Development of Anomalous Temperature Regimes over the Southeastern United States: Synoptic Behavior and Role of Low-Frequency Modes

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ABSTRACT

During winter, anomalous temperature regimes (ATRs), which include cold-air outbreaks (CAOs) and warm waves (WWs), have important impacts in the southeastern United States. This study provides a synoptic–dynamic characterization of ATRs in the southeastern United States from 1949 to 2011 through composite time-evolution analyses. Events are categorized by the sign and amplitude of relevant low-frequency modes. During CAO (WW) onset, negative (positive) geopotential height anomalies are observed in the upper troposphere over the Southeast with oppositely signed anomalies in the lower troposphere over the central United States. In most cases, there is a surface east–west geopotential height anomaly dipole, with anomalous northerly (CAO) or southerly (WW) flow into the Southeast leading to cold or warm surface air temperature anomalies, respectively. Companion potential vorticity anomaly analyses reveal prominent features in the mid- to upper troposphere consistent with the coincident geopotential height anomaly patterns. Ultimately, synoptic-scale disturbances are found to serve as dynamic triggers for ATR events, while low-frequency modes provide a favorable environment for ATR onset. The results provide a qualitative indication of the role of low-frequency modes in ATR onset. In WW (CAO) events influenced by low-frequency modes, the North American geopotential height anomaly pattern arises in part as a downstream (regional) manifestation of the negative Pacific–North American pattern (North Atlantic Oscillation). Interestingly, the North Atlantic Oscillation contributes to both CAO onset and demise. Thus, these results indicate that low-frequency modes also affect event duration (CAOs). One general distinction found for ATRs is that CAOs involve substantial airmass transport while WW formation is more regional in nature.

1. Introduction

Anomalous temperature regimes (ATRs) occurring during the boreal cool season [December–January (DJF)], including cold-air outbreaks (CAOs) and warm waves (WWs), are common phenomena over eastern North America (Konrad and Colucci 1989), as well as other areas of the world (Ashcroft et al. 2009; Schultz et al. 1997, 1998). ATRs are generally defined as events where temperatures are significantly above (WWs) or below (CAOs) average on a synoptic time scale (2–7 days). These events can have major impacts on energy consumption, local agriculture, and human health (Westby et al. 2013). In response to anthropogenic influences on regional climate, temperature extremes are expected to change in severity, frequency, and duration in the future (Loikith and Broccoli 2012). Even against the general backdrop of globally warming temperatures, past studies find that midlatitude CAO events can be expected to at least persist (and in some cases increase) in intensity and duration throughout the twenty-first century based on inferences from historical observed trends (Walsh et al. 2001; Portis et al. 2006; Westby et al. 2013) coupled with modeling studies (Scherer and Diffenbaugh 2014; Kodra et al. 2011; Park et al. 2011). While some studies find that CAO frequency will likely decrease in the future (Kodra et al. 2011), other studies predict that CAOs may even increase in frequency in some regions (Vavrus et al. 2006). These opposing predictions can be linked to regional climate changes associated with forced changes in the large-scale circulation or extreme synoptic events (Vavrus et al. 2006; Francis and Vavrus 2012; Loikith and Broccoli 2014). Another consideration is that the mean and variability are distinct statistical moments (Park et al. 2011) that may vary differently in relation to climate change. For example, regional temperature variability in a future warmer climate may be similar to that in the present
colder climate and hence the behavior of anomalous temperature events may not vary appreciably. This also has important potential implications for agriculture in early spring. Freeze events depend more on the variability of temperature than the overall temperature trends, which can increase the risk of plant damage. This is especially true in regions where recent climate change includes warmer late winter to early spring air temperatures and advanced plant blooming, which may make the plants more vulnerable if CAOs occur during these developmental stages (Augspurger 2013).

On the other hand, WWs are generally expected to increase in frequency and severity in the future (Scherer and Diffenbaugh 2014). This trend is attributed to an increase in the odds of warm years and seasons due to anthropogenic influences (Christidis and Stott 2014) and globally warming temperatures. But unlike for CAOs, studies of boreal winter WWs are scarce in the literature. Nonetheless, these events can also have important implications for agriculture. Many plants use warming episodes as cues to begin the process of budding, leafing, and flowering (Peterson and Abatzoglou 2014), and WWs can pose a threat of increased freeze damage by inducing premature plant development (Gu et al. 2008) followed by a return to normal winter conditions. Freeze damage can also occur if a late winter WW event is followed by a prominent CAO. Such sequences have been responsible for damages to the agriculture industry ranging from $500 million to $2 billion (Peterson and Abatzoglou 2014).

The important regional impacts of boreal winter CAOs and WWs underscore the necessity for regional preparedness in anticipation of future changes in their behavior (Kodra et al. 2011; Park et al. 2011; Scherer and Diffenbaugh 2014). To better understand these changes in ATRs, it is necessary to have a stronger understanding of the synoptic and physical nature of ATR events in the current climate, including the role of low-frequency modes. An enhanced understanding of ATRs will allow us to more skillfully predict ATRs, adequately assess the ability of climate models in representing ATR events, and better characterize the uncertainties of future projections in ATR behavior.

Several prior studies have examined the typical synoptic behavior of CAOs, while parallel studies of boreal winter WWs are rare in the literature. Past studies have identified several favorable synoptic conditions for CAOs occurring east of the Rocky Mountains: a 500-hPa dipole pattern with ridging (trenching) over the western (eastern) United States (Hartjenstein and Bleck 1991; Colle and Mass 1995), high-latitude ridging at 500 hPa (Konrad 1998), a corresponding surface couplet with an anticyclone (cyclone) over the western (eastern) United States (Konrad and Colucci 1989; Colle and Mass 1995; Konrad 1998; Ashcroft et al. 2009), cold-air advection across 55°N (Hartjenstein and Bleck 1991), and transient waves (Konrad 1998). This upper-level ridge–trough pattern and surface couplet controls the strength of the meridional circulation, promoting strong northerly flow that transports cold air from high-latitude source regions, ultimately leading to CAO onset. However, a few studies have found that, for the southeastern (SE) United States, the anticyclone strength plays a more important role than the accompanying cyclone in CAO onset (Konrad and Colucci 1989; Konrad 1996, 1998). This is especially true when cyclogenesis occurs after the initiation of the CAO event, and has been found to be associated with the strongest CAO events (Konrad and Colucci 1989; Konrad 1996). The timing of the surface cyclogenesis relative to event onset also affects the relative contributions of cold-air advection and adiabatic processes during event onset (Konrad and Colucci 1989). On the other hand, very few studies on the synoptic behavior of winter WW events exist. A study by Teng et al. (2013) found that heat waves are typically preceded by a wavenumber-5 quasi-stationary wave pattern. Even though the study mainly focuses on summer heat waves, they remark that this wave pattern is observed in both winter and summer and thus may account for heat waves in both seasons. In principle, this type of circulation could provide the necessary warm-air advection into the region of interest characteristic of WW events. It also makes sense to expect that the anomaly patterns associated with warm events are of opposite sign to, but structurally similar to, those associated with CAOs (Loikith and Broccoli 2012). In the accounts presented thus far, it is clear that synoptic-scale circulations are critical to ATR development; however, it is also possible that these smaller-scale circulations may be embedded within larger planetary-scale circulation patterns that may persist for much longer (Konrad 1998). Therefore, circulation on a variety of scales may contribute to the formation of ATRs, ranging from frontal-scale features up to hemispheric low-frequency modes.

