Northeast United States heat waves: a statistical analysis and synoptic climatology

Scott Cooper Runyon

University at Albany, State University of New York, runyon@atmos.albany.edu

The University at Albany community has made this article openly available. Please share how this access benefits you.

Follow this and additional works at: https://scholarsarchive.library.albany.edu/legacy-etd

Part of the Atmospheric Sciences Commons

Recommended Citation


This Master's Thesis is brought to you for free and open access by the The Graduate School at Scholars Archive. It has been accepted for inclusion in Legacy Theses & Dissertations (2009 - 2024) by an authorized administrator of Scholars Archive. Please see Terms of Use. For more information, please contact scholarsarchive@albany.edu.
NORTHEAST UNITED STATES HEAT WAVES:
A STATISTICAL ANALYSIS AND SYNOPTIC CLIMATOLOGY

by

Scott C. Runyon

A Thesis
Submitted to the University at Albany, State University of New York
in Partial Fulfillment of
the Requirements for the Degree of
Master of Science

College of Arts and Sciences
Department of Atmospheric and Environmental Sciences
2011
ABSTRACT

The results of a statistical analysis of anomalously hot day and heat wave frequency in the United States (U.S.) are presented, along with a synoptic climatology of northeast U.S. heat waves (NHWs) for 1948–2001. This study used daily high temperatures retrieved from the National Climatic Data Center (NCDC) Daily Surface Dataset for 54 stations binned into the nine standard NCDC climate regions. Anomalously hot days were defined as when the daily high temperature exceeded the climatological 97.5-percentile temperature. Heat waves were defined as at least three consecutive anomalously hot days, whereas regional heat waves (e.g., NHWs) were defined as simultaneous heat waves at two or more cities within a region.

Time series of anomalously hot day and regional heat wave frequencies were prepared for 1948–1959, 1960–2001, and 1948–2001 for the nine individual NCDC regions, composites of all nine regions, and composites of all Eastern (Western) regions located east (west) of the Rockies. Statistically significant ($R^2>0.2$) positive trends of anomalously hot days were found in the Western regions composite and Southeast for 1948–1959 and 1960–2001, respectively. A subjective analysis of anomalously hot days and regional heat waves shows that the Eastern regions and entire U.S. experience higher frequencies in the 1950’s, 1980’s, and 1990’s, whereas the Western regions experience increasing frequencies throughout the period 1948–2001.
Composites analyses link NHWs to the dynamically driven amplification and subsequent eastward translation of an upper-level ridge over North America associated with a Rossby wave train (RWT) initiated by geopotential height rises over the east-central North Pacific that may be linked to an RWT originating over East Asia or to convection over the equatorial Pacific. This ridge amplification and translation strengthens and advects surface-based mixed-layers and low-level continental tropical air masses from over elevated arid and semiarid terrain in the Intermountain West towards the Northeast. This warm air is maintained along the equatorward flank of a strong subtropical jet (STJ) and along the northern periphery of a strong midlevel anticyclone, and arrives in the Northeast as the region experiences subsidence-induced warming beneath the equatorward exit region of the STJ.
ACKNOWLEDGMENTS

Firstly, I’d like to dedicate this thesis to my deceased father, Chapman L. Runyon, who passed away many years ago and about whom I still think about often. I’d also like to acknowledge the steadfast care, support, and love of my mother Lillian Runyon throughout these many years.

This project would never have been completed without the help, guidance, and unbelievable patience of Professor Lance Bosart and Daniel Keyser, whom I will be forever indebted to.

I'd like to thanks Dr. Alan Srock, Dr. Nick Metz, Joseph Villani, Dave Groenert, Tom Galaneau, Ross Lazear, Kevin Tyle and all the other graduate and instructors students I’ve met throughout the years. Thank you for your friendship and support.

Special thanks to my wife Dr. Heather Archambault for her continued patience and assistance with this thesis. The project would never have been of the quality it is without her love and support.
1. Introduction

1.1 Overview/Motivation

Heat waves are a significant contributor to the total number of weather-related fatalities across the entire United States (U.S.). From 1988 through 2004, heat-related fatalities outnumbered all those associated with other types of severe weather. During this period, hurricanes killed 312 people while tornadoes took 886 lives; lightning strikes and floods killed 925 and 1411 people, respectively. Extreme heat, on the other hand, claimed a total of 2518 people within the same time period. (NOAA/NWS 2004) A better understanding of heat waves could lead to improved forecasts and possibly result in decreased mortality rates. These forecasts may become critical, given the assumption that the frequency, severity, and duration of heat waves may increase due to a warming climate. Published work by Robinson (2001) and a joint report by Ozone Action and Physicians for Social Responsibility (Davies et al. 2000) have already shown an increase in the frequency of heat waves in several regions across the U.S.

