contours every 6 dam), and (b) DT (1.5-PVU surface) potential temperature (K, shaded), wind barbs (knots), 925–850-hPa layer averaged cyclonic relative vorticity (black contours, every 0.5 x 10$^{-4}$ s$^{-1}$), at 1200 UTC 22 February 2018. Plots created by Heather Archambault: (http://www.atmos.albany.edu/student/heathera/).
Fig. 4.5. (a) 250-hPa wind (m s$^{-1}$, color shading), precipitable water (mm, grey shading), SLP (solid contours every 4 hPa), and 1000–500 thickness (dashed contours every 6 dam), and (b) DT (1.5-PVU surface) potential temperature (K, shaded), wind barbs (knots), 925–850-hPa layer averaged cyclonic relative vorticity (black contours, every 0.5 x 10$^{-4}$ s$^{-1}$), at 1200 UTC 26 February 2018. Plots created by Heather Archambault: (http://www.atmos.albany.edu/student/heathera/).
Fig. 4.6. (a) 250-hPa wind (m s$^{-1}$, color shading), precipitable water (mm, grey shading), SLP (solid contours every 4 hPa), and 1000–500 thickness (dashed contours every 6 dam), and (b) DT (1.5-PVU surface) potential temperature (K, shaded), wind barbs (knots), 925–850-hPa layer averaged cyclonic relative vorticity (black contours, every 0.5 x 10$^{-4}$ s$^{-1}$), at 1200 UTC 02 March 2018. Plots created by Heather Archambault: (http://www.atmos.albany.edu/student/heathera/).
Fig. 4.7. (a) 250-hPa wind (m s$^{-1}$, color shading), precipitable water (mm, grey shading), SLP (solid contours every 4 hPa), and 1000–500 thickness (dashed contours every 6 dam), and (b) DT (1.5-PVU surface) potential temperature (K, shaded), wind barbs (knots), 925–850-hPa layer averaged cyclonic relative vorticity (black contours, every 0.5 x 10$^{-4}$ s$^{-1}$), at 1200 UTC 08 March 2018. Plots created by Heather Archambault: (http://www.atmos.albany.edu/student/heathera/).
Fig. 4.8. (a) 250-hPa wind (m s$^{-1}$, color shading), precipitable water (mm, grey shading), SLP (solid contours every 4 hPa), and 1000–500 thickness (dashed contours every 6 dam), and (b) DT (1.5-PVU surface) potential temperature (K, shaded), wind barbs (knots), 925–850-hPa layer averaged cyclonic relative vorticity (black contours, every 0.5 x 10$^{-4}$ s$^{-1}$), at 1200 UTC 09 March 2018. Plots created by Heather Archambault: (http://www.atmos.albany.edu/student/heathera/).
Fig. 4.9. (a) 250-hPa wind (m s$^{-1}$, color shading), precipitable water (mm, grey shading), SLP (solid contours every 4 hPa), and 1000–500 thickness (dashed contours every 6 dam), and (b) DT (1.5-PVU surface) potential temperature (K, shaded), wind barbs (knots), 925–850-hPa layer averaged cyclonic relative vorticity (black contours, every 0.5 x 10$^{-4}$ s$^{-1}$), at 1200 UTC 13 March 2018. Plots created by Heather Archambault: (http://www.atmos.albany.edu/student/heathera/).
Fig. 4.10. (a) 250-hPa wind (m s$^{-1}$, color shading), precipitable water (mm, grey shading), SLP (solid contours every 4 hPa), and 1000–500 thickness (dashed contours every 6 dam), and (b) DT (1.5-PVU surface) potential temperature (K, shaded), wind barbs (knots), 925–850-hPa layer averaged cyclonic relative vorticity (black contours, every 0.5 x 10$^{-4}$ s$^{-1}$), at 1200 UTC 20 March 2018. Plots created by Heather Archambault: [http://www.atmos.albany.edu/student/heathera/](http://www.atmos.albany.edu/student/heathera/).
Fig. 4.11 ECC density distribution and frequency for the 2–12 February 1998 case study of the West Coast of the United States.
Fig. 4.12. (a) 300-hPa geopotential height anomalies (m) and (b) 300-hPa geopotential height mean field composites (m) for 2–12 February 1998.
Fig. 4.13. 1000–500-hPa thickness (dashed blue; every 6 dam, with the 540-dam contour boulded), 500-hPa absolute vorticity (shaded; beginning at $5 \times 10^{-5} \text{s}^{-1}$ according to the color bar), and MSLP (solid black; every 5 hPa). The five-day period surrounding the onset of a section of the ECC event for the western United States, where (a) T−48 to T+0 (0000 UTC 2 February–0000 UTC 7 February 1998) and (b) T+24 h to T+72 h (0000 UTC 5 February–0000 7 February 1998).
5. Discussion, conclusions, and suggestions for future work