Indeed, prior statistical research has identified linkages between ATRs and the amplitude and phase of several low-frequency modes of variability. The most extreme temperatures tend to occur when the index value of at least one low-frequency mode is in the upper or lower quartile of the distribution (Loikith and Broccoli 2014). Specifically, CAOs have been linked to the positive phase of the Pacific–North American (PNA) teleconnection pattern (Downton and Miller 1993; Vavrus et al. 2006; Cellitti et al. 2006; Rogers and Rohli 1991; Westby et al. 2013; Loikith and Broccoli 2014), and the
negative phases of the North Atlantic Oscillation (NAO) (Walsh et al. 2001; Cellitti et al. 2006; Westby et al. 2013), Arctic Oscillation (AO) (Lim and Schubert 2011), and northern annular mode (NAM) (Loikith and Broccoli 2014) (noting that the AO and NAM represent the same phenomenon). Connections between CAOs and the positive phases of the Pacific decadal oscillation (PDO) (Westby et al. 2013) and El Niño–Southern Oscillation (ENSO) (Lim and Schubert 2011; Westby et al. 2013; Loikith and Broccoli 2014) have also been identified. Meanwhile, WW frequency is modulated by the positive phase of the NAO and the negative phases of the PNA, PDO, and ENSO (Westby et al. 2013). However, some of these low-frequency modes seem to play a more important role than others. For instance, Loikith and Broccoli (2014) find that the PNA and NAM modes are more important than ENSO in modulating extreme temperatures over North America, and that the PNA is slightly more dominant than NAM. Westby et al. (2013) also find a weaker relationship between ENSO and ATRs compared to the PNA or AO. This is likely because the primary atmospheric circulation anomalies associated with these modes of variability are more prominent over the continent and thus provide a stronger direct influence on temperatures in these regions. Low-frequency modes also play a more prominent role in modulating ATRs in particular regions (e.g., the Southeast) of the United States compared to others, and the simultaneous contribution of multiple modes may also be important (Loikith and Broccoli 2014; Westby et al. 2013). For example, over the SE region, nearly 30% of the variability in CAOs can be attributed to the NAO, while over 50% of the variability in WWs can be attributed to the combined influence of the NAO and PNA (Westby et al. 2013).

Although there is a significant relationship between ATRs and the phase and magnitude of several low-frequency modes, this is an imperfect relationship since the modes only explain a fraction of the interannual variability in ATRs (Westby et al. 2013). Therefore, other factors are required to account for the remaining variability. Loikith and Broccoli (2014) found that local, transient, synoptic-scale weather features represent another important factor for ATRs in many locations, and they often occur in conjunction with an unusually strong phase of an influential low-frequency mode. For example, a CAO case study by Bosart et al. (1996) found that a 1993 CAO event resulted from an interaction between the slowly evolving planetary-scale circulation (perhaps associated with a teleconnection pattern) and a rapidly moving smaller-scale transient feature. Thus, there is not a one-to-one relationship between ATRs and low-frequency modes. Other factors, such as synoptic-scale disturbances (Bosart et al. 1996) or internal variability (Teng et al. 2013), likely come into play and may interact with the low-frequency modes through a positive feedback process that leads to the formation of ATRs. Given a collective consideration of these studies, it seems possible that the large-scale modes may be important in helping to form the predecessor cold or warm air masses, while the synoptic features provide a physical trigger to transport these air masses elsewhere and hence lead to ATR formation.

This current study applies categorical composite analyses to identify the primary synoptic and dynamic features associated with historical ATR onset over the SE United States. The objective of our study is to provide a structured, comprehensive, and updated analysis of the typical nature of SE ATRs emphasizing their linkage to prominent low-frequency modes. Many studies thus far have focused on the statistical characteristics of ATRs [Walsh et al. (2001), Portis et al. (2006), and Westby et al. (2013), among many others], examined the synoptic nature of individual case studies (Bosart et al. 1996; Schultz et al. 1997; Colucci et al. 1999; Hartenstein and Bleck 1991; Konrad and Colucci 1989), or have performed composite analyses without explicit consideration of the roles of low-frequency modes (Portis et al. 2006; Dallavalle and Bosart 1975; Konrad 1996; Colle and Mass 1995; Loikith and Broccoli 2012). Further, as discussed above, there is little or no existing research on the behavior of boreal cool season WWs. The current research aims to fill these gaps.

Given the prominent role of low-frequency modes in modulating SE ATRs (Westby et al. 2013), we choose to focus our attention on CAOs and WWs occurring in the SE United States in the current study. In our approach, we will first categorize events in terms of the phase and amplitude of the primary implicated low-frequency modes, allowing for an explicit delineation of their physical role. We will then study the composite time evolution of the temperature and circulation anomaly patterns leading up to and during ATR onset. These analyses aim to identify key atmospheric features associated with high-impact ATR events to aid local forecasters with precursor pattern recognition and improve the accuracy of predictions of ATRs over the region of interest. We describe our data and methodology in section 2. In section 3, we present the results of our composite analyses and describe the key atmospheric circulation features associated with each category of ATR events. The final section summarizes our findings, discusses their implications, and provides concluding remarks.

2. Data and methods

The main dataset for our study is the four-times-daily National Centers for Environmental Prediction–National
Center for Atmospheric Research reanalyses (NNR; Kalnay et al. 1996) for the period from January 1948 to February 2011. This dataset is advantageous because its length allows for a relatively large sample of events. ATR events are identified in terms of surface air temperature (SAT) at the $\sigma = 0.995$ level, the closest level to the surface. Other NNR fields used in our analysis are 1000-hPa zonal and meridional winds, 1000- and 2500-hPa geopotential heights, sea level pressure (SLP), and the 1000–500-hPa thickness field. The horizontal resolution of the gridded data is 2.5° latitude $\times$ 2.5° longitude. The usage of the four-times-daily instantaneous data (versus daily means) helps facilitate a better understanding of the detailed synoptic conditions leading to ATRs since smaller-scale features, such as fronts, become more readily apparent.