Previous research has largely focused on long-term heat wave–droughts and climatologies published on the subject have been limited in geographical scope. In addition, research on individual synoptic time-scale severe heat waves is difficult to find. For these reasons, short-lived northeast U.S. heat waves (NHWs) have largely been overlooked by the research community. The purpose of this research is to investigate heat wave frequency and trends across the U.S.,
to compare the results with previously published studies, and to determine if both
the antecedent conditions to and the synoptic evolution of NHWs exhibit
consistent and identifiable patterns.

1.2 Literature Review

1.2.1 North American anticyclone

Today, atmospheric scientists are well aware of the continental
anticyclone that typically resides over North America during the summer season.
This awareness, however, was not always the case as evidenced in the
published works of Reed (1933, 1937, 1939), which represent some of the
earliest research done on this topic. The following excerpt from Reed (1933)
summarizes the state of knowledge concerning the North American anticyclone
at that time: “The existence of an anticyclonic system...at high levels over the
southwestern portion of the North American continent during the warm season is
one of the most interesting and important revelations yielded by aerological
investigations...” It had been previously thought, Reed (1933) comments, that a
type of “monsoonal wind circulation existed over California and the neighboring
ocean [during the] summer.” This investigation, aided by data collected from
newly installed (circa 1928) upper-air observation sites across the southwest
U.S., determined that the anticyclone was a warm-season phenomenon, existing
between April and October, reaching its maximum intensity during the months of
July and August. Reed (1933) estimated that the mean position of the North
American anticyclone, at peak intensity, was over the southern Rocky Mountains, but that its position was “…quite variable, the crest being sometimes west of the Continental Divide and sometimes east of it”. Reed (1933) also noted that the anticyclone began its life over the elevated terrain of Mexico and migrated northward as the summer season progressed, later retreating back to Mexico as summer came to an end.

Reed (1933) also established a correlation between the anticyclone position and anomalously high surface temperatures, concluding that surface heating was a ‘controlling factor’ in determining the location of the anticyclone. He noted that when the anticyclone was located east (west) of the Rockies, surface temperatures were significantly above normal east (west) of the Rockies. Reed’s (1937) research would later prove that certain summertime weather regimes coexist with the positioning of the North American anticyclone by noting the “…preponderance of times [in the 1930’s] at which the anticyclone was centered east of the Rocky Mountains”, a time presently referred to as the “Dust Bowl”, which is associated with intense heat and deficit rainfall amounts in that region. Reed (1937) also established an apparent correlation between the anticyclone center and those areas that exhibited below normal precipitation. Reed (1937) realized that above normal heat (and drought conditions) occurred when a “…uniformity of circulation at all levels…” existed, referring to the co-location of an intruding low-level North Atlantic anticyclone (“Bermuda high”) with the mid-level North American anticyclone. These coupled anticyclones, a climatological anomaly, produce deep subsidence through the troposphere and
are closely related to heat waves. Enhanced tropospheric static stability, a result of subsidence, leads to decreased cloud production and precipitation, which allows above normal levels of sunshine to heat the surface, producing above normal temperatures. An example, as reported by Reed (1933), correlates the strong static stability associated with the North American anticyclone with a dearth of convection over California in the summer. Reed (1937) also recognized that a deep anticyclonic regime makes it highly unlikely that cooler air masses could encroach on the anticyclone’s position, allowing for additional heating and greatly reducing the chances of frontal precipitation.

As was shown in Parker et al. (1989), Bell and Bosart (1989), and Galarneau et al. (2008), the North American anticyclone, during the summer season, is preferentially located over the mountainous southwestern U.S. and northern Mexico. Parker et al. (1989) prepared a 36-year climatology (1950–1985) of 500-hPa closed anticyclones (CAs) over North America by counting the percentage of times at least one closed 6-dam contour of higher geopotential heights could be identified in a grid system divided into 10° by 10° latitude–longitude quadrangles during the period. Figures 1.1a–c show the tendency for the North American anticyclone to be positioned in the southern and southwestern regions of the U.S. during meteorological summer (June–August). Bell and Bosart (1989) constructed a 15-year climatology (1963–1977) of 500-hPa CAs across the Northern Hemisphere (NH) using a 2° by 5° latitude–longitude gridded dataset bounded meridionally between 24°N and 82°N. CAs were defined as those circulation centers containing at least one closed 30-m
contour around a central maximum geopotential height value. Consistent with Parker et al. (1989), Bell and Bosart (1989) (Fig. 1.2) found that closed anticyclone centers are most frequently positioned over subtropical continents during meteorological summer and that their occurrence is maximized over the southwestern U.S. Galarneau et al. (2008) constructed a global 54-year climatology (1950–2003) of CAs using a methodology adapted from that used in Bell and Bosart (1989) using twice-daily 2.5° by 2.5° latitude–longitude gridded National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis 200-, 500-, and 850-hPa geopotential height fields. (Kalnay et al. 1996) Galarneau et al. (2008) (Figs. 1.3a,b) showed that 200- and 500-hPa CAs are, in general, more frequent over subtropical continents during meteorological summer, and that a local maximum exists over the southwest U.S. and northern Mexico, a result consistent with that of Parker et al. (1989) and Bell and Bosart (1989).