5.1 Discussion and conclusions

The goals of this research were to improve the understanding of the linkages between preferred spatiotemporal extratropical cyclone clustering (ECC) and persistent large-scale flow regimes (i.e., teleconnection patterns). Although previous studies have examined the serial clustering of ECs in regions such as the North Atlantic, no previous study has investigated ECs that cluster in a spatiotemporal distribution around the entire NH, nor analyzed those ECCs from a climatological perspective during large-scale flow regimes. Climatologies of spatiotemporal ECC events during $\geq 1$ and $\leq -1$ index values of large-scale flow regimes such as the NAO and PNA were constructed to gain insight on regions favorable for ECCs to develop during the aforementioned large-scale flow patterns. Mean and anomaly composites of selected dynamic and thermodynamic variables (e.g., precipitable water, 300-hPa geopotential heights, and 850-hPa temperatures) were constructed to better understand linkages between ECCs and large-scale flow regimes. These composites were used to identify preferred large-scale flow patterns and characteristics associated with ECCs for selected regions of the Northern Hemisphere. Case study investigations of two representative ECC events that were linked to large-scale flow patterns were performed to gain further understanding of the synoptic evolution and dynamic linkages during the ECC events.

5.1.1 Climatologies

5.1.1.1 ECCs and the ONI

Many of the results of the ECC climatology are in agreement with past studies of
ECs and their preferred genesis and tracking regions during both the positive and negative phase of the Oceanic Niño Index (ONI) and their associated NH large-scale flow patterns. Storm tracks during the positive phase of ENSO (i.e., El Niño) are typically characterized by an equatorward shift in the storm track in the North Pacific and an enhanced and equatorward-shifted North American East Coast storm track (e.g., Hoerling and Kumar 1997; Hoerling and Kumar 2000; Higgins et al. 2002; Eichler and Higgins 2006). Storm tracks during the negative phase of ENSO (i.e., La Niña) are typically characterized by a poleward shift in the storm tracks in both the North Pacific and North American East Coast (e.g., Hoerling and Kumar 1997; Hoerling and Kumar 2000; Higgins et al. 2002; Eichler and Higgins 2006). ECCs preferentially occur in the Gulf of Alaska, north central Russia, and along an equatorward-shifted storm track across the North Atlantic during the positive phase of the ONI. Likewise, ECCs and their associated large-scale height patterns preferentially occur over the north-central Pacific, east of Greenland, over eastern Russia, and along a poleward-shifted storm track across the North Atlantic during the negative phase of ONI. Winters et al. (2018) have shown that a positive phase of the ONI aids in modifying the configuration of the NPJ through anomalous convection and above-normal sea-surface temperatures over the central and eastern equatorial Pacific by favoring an extended and equatorward-shifted NPJ. Conversely, as shown by Winters et al. (2018), a negative phase of the ONI produces anomalous convection and above-normal sea-surface temperatures over the western equatorial Pacific favoring a retracted NPJ. An extended and retracted NPJ favors a Gulf of Alaska ECC maximum and a north-central Pacific ECC maximum, respectively, in this
study.

5.1.1.2 ECCs and the NAO

Similar to the results for the ONI, results of the NAO ECC climatology are in agreement with past studies of ECs and their preferred genesis and tracking regions during both the positive and negative phases of the NAO, and their associated NH large-scale flow patterns. The positive phase of the NAO is typically characterized by a strong Icelandic low and Azores high, or negative height anomalies near Iceland and positive height anomalies near the Azores. This phase is established with an anomalously strong southwest–northeast-positioned jet stream and storm track that extends from near the New England coast to northern Europe. The negative NAO phase is typically characterized by a weak Icelandic low and Azores high with the corresponding positive and negative height anomalies, respectively. This positive NAO phase is established with a weak west–east-positioned jet stream and storm track that extends from the Mid-Atlantic coastline to central Europe (Lau 1988; Rogers 1990; Archambault et al. 2010).