Besides characterizing the synoptic behavior of ATR development, it is of interest to identify significant dynamical features, which can be succinctly described using potential vorticity (PV). Although PV is not readily available on pressure surfaces from NNR, it is easily obtained from the European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA; the ERA data used in this study have been obtained from the ECMWF data server; http://apps.ecmwf.int/datasets/). The dataset comprises the 40-yr ECMWF Re-Analysis (ERA-40; which spans from 1957 to 2002) and the ECMWF interim reanalysis (ERA-Interim; extending from 1979 to the present), which have different horizontal spatial resolutions (2.5° latitude $\times$ 2.5° longitude and 1.5° latitude $\times$ 1.5° longitude, respectively). To make use of the ERA data for the same time span as NNR, we tested whether it was possible to “merge” these two datasets to create a longer dataset, via spatial interpolation of the ERA-Interim data to be comparable with ERA-40. We first verified that the ERA-40 and ERA-Interim datasets similarly represent large-scale variability in the field variables (e.g., SAT and 500-hPa geopotential height) in the region of interest. To achieve this, we conducted spatial correlation analyses between the two datasets for the period of overlap (1980–2002) over a domain encompassing North America (30°–70°N, 230°–300°E), including the airmass source regions for CAOs and WWs. The ERA-Interim data were interpolated onto the ERA-40 grid using bilinear interpolation. The spatial correlation values for SAT and 500-hPa geopotential heights, performed at daily, monthly, and yearly time scales, were all at or above 0.999, suggesting excellent quantitative correspondence. Therefore, we ascertain that the ERA-40 and ERA-Interim datasets can be usefully merged into one dataset (denoted ERA) to provide a supplementary long-term dataset to NNR. We then create a PV dataset that extends from September 1957 to February 2011 with a grid resolution of 2.5° latitude $\times$ 2.5° longitude. We analyzed PV at 12 levels ranging from 925 to 100 hPa.

The low-frequency modes of interest are the NAO and the PNA teleconnection pattern, recognized as leading factors in SE ATR modulation (e.g., Westby et al. 2013). The monthly low-frequency mode indices used in this investigation are from the National Oceanic and Atmospheric Administration’s (NOAA) Climate Prediction Center (CPC) (http://www.esrl.noaa.gov/psd/data/climateindices/).

To identify ATR events during winter (December–February), we first calculate detrended surface air temperature anomalies from NNR following the approach described in Westby et al. (2013). However, in the current study this method is applied to SAT averaged over a specific region instead of at individual grid points. The region of interest for this study is the SE United States, as defined in Fig. 6 in Westby et al. (2013), because ATRs in this region exhibit the strongest connection to low-frequency modes. Once the detrended SAT anomalies were calculated for the period 1948–2011 for this region, discrete ATR events were identified using magnitude, duration, and separation criteria. The first requirement is that each warm wave (cold-air outbreak) consists of times when the SAT anomalies exceed (fall below) $+1\sigma$ ($-1\sigma$). Using this threshold ensures an adequate sample size for statistical significance testing. Second, each event is required to be at least 5 days long. The duration requirement of 5 days allows us to study relatively high-impact events. Third, to ensure independence, events in each category are required to be separated from one another by at least 10 days.

Composite analyses are often used to identify fundamental synoptic-scale circulation signatures and precursor features unique to specific classes of extreme events (Chen and Zhai 2014; Sisson and Gyakum 2004; Grotjahn and Faure 2008; Milrad et al. 2009) and are used extensively in the current analysis. The composites were created by first categorizing WW or CAO events by the sign and magnitude of the monthly low-frequency mode index for the month of occurrence. Following the results of Westby et al. (2013), where it was determined that the NAO and PNA both strongly modulate ATR frequency in the Southeast (their Fig. 4), ATR events are classified based on the sign and magnitude of these two low-frequency modes. The positive (negative) event category is defined as when the normalized low-frequency mode index is $>+1\sigma$ ($<-1\sigma$), while the neutral event category is when $-1\sigma <$ normalized low-frequency mode index $< +1\sigma$. The number of events in each category, as well as the average length of events, is provided in Table 1. Interestingly, the mean duration of events occurring during negative phases of the low-frequency modes is longer than the other categories.
Because of the infrequent occurrence of ATRs during the positive phases of the low-frequency modes, those categories are not considered further. Thus, our composite analyses focus on the following four most commonly populated categories:

1) SE CAOs, neutral NAO, and varying PNA;
2) SE CAOs, negative NAO, and varying PNA;
3) SE WWs, neutral PNA, and varying NAO; and
4) SE WWs, negative PNA, and varying NAO,

where “varying” indicates no categorization in terms of this index. The inclusion of neutral cases allows us to directly compare and contrast with the low-frequency mode cases, in order to isolate the distinct roles of the low-frequency modes. The individual events included in each category are listed in Table 2.

Time-evolving composite fields are constructed for each of the four categories outlined above in order to isolate the synoptic and dynamic structural distinctions among the different categories of ATR events. For the composite anomaly calculations, the background 6-hourly climatological value of each parameter is calculated for the period 1948–2011 (NNR) or 1958–2011 (ERA). The time evolution of these composite fields are analyzed relative to the onset day (day 0, when the anomaly threshold value is first crossed) with day \( n \) referring to the \( n \)th day prior to (negative value) or after (positive value) ATR onset. We construct composites from day \(-8\) to \(+4\) at 6-hourly intervals, although for space considerations the results are presented at 2-day intervals. We display results for specific days in a temporal window encompassing the period characterized by the strongest statistically significant features (which varies from case to case). Statistical significance is assessed using a two-tailed Student’s \( t \) test with a confidence level of 95\%. Statistical significance is displayed as color shading, green contours, or boldface black contours, and is detailed in the figure captions. Nevertheless, we note that a statistically significant feature in the composite may not occur in every ATR event the composite comprises (Grotjahn and Faure 2008), but will likely be present in most cases. Further, local maxima or minima in the composite anomaly field may indicate either 1) the presence of a stronger feature in many of the events or 2) that a given feature is more consistently located in a given location (Konrad 1996).

Vertical cross sections of the composite PV anomalies are also examined. Because we are primarily interested in statistically significant features surrounding the time of ATR onset, days \(-2\), \(0\), and \(+2\) are studied. The longitudes examined span \(60^\circ-120^\circ W\), which covers the continental United States (CONUS). Vertical cross sections are constructed at \(40^\circ, 45^\circ\), and \(50^\circ\)N. The variety of latitudes and time steps chosen accounts for the fact that the PV anomalies are often mobile features and also allows us to better discern the three-dimensional spatial extent of individual PV anomaly features.

### Table 1. Number and average length of CAO and WW events that occurred in the SE United States from 1949 to 2011 during various phases of low-frequency modes. The asterisks indicate events in the positive low-frequency mode categories that are not well represented and therefore are not assigned a case number or considered further in our analysis.

<table>
<thead>
<tr>
<th>Category</th>
<th>No. of events</th>
<th>Avg length of event (days)</th>
</tr>
</thead>
<tbody>
<tr>
<td>SE CAOs, neutral NAO (case 1)</td>
<td>12</td>
<td>7.18</td>
</tr>
<tr>
<td>SE CAOs, negative NAO (case 2)</td>
<td>17</td>
<td>9.28</td>
</tr>
<tr>
<td>SE CAOs, positive NAO*</td>
<td>0</td>
<td>—</td>
</tr>
<tr>
<td>SE WWs, neutral PNA (case 3)</td>
<td>11</td>
<td>7.20</td>
</tr>
<tr>
<td>SE WWs, negative PNA (case 4)</td>
<td>12</td>
<td>8.77</td>
</tr>
<tr>
<td>SE WWs, positive PNA*</td>
<td>1</td>
<td>—</td>
</tr>
</tbody>
</table>

### 3. Results

Composite analyses of the four cases examined in this study indicate that there are large-scale as well as synoptic-scale signatures in several atmospheric variables that can be tracked backward in time for several days and hence may provide telling information on the impending arrival of such ATR events. Broad quasi-stationary circulations are usually indicative of low-frequency mode modulation, whereas smaller-scale mobile circulations generally indicate synoptic features or perturbations. In addition, several structurally coherent and statistically significant features having moderate to strong amplitude are observed in our composite PV anomaly analyses. The results of each of the four cases are detailed in the following subsections.