The relationship between CAs and significant above normal surface temperatures is now a well-recognized phenomenon. In order for those areas east of the Rockies to experience extreme heat, the North American anticyclone must be displaced eastward. This fact is nicely seen in the work done by Karl and Quayle (1981) on the 1980 season-long heat wave that occurred throughout the Southern Plains and southeast U.S. Figures 1.4a–f depict the broad coexistence of areas of both strongly positive 700-hPa geopotential height (Figs. 1.4a–c) and anomalously warm surface temperature (Figs. 1.4d–f). Note the geographical displacement of the North American anticyclone into the southern
and southeastern U.S. As the anticyclone strengthened and progressed eastward, over time, the intensity (denoted by shading) and eastward extent of the heat increased as well.

1.2.2 Extended heat wave–droughts

The bulk of the research done on heat waves has, for the most part, been focused on long-term heat wave–droughts. Although heat waves had been studied before, the multiyear heat wave–drought that occurred during the summers of 1952–1954 was the first to be studied in detail. Monthly summaries published during this period in *Monthly Weather Review (MWR)* detail the NH circulation associated with anomalously warm surface temperature. Klein (1952a) noted, “Almost the entire United States east of the Rocky Mountains was under the influence of anticyclonic curvature, anticyclonic shear, and above-normal geopotential heights north of a well-developed High centered on the east Gulf Coast.” Figure 1.5 shows the contoured mean and anomaly 700-hPa geopotential height fields for the period of 1–29 June 1952. Klein (1952a) quantified the abundance of clear skies, sunshine, and solar heating by calculating the percentage deviation from normal values. Each of the quantities was well above normal in areas located underneath the mean position of the 700-hPa anticyclone. Klein also observed two key features of the 700-hPa geopotential height field of June 1952. One of these features was a tight gradient that produced a consistently zonal flow across the U.S.–Canadian
border, which inhibited cooler air masses from pushing southward into the U.S. Secondly, the gradient created between the ridge over the eastern U.S. and a trough centered over the West Coast likely led to the transport of hot, desert-like (continental tropical; cT) air from the southwestern U.S. eastward.

The characteristics discussed by Klein (1952a) were consistent with the conclusions reached by Reed (1933, 1937, 1939). Klein (1952a) also asked a simple question: "Was the circulation pattern of June 1952 typical of heat wave situations in summer in the central United States?" Klein (1952a) averaged the 700-hPa geopotential height anomalies observed during the 10 hottest 5-day mean periods (non-consecutive) at Kansas City, MO, during meteorological summer from 1947 through 1951. These values were then subtracted from normal geopotential height values for the month of June to create a composite anomaly map (Fig. 1.6). Characteristics common to both plots (June 1952 vs. the composite map) are above normal ridging over the Mississippi Valley, a deeper-than-average trough along the West Coast, a weak trough along the East Coast, an area of confluence in southern Canada, and generally below-normal 700-hPa geopotential heights across Canada.

Klein (1952a) next expanded the statistical analysis from a composite anomaly of 10 cases to a simple linear correlation between surface temperature anomalies at Kansas City and 700-hPa geopotential height anomalies. Selected were the 105 hottest 5-day mean periods at Kansas City from 1946 through 1949. The resulting correlation field showed that "above-normal temperatures in Kansas City in summer are generally accompanied by above-normal 700-mb
geopotential heights in the eastern two-thirds of the [U.S.] and the eastern Pacific (Fig. 1.7)." Areas in central Canada and along the West Coast show a negative correlation, indicative of below-normal geopotential heights (Fig. 1.7). Klein (1952a) discovered a “striking parallelism between the lines of equal correlation…and the lines of equal 700-mb geopotential height anomaly…” and concluded that a “fundamental [interrelation] between temperature and circulation [exists]…for extremely warm cases…” in the eastern U.S. Klein (1952a) was of the belief that central and eastern U.S. heat waves occur under specific conditions, and that the circulation of June 1952 was simply one such example of this unique pattern.