During the positive phase of the NAO, ECCs preferentially occur over Iceland, east of Greenland, as well as along a poleward-shifted North Atlantic storm track. Conversely, during the negative phase of the NAO, ECCs and their associated large-scale height patterns preferentially occur over the British Isles, off the coasts of Newfoundland and Portugal, as well as along an equatorward-shifted storm track across the North Atlantic as shown by Wallace and Gutzler (1981), Hurrell (1995), and Archambault et al. (2010). During the positive phase of the NAO, an anomalously strong southwest–northeast oriented North Atlantic jet stream and storm track extends from the Mid-Atlantic coast to
central Europe as was also shown in previous studies (Wallace and Gutzler 1981; Hurrell 1995; Archambault et al. 2010). Conversely, during the negative phase of the NAO, an anomalous weak North Atlantic jet stream is expected, and storm track extending from the Mid-Atlantic coast to the Iberian Peninsula was observed in the ECC climatology. This negative NAO storm track pattern is consistent with the ECC climatology and anomalous 300-hPa height field composites constructed.

5.1.1.3 ECCs and the PNA

Similar to the results for the ONI and NAO, results of the PNA ECC climatology are in agreement with past studies of ECs and their preferred genesis and tracking regions during both the positive and negative phases of the PNA, and their associated NH large-scale flow pattern. The positive phase of the PNA is characterized by a ridge–trough pattern over North America, where a ridge is observed over western North America and a trough over eastern North America (e.g., Leather et al. 1991; Archambault et al. 2010). Conversely, the negative phase of the PNA is characterized by the opposite ridge–trough pattern, with a trough observed over western North America and a ridge over eastern North America (e.g., Leather et al. 1991; Archambault et al. 2010). During the positive phase of the PNA, ECCs preferentially occur across the southeastern United States extending from the Gulf of Mexico towards Greenland, the Gulf of Alaska, and an equatorward-shifted storm track and ECC distribution across the North Atlantic is present. Conversely, during the negative phase of the PNA, ECCs preferentially occur across eastern North America, over western North America and the Kamchatka Peninsula, and a poleward-shifted storm track and ECC distribution is present across the North Atlantic as shown by (e.g., Wallace and Gutzler 1981; Barnston and Livezey 1987;
Archambault et al. 2010). As previously discussed, the occurrence and location of ECC events are clearly linked to the structure, evolution, and location of the NPJ. During the positive phase of the PNA, an anomalous trough is observed across the central North Pacific, and an anomalous ridge is observed across the subtropical regions of the central North Pacific. As shown by numerous authors, (Wallace and Gutzler 1981; Barnston and Livesey 1987; Franzke and Feldstein 2005; Strong and Davis 2008; Athanasiadis et al. 2010; Franzke et al. 2011; Griffin and Martin 2017; Winters et al. 2018), this positive phase of the PNA favors an extended NPJ and is consistent with the Gulf of Alaska ECC maximum observed in this study. Conversely, during the negative phase of the PNA, an anomalous ridge is present across the central North Pacific and an anomalous trough is located across the subtropical regions of the central North Pacific and western North America. As shown by previous studies, this pattern favors a retracted NPJ and a western North America ECC maximum as observed in this study, patterns that are consistent with the ECC climatology and the anomalous 300-hPa height field composites constructed during this study. The wavenumber-three NH flow pattern observed during a negative PNA pattern in the composite mean 300-hPa height field may be associated with a more stable NH and persistent flow regime as compared to the wavenumber-four positive PNA pattern which may be more transient due to its more amplified state. The 300-hPa composite anomalies of the negative PNA and NAO patterns revealed that these phases of the PNA and NAO can occur simultaneously, which teleconnect with a greater frequency of ECC events.