#### a. SE CAOs, neutral NAO (case 1)

For SE CAOs, we try to delineate the role of the NAO pattern during CAO onset, given that CAOs are favored during the negative phase of the NAO. Case 1 considers CAO events occurring during the neutral phase of the NAO. Composite time-evolution analyses for case 1 are presented in Fig. 1 (horizontal maps of several fields) and Fig. 2a (vertical cross sections of PV anomalies). In this case, a mass of cold air hovers over Canada between days \(-8\) and \(-4\) (not shown). By day \(-2\), the western portion of this air mass starts to traverse southward into the central United States, in association with northwesterly winds to the east of the Rockies (Fig. 1a). Meanwhile, northwesterly winds at the trailing (northward) edge of the cold air mass continue to feed cold air into the system, allowing for airmass intensification. By
day 0, the cold air covers almost all of the central and eastern United States, with northwesterly winds still feeding the air mass via an anomalous cyclonic circulation over southeastern Canada. The demise of the CAO event begins at day +2 when the northerlies and resulting cold-air advection behind the air mass are shut off (due in part to an anticyclonic circulation pattern that migrates southeastward from southwestern Canada and intensifies). By day +4, the temperature anomalies have decreased significantly in magnitude (not shown), likely because of the significant reduction in cold-air advection.

The role of cold-air advection, which has been found to be a key player in CAOs (Colle and Mass 1995), and its association with several prominent synoptic features is confirmed in Fig. 1b. Several days prior to the onset of the CAO events, there is a high pressure ridge extending over western Canada (day +2), coupled with a region of low SLP over eastern Canada. This synoptic structure may be the factor that initially set the cold air mass into motion. Some prior studies had found that the cold air responsible for CAOs may be forced southward from its high-latitude source region by northerly winds that develop between an anticyclone–cyclone couplet in the lower troposphere (Konrad 1996; Konrad and Colucci 1989). By day +2, this region of high pressure has intensified and become meridionally elongated, leading to strong cold-air advection along its eastern flank where northwesterly winds are crossing the thickness contours. Further, the cold-air advection is collocated with the region of statistically significant wind anomalies (Fig. 1a). By day 0, the ridge has slightly intensified and becomes tilted with a northwest–southeast orientation, while a robust surface low pressure system has formed over easternmost Canada. The circulation associated with this strengthening couplet is favorable for both sustaining and spreading the cold air mass southeastward toward the SE United States (as discussed above). By day +2, the anticyclone has migrated into the central United States and taken on a more circular shape while the low pressure system has propagated well to the east. This change in structure effectively shuts off the cold-air outbreak, as the closed anticyclonic circulation over the central United States begins to feed relatively warm air.

### TABLE 2. Discrete event dates for CAOs and WWs that occurred in the SE United States from 1949 to 2011. The asterisks indicate dates that are included in the NNR composites but that are not included in the ERA composites as a result of the availability of the data.

<table>
<thead>
<tr>
<th>Year</th>
<th>Start month</th>
<th>Start day</th>
<th>Start hour</th>
<th>Length of event (days)</th>
</tr>
</thead>
<tbody>
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<td>1955*</td>
<td>12</td>
<td>10</td>
<td>12</td>
<td>5.5</td>
</tr>
<tr>
<td>1958</td>
<td>8</td>
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<td>6</td>
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</tr>
<tr>
<td>1962</td>
<td>1</td>
<td>9</td>
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</tr>
<tr>
<td>1967</td>
<td>2</td>
<td>22</td>
<td>18</td>
<td>6</td>
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<td>6</td>
<td>5</td>
</tr>
<tr>
<td>1978</td>
<td>1</td>
<td>27</td>
<td>0</td>
<td>15.5</td>
</tr>
<tr>
<td>1983</td>
<td>12</td>
<td>16</td>
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<td>12</td>
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<td>2000</td>
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<td>18</td>
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### SE CAOs, negative NAO (case 2)

<table>
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### SE WWs, neutral PNA (case 3)

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into the upper Midwest. Finally, by day +4 (not shown),
the cold advection is substantially weaker and limited to
the far eastern United States, while warm advection
continues to occur on the western and northwestern
flanks of the high pressure center. This temporal se-
quencing is consistent with Konrad and Colucci (1989),
who found that some CAOs over the eastern United
States developed prior to cyclogenesis over the western
Atlantic Ocean, and also contains many of the aspects
described in the conceptual model by Colle and Mass

These synoptic features are also apparent in the 1000-hPa
geopotential height anomaly field (Fig. 1c), with positive
(negative) anomalies corresponding to the regions of
high (low) pressure. One key difference between the total
SLP field and 1000-hPa geopotential height anomaly field
is that the developing surface cyclone over the eastern
United States is clearly evident slightly earlier (day −2) in
the time evolution of the latter. This cyclone seems to
form just ahead of the cold air associated with the CAO
event. Otherwise, the two fields match each other quite
well. Meanwhile, at 925 hPa, we note that statistically
significant negative PV anomalies are found in the base of
the surface ridge at days −2 and 0, but no other significant
anomaly features are observed. This PV anomaly feature
is relatively stationary and would help enact a clockwise
circulation. Therefore, this dynamical feature may help
contribute to northerly or northwesterly winds over the

FIG. 1. Composite time-evolution plots for days −4, −2, 0, and +2 for SE CAOs, neutral NAO (case 1). (a) 1000-hPa wind anomalies
(m s⁻¹; vectors) and SAT anomalies (°C; shading). (b) SLP (hPa; black lines with a contour interval of 2 hPa) and total 1000–500-hPa
thickness (m; purple dashed lines with a contour interval of 100 m). (c) 1000-hPa geopotential height anomalies (m; black lines with
a contour interval of 10 m) and 925-hPa PV anomalies (shading). (d) 250-hPa geopotential height anomalies (m; black lines with a contour
interval of 50 m) and 250-hPa PV anomalies (shading). The shaded anomalies are statistically significant at the 95% confidence level. Blue
(red) shading represents negative (positive) values. In (c) and (d), the shaded PV anomalies indicate the sign only, while the boldfaced
contours indicate where the geopotential height anomalies are statistically significant.

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At upper levels, a significant long-wave trough feature is observed over eastern North America, as indicated by the intensifying negative geopotential height anomaly signature at 250 hPa (Fig. 1d). Such a feature has also been cited in prior work [Colle and Mass (1995), among others] and provides dynamical support for the surface cyclogenesis. Meanwhile, weaker-amplitude positive geopotential height anomalies are present over western North America. The primary PV anomaly signature found at this level is a positive anomaly located near the center of the upper-level trough. Unlike the lower-tropospheric PV anomaly pattern, the 250-hPa PV anomaly signature is mobile and propagates southeastward as the trough deepens. This feature would result in a counterclockwise circulation, which in most of the time steps shown would be associated with northerly or northwesterly winds over the north-central United States.

The superposition of the upper- and lower-level PV anomalies would contribute simultaneously to the CAO onset by moving cold air from Canada into the central and eastern United States, consistent with the behavior observed in the lower troposphere.

A consideration of the vertical cross sections of the composite PV anomaly field (Fig. 2a) helps us to establish which PV anomaly features are most significant during ATR onset, as well as their spatial extent. At day −2, statistically significant positive PV anomalies in the upper atmosphere (at 50°N) extend from the upper troposphere down to ~750 hPa, while the statistically significant negative PV anomaly signature near the surface extends upward to 850 hPa. By day 0, the upper-level PV signature has strengthened, spatially expanded, and propagated southward (centered at 45°N) while the near-surface PV feature has primarily strengthened. At this time,

Fig. 2. Composite time-evolution vertical cross sections of PV anomalies for days −2, 0, and +2 for all four cases. The color shading indicates the sign and magnitude of the PV anomalies (PVU), while the green contours indicate statistical significance at the 95% confidence level. Blue (red) shading represents negative (positive) values. The latitude of each cross section is specified.

The north-central United States consistent with the other composite analyses presented.
significant positive PV anomalies extend downward to 850 hPa, or just above the planetary boundary layer (PBL). Thus, the positive PV anomaly signature is likely close enough to the surface to promote northerly surface flow. A similar structure is present on day 1, except that the upper-level PV feature has progressed eastward and southward (to 40°N) while most of the near-surface feature is no longer significant. We generally find that for case 1 the most significant PV anomaly features are found in middle to upper (positive) and lowermost (negative) portions of the troposphere, with the largest anomaly magnitudes above 500 hPa. In addition, the longitudinal phase relationship between the lower-level negative PV anomalies and the upper-level positive PV allows for the two dynamical features to collectively produce anomalous northerly winds at the surface. The role of the former is most prominent just prior to and during onset while the latter feature persists longer and tends to progress eastward, leading the cold-air progression.

**b. SE CAOs, negative NAO (case 2)**

Unlike case 1, case 2 CAO events do not begin with a large pool of cold air over Canada that hovers for several days prior to event onset. Significant cold anomalies do not emerge until day 2 when relatively weak signatures appear over the Pacific Northwest and south-central Canada, respectively (not shown). By day 2, these two cold anomaly features have merged, intensified, and spread into central North America with a northerly wind signature at the leading edge of the cold air mass (Fig. 3a). Unlike case 1, the anomalous wind pattern in case 2 does not directly coincide with the temperature anomaly signature, which may account for the weaker temperature anomaly signature at this time lag. Another feature of interest at day 2 is the emerging region of significant positive temperature anomalies in the far northeastern corner of the map. As will be discussed further, this feature arises from the large-scale quasi-stationary circulation associated with the negative
NAO phase and is present at all time lags in Fig. 3. This feature, not present in case 1, highlights a fundamental difference between these two types of CAO events. Of course, this is largely anticipated, given the prescribed NAO categorization used in creating the composites. In fact, a significant NAO signature is observed in the height anomaly field over the North Atlantic Ocean at least 8 days prior to CAO onset. The longitudinal extent of the NAO circulation anomaly pattern during winter allows it to directly influence the eastern portions of North America (Thompson and Wallace 2001; Cellitti et al. 2006; Hurrell 1995). By day 0, there is a noticeable increase in the intensity and spatial extent of the cold air mass, which now covers most of the central and eastern United States. At this time, there is a substantial north-northwesterly circulation anomaly signature coincident with the cold pool. The eastern portion of this pattern is part of a broader large-scale cyclonic circulation signature that emanates from the North Atlantic in association with the concurrent negative NAO pattern. Ironically, this easterly onshore flow into northeastern Canada is also responsible for the surface warm anomaly pattern that develops over northeastern Canada. However, the role of temperature advection in producing the cold pool anomaly intensification between days −2 and 0 is unclear (given the weak circulation that exists initially). It is possible that some other in situ mechanism (or aspects of case-to-case variability) helps account for this rapid intensification. By day +2, the cold-air pattern has intensified and expanded farther with the largest anomaly magnitudes located over the eastern and southeastern United States. However, northeasterly flow coming from the emerging warm anomaly over northeastern Canada (which has intensified and expanded) is now funneling warm, moist air toward the northern edge of the cold air mass. This subsequently serves to effectively ‘shut off’ the ability of the large-scale circulation to tap into cold continental air from Canada, not only preventing the CAO from further intensification, but also leading to its eventual demise. It is evident that this weakening process has begun by day +4. We normally think of the NAO as creating favorable conditions in the atmospheric circulation for cold air to funnel into the eastern United States since it is associated with a trough over the eastern United States (e.g., Fig. 3d). However, as illustrated here, the role of long-lived NAO events may not be quite so straightforward, as the NAO can also contribute to CAO demise at longer time lags. Thus, the role of the NAO in CAO events may be self-limiting, as it can both help initiate CAO onset (at short time lags) as well as contribute to CAO demise (at longer time lags).

Several days prior to CAO onset (day −4), a high pressure region exists over southwestern Canada with low pressure over eastern Canada and the western Atlantic (not shown), similar to the initial setup for case 1. However, the two case evolutions diverge thereafter. By day 0, the region of high pressure has intensified and expanded southward into the western United States, leading to cold-air advection along the eastern flank of the surface ridge, while the low pressure center also strengthens (Fig. 3b). This general pattern persists through day +4, with minimal propagation eastward. By day +4, the cold advection over the eastern United States has weakened, consistent with a weakening of the SLP couplet. This sequence of events confirms the anomalous circulations in the wind field and is consistent with the evolution of the temperature anomalies.

These features are also evident in the 1000-hPa geopotential height anomaly field (Fig. 3c), with significant positive (negative) height anomalies over the western (eastern) United States. Another feature of interest is the region of positive geopotential height anomalies observed just off the western coast of British Columbia. These anomalies are initially present at day −4, but have strengthened by day −2. By day 0, this feature has moved eastward and merged with another ridge over the eastern slopes of the Rockies. Therefore, unlike case 1, case 2 onset is strongly affected by a large-scale low-frequency mode juxtaposed with a migratory synoptic-scale disturbance. It appears that the synoptic-scale disturbance originally over western North America interacts with the NAO circulation to produce a more intense ATR. This observation is consistent with prior research that has shown that high-frequency transient eddies can amplify teleconnection patterns via positive feedback [Barnes and Hartmann (2010), and references therein]. At 925 hPa, statistically significant negative PV anomalies are located within the surface ridge at days −2 and 0, similar to case 1. These PV anomalies remain stationary, as they do in case 1, over the northern Great Plains and would help to enact a clockwise surface circulation. This would produce northerly or northwesterly winds over the north-central United States consistent with the other fields presented.