5.1.2 Case studies

The case study results presented in Chapter 4 established that ECC events can be
linked dynamically to the large-scale flow patterns through observations of the
amplitudes of Rossby wave trains, positions of cyclonic and anticyclonic wave-breaking
episodes, and to the position and structure of the midlatitude jet stream. It is evident that
quasi-stationary North Atlantic and North Pacific jets play a necessary role in the
production and maintenance of ECC events over specific regions across the NH by
continually forming ECs in the left-exit or right-entrance regions of these jets. It is also
evident that the ECs that develop during these ECC events aid in maintaining the large-
scale flow pattern by continually tracking over the same regions to reinforce the ridge
downstream of the ECC event and maintain the current large-scale flow pattern and
midlatitude jet structure. This process was clearly apparent in the 2–12 February 1998
ECC event, where numerous ECs developed off the West Coast of the United States from
the trailing cold fronts associated with occluding cyclones from the central North Pacific.
These ECs progressed poleward as they occluded and advected copious amounts of heat
and moisture poleward to reinforce the ridge over western North America. ECC events,
such as described in the March 2018 case study, may develop from anomalously strong
EC events that aid in perturbing the large-scale flow pattern and inducing an associated
downstream Rossby wave train. For example, the March 2018 ECC event was initiated
by an anomalously strong midlatitude jet and ridging pattern over the eastern United
States. This anomalously strong jet aided in the development of a strong EC in its left-
exit region of the North Atlantic jet. The strong EC recurved poleward and westward
over Greenland, and assisted in the development of a negative NAO pattern in
conjunction with high-latitude ridge building over the central and eastern North Atlantic.
This ridging created a reversal in the upper-level flow pattern from westerly to easterly
over Greenland in conjunction with the establishment of atmospheric blocking across eastern North America. This blocking pattern prevented disturbances from propagating downstream and also aided in shifting the storm track equatorward from the Canadian Maritimes to the Northeast and Mid-Atlantic Coasts of the United States. The blocking-related development of a negative NAO pattern in tandem with a negative PNA pattern allowed for the enhancement of the ECC event by developing a large-scale flow pattern that was conducive for ECs to occur over the eastern United States.

These case studies highlighted that the negative phases of the NAO and PNA, and the positive phases of the ONI and PNA, can occur simultaneously, which aids in a greater frequency of ECC events. This situation is evident in both the 2–12 February 1998 and the 3–23 March 2018 ECC events, where the 2–12 February 1998 ECC event closely resembles the synoptic flow pattern associated with the composite analyses of the positive phases of the ONI and PNA, and the 3–23 March 2018 ECC event closely resembles the synoptic flow pattern associated with the composite analyses of the negative phases of the PNA and NAO. Therefore, it is likely that these ECs and their associated ECC events aid in the development of these quasi-stationary, large-scale flow patterns that occur on sub-seasonal timescales. The development of these quasi-stationary large-scale flow patterns from ECC events are likely caused from recurring downstream ridge building associated with warm-air advection and strong diabatic heating from repetitive ECs tracking over the same areas over a short period of time. These ECC events were found to be initiated two–four weeks prior to the ECC events by anomalous disturbances such as strong midlatitude disturbances or anomalously strong ECs that interacted with the midlatitude jet stream to cause atmospheric blocking. The
reversal of the flow field and blocking pattern allowed for consecutive cyclonic wave breaking events to develop in conjunction with EC events over the same regions.

5.2 Suggestions for future work

The results presented in this thesis motivate a number of future research opportunities. The methodology used to construct ECC events and ECC frequency, and their distribution on seasonal and yearly time scales could be modified in two ways: 1) expand the radius around each grid-point used to cluster ECs to either 600-km or 700-km to determine whether changing the radius size will affect ECC frequency and distribution, and 2) filter out ECs above a certain MSLP threshold. Modifying the methodology by filtering ECs by certain MSLP threshold may provide an improved understanding of the scale and impact that ECC events have on large-scale flow patterns on a sub-seasonal timescale. Further additional case studies should be performed in locations over the central North Pacific, northern Siberia, and the Arctic during the summer season, the northwestern Europe and the Mongolian Plateau, where initial findings suggested that these regions is a convergence zones of ECs that develop in the lee of the Himalaya Mountain range, and in the lee of the Ural Mountain range that track eastward through Siberia.

REFERENCES


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