In the upper troposphere, a persistent long-wave trough is observed over eastern North America, associated with negative geopotential height anomalies at 250 hPa (Fig. 3d), and intensifies through the onset period. As expected during an NAO event, this trough is paired with an oppositely signed ridge feature to the north. This pairing is further exemplified in Fig. 4 (left), where the negative NAO signature over the North Atlantic Ocean is clearly evident. This quasi-stationary north–south dipole structure is distinct from the more transient and mobile trough structure observed in case 1. There is also a weaker-amplitude positive height
anomaly feature over the western coast of North America. The parallel 250-hPa PV anomaly evolution reveals two persistent and roughly quasi-stationary features consistent with the geopotential height field: a low-over-high dipole over the eastern portion of the domain and a negative anomaly feature over the western coast. These features are more expansive and geographically fixed than those observed in case 1. The eastern PV dipole structure would be associated with a circulation that would favor both northerly flow over the central United States and easterly flow over northeastern Canada, contributing to both CAO onset as well as its follow-up demise via onshore flow over eastern Canada. The region of negative PV anomalies in the upper ridge over the western coast is probably located too far away from the CAO event as well as its follow-up demise via onshore flow over eastern Canada. The region of negative PV anomalies in the upper ridge over the western coast is probably located too far away from the CAO event to make an appreciable direct contribution, but may indirectly play a role by developing the surface high pressure system east of the Rockies, which in turn contributes to the CAO. This upper-level anticyclonic signature is not as prominent in case 1, but has been identified in prior studies as an important feature during CAO events in the SE United States (e.g., Konrad 1998).

The vertical PV anomaly cross sections for case 2 are discussed next (Fig. 2b). Although there is some qualitative commonality with case 1, there are also differences that parallel the discussion of Fig. 3. First, the upper-level positive PV anomaly signature is longitudinally more expansive and less mobile in case 2 than case 1. In addition, the relatively stationary nature of the upper-level PV anomaly feature identified in the horizontal PV anomaly maps (Fig. 3) is also evident in the vertical cross sections, as the primary vertical axis remains centered near 90°W during onset. Similar to case 1, at day 0 significant positive anomalies extend down to about 775 hPa or just above the PBL. There is also a region of significant negative anomalies near the surface over the western United States at this time, as noted in Fig. 3c. However, the lower-level negative PV anomaly feature (located in the surface ridge) is weaker and less statistically robust in case 2. By day +2 the two features are more clearly detached from one another, as the midtropospheric signature erodes above the PBL.
As in case 1, the upper-level positive anomaly signature is strongest at this time lag. Further, the strongest lower-level negative PV anomalies are located to the west of the positive PV anomalies at upper levels. This placement allows for the collective influence of both anomalies in producing northerly winds at the surface, as discussed earlier.

c. SE WWs, neutral PNA (case 3)

The WW counterparts to CAOs are discussed next. For SE WWs, we attempt to delineate the role of the PNA pattern in WW onset, noting that WWs are favored during the negative phase of the PNA (Westby et al. 2013). Case 3 considers WW events occurring during the neutral phase of the PNA with the composite synoptic analyses presented in Fig. 5. Prior to WW onset, an anomalous cyclonic flow develops over the western coast of North America with anomalous southerly flow at its leading edge producing local surface warming (Fig. 5a).

At day −2, we also see the initial circulation signature over the SE United States of an anomalous high pressure system forming in situ over the western North Atlantic. By day 0, the cyclone has extended southeastward and the combined circulation signature of this cyclone–anticyclone couplet leads to a broad swath of anomalous southerly flow extending from Texas to New England. An elongated warm-air anomaly forms in association with this southerly flow. By day +2, the warm anomaly has intensified and expanded into the upper Midwest. Around day +4, the event starts to weaken as the warm-air advection reaches a peak. The evolving wind field in this case appears to be associated with both mobile (cyclone) and quasi-stationary (anticyclone) synoptic features.

The cyclone–anticyclone couplet is also evident in the SLP field (Fig. 5b). On day −2, there is a trough of low pressure extending into the Pacific Northwest from the Gulf of Alaska and a developing high pressure system...
just off the SE coast. These pressure features are linked to the anomalous circulations presented in Fig. 5a, and warm advection occurs over the north-central United States in association with the low pressure feature. By day 0, both have intensified, with the trough (high pressure system) expanding southeastward (northeastward) and a broad area of warm advection extending in between. A distinct low pressure system emerges by day +2 over the southern United States, while the trough over the Pacific Northwest becomes less distinct. These features are separated by a weak high pressure system. Meanwhile, the high pressure system to the east of the United States strengthens and becomes anchored over Bermuda (akin to a Bermuda high). Southerly flow becomes concentrated in between the southern low and the Bermuda high, feeding warm air into the eastern United States. However, by day +4 the areal extent of the warm advection is reduced, in association with the eastward movement of the surface low.

Many of these synoptic features are also evident in the 1000-hPa height anomaly field (Fig. 5c), with negative (positive) anomalies corresponding to anomalous low (high) pressure. The height anomaly field isolates the migratory behavior of the surface low pressure system as well as the more in situ development of the offshore high pressure system. Little to no coherent structure exists in the accompanying 925-hPa composite PV anomaly field, in contrast to the previously discussed CAO cases.

In the upper troposphere, a weakening (strengthening) trough (ridge) anomaly is observed over western (eastern) North America (Fig. 5d). This pattern is generally opposite to those observed for the CAO cases. The negative geopotential height anomalies are associated with the trough, while the positive anomalies are indicative of the ridge. On day −2, the primary features of significance are (i) the broad negative height anomaly feature off the Pacific coast and (ii) a short-wave ridge anomaly over the southwestern United States. By day 0, both have propagated eastward with a new ridge feature arising over the northeastern United States. Thereafter, the trough dissipates while the two ridge features merge, intensify, and stall over the northeastern United States. The simultaneous weakening of the upstream trough and strengthening of the downstream ridge is consistent with eastward dispersion of Rossby wave energy (e.g., Blackmon et al. 1984a,b). The upper-level ridge feature is likely dynamically linked to the oceanic surface high located to its east. Both the western trough and eastern ridge are associated with statistically significant PV anomaly features at 250 hPa that track the movement and intensity of the upper-level height anomaly features. Both a decaying positive PV anomaly to the west and a strengthening negative PV anomaly to the east are configured in such a manner to contribute (intermittently) to anomalous southerly flow observed initially over the upper Midwest (day −2) and thereafter over the eastern United States.

Vertical cross sections of composite PV anomalies for case 3 are presented in Fig. 2c. Two days prior to WW onset, there are significant large-amplitude upper-level PV anomalies at 45°N in association with the western trough (positive anomalies) and the two eastern ridge features (negative anomalies). Below 500 hPa, the PV anomaly signature is locally either weak in amplitude or insignificant (or both). The sole exception to this rule is a modest negative anomaly feature centered around 775 hPa at 95°W. By day 0, the two significant negative PV anomaly features begin to merge at upper levels. Appreciable anomaly amplitudes extend downward from these two features to the midtroposphere (~500 hPa). There are no notable significant PV anomaly structures found in the lower troposphere. The upper-level negative PV anomalies may contribute to the southerly surface winds observed over the south-central United States at this time. By day +2, the two lobes have shifted eastward slightly while continuing to slowly merge longitudinally. It is useful to note that we end up focusing our WW cross sections primarily along 40°N (except for day −2 of case 3). This reflects the general latitudinal stationarity of the most prominent WW dynamical features in contrast to the gradual southward movement of key dynamical features for the CAO events. Another distinction between the WW and CAO events is the general lack of coherent lower-level PV anomaly features having appreciable amplitude and/or significance in WW cases, suggesting that WW events may be largely driven by upper-tropospheric dynamics.

d. SE WWs, negative PNA (case 4)

The evolution of case 4 is quite different from the previous WW case just considered. In case 4, the warm anomaly forms more abruptly around the time of event onset. Unlike the previous case, at day −4, no significant warm anomaly features are observed (Fig. 6a), nor are there any statistically significant southerly winds in the region of interest. By day −2, weak warm anomalies begin to emerge over the SE United States, in association with a robust anomalous anticyclonic circulation that forms along the mid-Atlantic coast. This circulation promotes warm advection over a broad region encompassing the SE United States. By day 0, the warm pattern has expanded to now include the southeastern, northeastern, and south-central United States, and has intensified in strength. Meanwhile, the eastern edge of an emerging anomalous cyclonic circulation is now observed in the wind anomaly field over the Midwest. The east–west dipole pattern
across the United States is highly reminiscent of the surface manifestation of the PNA. This feature is not surprising given the prescribed sign of the PNA used in creating the composites. As will be discussed in relation to Fig. 4 (right), the PNA signature is visible in the upper-tropospheric geopotential height field at least 4 days prior to CAO onset. This feature was not present in case 3, and highlights a fundamental difference between the two categories of WW cases. By day $12$, the warm SAT anomalies reach their maximum amplitude, as southerly winds between the anticyclone–cyclone couplet continue to advect warm air into the region. Another interesting feature present in the composite at day $12$ is the distinct front/dryline over the south-central United States. Such frontal-type features are evident in several of the near-surface fields presented in Fig. 6, illustrating one of the benefits of using four-times-daily data. By day $14$ (not shown), the warm anomaly begins to weaken and retract in areal extent.

The role of the warm advection and its connection to several prominent synoptic features are confirmed in Fig. 6b. Several days prior to onset (day $-4$), there is a region of relatively high surface pressure located over the central United States, which, if anything, is associated with cold advection over the SE United States. However, by day $-2$, this area of high pressure has migrated eastward and becomes centered just off the SE coast, leading to warm advection over the SE region consistent with the anomalous wind field (Fig. 6a). The anticyclone weakens slightly by the time of onset (day $0$), while surface pressure falls are observed east of the U.S. Rockies. These lower pressures are associated with the cyclonic circulation signature noted earlier in the wind anomaly field. By day $+2$, the anticyclone strengthens once again and expands northward, leading to concentrated warm advection over the SE region. In addition, a closed low pressure center emerges over the central United States. The counterclockwise circulation associated with this feature...
contributes to the southerly winds and warm advection over the SE region. The frontal feature is also evident as a trough trailing southward from the center of the low pressure system. By day +4 (not shown) both of the pressure systems have weakened in magnitude and migrated such that the southerly flow over the SE region is reduced, ultimately leading to event demise.

These near-surface synoptic features are also evident in the 1000-hPa geopotential height anomaly field (Fig. 6c), with positive (negative) height anomalies corresponding to the regions of high (low) pressure. However, the degree of similarity between the SLP and 1000-hPa height anomaly fields appears to be less for this case compared to the previous three composites, particularly over the western United States. This is likely due to mismatches in the regional patterns of the climatological and anomalous height–pressure fields. As for case 3, no significant PV anomaly structures are observed near the surface.

In the upper troposphere, we observe a persistent (strengthening) quasi-stationary trough (ridge) anomaly feature over northwestern (southeastern) North America (Fig. 6d). This pattern and its temporal evolution are qualitatively similar to case 3, but the centers of action in the current case are on average (i) stronger and (ii) less mobile (more geographically fixed). At day −4, a significant trough feature is observed over western Canada. By day −2, the trough anomaly has strengthened and extended northeastward, resulting in a southwest-to-northeast horizontal “tilt” in the anomaly structure. Such a structure is favorable for southward dispersion of Rossby wave energy toward the continental United States (e.g., Hoskins et al. 1983). This tilted structure is maintained through day 0 and during this time the ridge anomaly over the eastern United States intensifies, consistent with expectations of Rossby wave theory. By day +2, the large-scale wave pattern has become more amplified as the ridge over the eastern United States strengthens. However, the upstream trough has altered its structure with more of a north–south anomaly tilt. Thereafter (not shown), the overall flow pattern continues, but the ridge amplitude weakens as the trough structure is less favorable for southward energy dispersion. The 250-hPa composite PV anomaly evolution parallels the height anomaly evolution, with positive (negative) PV anomalies located over western Canada (eastern United States). Like case 3, the configuration of these upper-level PV anomaly features leads to an associated circulation anomaly pattern that is consistent with the observed anomalous southerly flow over the central United States. However, in contrast to case 3, the anomalies are quasi-stationary during the evolution and there seems to be a pulsing of energy through the quasi-stationary wave pattern (with the downstream center strengthening at the expense of the upstream center). These characteristics suggest a prominent role for a quasi-stationary wave train feature in case 4 WW onset. To further investigate the role of a stationary wave train feature, we have studied the 250-hPa geopotential height anomaly field over an expanded domain that includes much of the North Pacific Ocean (Fig. 4, right). There is, in fact, a statistically significant east–west wave train anomaly structure (of alternating sign) emanating from near the Aleutian Islands and arching northeastward into Canada and then southeastward into the eastern United States, consistent with the negative PNA pattern. We additionally note that the evolution shows that the downstream ridge anomaly strengthens after the formation of significant anomaly centers upstream, consistent with Rossby wave dispersion through a quasi-stationary wave train. Interestingly, in contrast to case 2 where the negative NAO interacts with a synoptic-scale disturbance, case 4 does not include a synoptic disturbance that interacts with the negative PNA. This observation is consistent with prior studies that have found that the NAO is governed by nonlinear dynamics, whereas the PNA is primarily driven by linear dynamics (Feldstein 2002, 2003; Evans and Black 2003).

The PV anomaly cross sections for case 4 (Fig. 2d) mainly reflect the structure associated with the ridge anomaly over the eastern United States, as the western trough anomaly is largely confined to higher latitudes. Similar to case 3, the primary signatures of note are located at 500 hPa and above. Generally, we find that the most prominent PV anomalies are found in the mid- to upper troposphere, with the largest anomalies located near the tropopause. The initially amorphous upper-level negative PV anomaly signature (related to the eastern ridge) coalesces and strengthens during WW onset. The relatively stationary nature of the upper-level PV anomalies that was identified in the horizontal maps in Fig. 6 is also evident in the vertical profiles, as the negative upper-level anomalies remain centered between 80° and 90°W. In this case, however, we do observe a significant near-surface negative PV anomaly feature over the western United States that extends up to about 775 hPa. However, this feature does not appear to play a role in WW onset.

4. Conclusions

Our study identifies the main synoptic and dynamic signatures associated with different categories of ATR onset over the southeastern (SE) United States with an emphasis on delineating the role of large-scale low-frequency modes. This is approached via composite time-evolution analyses of several different field variables for four different ATR categories (two CAO and two WW categories) based upon
the concurrent sign and amplitude of relevant low-frequency modes. Our composite analyses reveal significant roles for both synoptic and large-scale disturbances during CAO and WW onset over the SE United States. Each of the four categories of events involves transient synoptic features, a low-frequency mode influence, or a juxtaposition of the two. We summarize the main features identified as follows.

- **Case 1 (SE CAOs, neutral NAO):** CAOs of this type are linked to a transient high pressure system propagating southeastward from Canada followed by synoptic-scale cyclogenesis over eastern Canada.
- **Case 2 (SE CAOs, negative NAO):** Although CAOs of this type are also associated with a mobile high pressure system from the west, it is linked to a large-scale quasi-stationary low pressure system over the eastern United States, which represents part of the southern manifestation of the NAO pattern. Thus, the NAO pattern plays a direct role in establishing the regional circulation leading to the CAO.
- **Case 3 (SE WWs, neutral PNA):** WWs of this type are initially (subsequently) associated with anomalous low (high) pressure near the Pacific Northwest (SE) coast. Both of these features are transient and exhibit eastward propagation.
- **Case 4 (SE WWs, negative PNA):** Although WWs of this type are also associated with anomalous low (high) pressure to the northwest (east), these features are nearly quasi-stationary and represent the downstream centers of the negative phase of the PNA pattern. Thus, these events have dynamical roots in the extratropical North Pacific.

Thus, our results provide a qualitative indication of the role of low-frequency modes in ATR onset via their impact on regional circulation and temperature advection.

The composite circulation anomaly patterns observed preceding and during ATR onset generally consist of either negative (CAOs) or positive (WWs) geopotential height anomalies in the upper troposphere near the region of interest (SE United States) with oppositely signed anomalies located in the lower troposphere west of the region of interest (over the central United States). In some of the cases, an east–west dipole pattern is present at the surface, resulting in enhanced northerly or southerly flow. In any case, the anomalous circulation patterns act to promote cold (CAO) or warm (WW) advection into the region of interest, ultimately forming the cold or warm SAT anomalies that define the extreme events. The primary differences among these cases lie in the origin and nature of the circulation anomaly features linked to ATR onset. These results generally confirm aspects of prior CAO studies (e.g., Konrad 1996, 1998; Konrad and Colucci 1989; Colle and Mass 1995) while providing distinctly new results on the synoptic behavior of WW development and establishing the synoptic role of low-frequency modes in both CAO and WW onset.

Our composite PV anomaly analysis provides additional information regarding coherent dynamical features associated with ATR onset. The composite PV anomaly structure is generally characterized by prominent PV anomaly patterns in the mid- to upper troposphere consistent with the coincident geopotential height anomaly patterns (i.e., anomalous ridge features are linked to negative PV anomalies while anomalous troughs are related to regions of above normal PV). The largest PV anomaly amplitudes are found in the upper troposphere with weaker anomalies in the lower atmosphere nearest the surface. The weakest anomalies are located in the midtroposphere. These upper-tropospheric PV anomaly patterns have horizontal structures that would be associated with southerly (northerly) winds over the central and eastern United States during WW (CAO) onset. In some cases, these induced flow patterns are supplemented with contributions from upstream PV anomaly features of opposite sign. Comparatively, even though the PV anomaly amplitudes near the surface are smaller in magnitude and area than those in the upper troposphere, their closer proximity to the surface, where the key meridional air temperature transport occurs, gives them the potential to provide a fundamental contribution to ATR onset.

The specific role of low-frequency modes in ATR evolution is interesting and includes some unexpected aspects. In particular, in case 2 the NAO contributes to both SE CAO onset as well as event demise. This result is counterintuitive as the low-frequency forcing ends up limiting the event time scale (even though the low-frequency mode, itself, persists). A prominent role for low-frequency modes is also indicated for case 4, in which case the PNA teleconnection pattern helps to establish the essential North American circulation centers. Thus, our results indicate that low-frequency modes serve to both favor ATR formation (cases 2 and 4) as well as modulate the event duration (case 2). In principle, such links to low-frequency modes suggest some potential for longer-term predictability of ATR events. For instance, a recent study by Scaife et al. (2014) demonstrated that aspects of the NAO behavior can be predicted months in advance. Because the NAO strongly modulates winter weather over North America and Europe, the predictability of the NAO may ultimately lead to increased skill in predicting regional winter climate behavior (Scaife et al. 2014). However, advances in predicting
specific ATR episodes (or their likelihood) will be predicted on first obtaining a more complete mechanistic understanding of the physical linkages between NAO events and ATR life cycles. Such knowledge would help in moderating or reducing the risk of ATR-related impacts on society.

This research has also provided some insight into the general behavior and development of CAOs and WWs. One distinction between these two classes of events is that CAOs appear to be related to substantial airmass transport (normally from Canadian latitudes) while WW formation occurs more locally over the region of interest (with the air mass emanating from the adjacent Gulf of Mexico or North Atlantic). This behavior is indicated in our composite wind analyses, as during CAOs the strong northerlies are typically observed at the leading edge of the cold air mass (Figs. 1a and 3a), while during WWs strong southerlies spatially coincide with the developing warm air mass (Figs. 5a and 6a). In addition, CAO cold air masses form in high latitudes and move southward over at least 20° latitude, while WW warm air masses form in more southerly latitudes but typically only traverse 5° latitude or less.

Although the current research provides an important synoptic–dynamic characterization of ATR formation, additional dynamically based diagnostic research will be required to obtain a more complete mechanistic understanding of ATR life cycles, including the specific role of low-frequency modes. Our follow-up research efforts will include regional heat budget, wave activity, and potential vorticity inversion diagnoses. The thermodynamic budget will provide quantitative insight into the relative importance of temperature advection, and adiabatic and diabatic processes in ATR formation, while the wave activity flux analyses will help clarify the role of low-frequency modes for different ATR categories. Under the assumption that horizontal temperature advection plays a leading role in ATR onset, the PV inversion analyses can be used to directly assess which of the dynamical features identified in the current study directly contribute to local temperature changes that occur. For example, this will allow us to directly assess the respective roles of the upstream anticyclone and downstream cyclone during CAO onset (e.g., Konrad 1998). Such additional knowledge of ATR physics is essential for optimizing the simulation of ATR events in both weather forecasts and climate simulations.

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REFERENCES