The unusual heat of June 1952 was the beginning of what would become a triennium of abnormally warm summers marked by an unusually persistent circulation across the U.S. MWR summaries by Winston (1953), Klein (1953), Hawkins (1953), Holland (1954), Hawkins (1954), and Winston (1954) reinforced many of the conclusions reached by both Reed (1933, 1937, 1939) and Klein (1952a). Winston (1953) noted similarities between the temperature and precipitation anomalies, as well as the circulation patterns, of both June 1952 and June 1953 across North America. Winston (1953) identified this pattern as a ‘wave train’, noting the clear ridge–trough–ridge pattern that existed in June 1953, and found that a similar pattern was present in June 1952 as well. Winston (1953) details the mechanisms that were in place for the growth and maintenance of the continental anticyclone, a necessary step for a heat wave to develop. Winston (1953) describes an unusually deep trough along the West
Coast, abnormally fast westerlies at the U.S.–Canada border, and a strong ridge over the east central Pacific. This description concurs with what Klein (1952a) had found earlier, with Winston (1953) believing that the June 1953 heat wave matched the typical heat wave circulation that Klein (1952a) had discovered. Winston (1953) also calculated the 700-hPa mean relative vorticity field for the month and showed that anomalously high 700-hPa geopotential heights over the U.S. were co-located with the largest values of anticyclonic vorticity. Winston (1953) noted that both anticyclonic curvature in the geopotential height field and anticyclonic speed shear in the wind field, both located in areas north of the 700-hPa high center, were significant contributors to the total vorticity.

Klein (1953) noted that most of the eastern U.S. during August 1953 was dominated by a strong anticyclone, a continuation of the hot and dry conditions that had started in June and persisted through July. Klein (1953) moved towards producing a wave energy analysis of the overall circulation. Klein (1953) hypothesized that the re-intensification of the ridge, and therefore the surface anticyclone, may have occurred due to barotropic energy dispersion from over the Pacific Ocean. Analysis by Klein (1953) of the antecedent conditions to the ridge building over the U.S. showed a chain of events that started in the western Pacific, and then moved eastward. The deepening of a pre-existing trough in the western Pacific was followed, a few days later, by ridge building over the eastern-central portions of the Pacific, which in turn was followed closely by the deepening of a trough over the West Coast of North America, and then by the amplification of the ridge over eastern North America. Klein (1953) showed that
a wave packet had propagated eastward with a group velocity equal to approximately 32 km h\(^{-1}\). Klein (1953) expanded upon previous research and began to study the evolution of longwave patterns that can potentially lead to heat waves. Klein (1953) also reiterated the characteristic circulation associated with heat waves: “strong ridge in the east-central Pacific, deep trough along the West Coast, and fast flat westerlies in southern Canada…” Hawkins (1954) and Winston (1954) discussed the continuation of the heat wave during July and August and agreed with Klein’s conceptual model for heat wave development, thus reinforcing most of his conclusions.

A climatological analysis that concerned aspects of the 1952–1954 drought was published by Namias (1955). Namias (1955) observed the remarkable consistency of the temperature and precipitation patterns as well as the hemispheric circulations during the summers of 1952–1954. He demonstrated that the 700-hPa mean geopotential heights and anomalies for the period are the classic ridge–trough–ridge pattern over the eastern Pacific and the U.S. He also concluded that the anomalous easterly flow found on the south side of the anticyclone in the southeast U.S. would prevent moist air from leaving the Gulf of Mexico and penetrating into the eastern portions of the U.S., leading to the inhibition of cloud formation and precipitation near the anticyclone.

Another result of Namias’s (1955) work came with his planetary wave analysis. It had already been understood that individual parts of the planetary wave system should not be considered independent features. Namias (1955) stated that all parts of the circulation are interrelated with each other, and that
every feature has an influence on every other feature of the entire pattern. It also had been known that certain long-lasting, persistent patterns were possible in the planetary wave system. Although dynamical theories had not been developed enough to explain planetary wave behavior over a period of more than a few days, Namias (1955) used several empirical approaches to help resolve whether a persistent wave pattern played a role in determining the circulation seen over the U.S. during the summers of 1952–1954. A key approach was the use of 5-day mean 700-hPa geopotential height plots for those summers characterized by large geopotential height anomalies in one specific region of the world. From these plots, it was then possible to determine the probabilities of either positive or negative anomalies in other regions across the NH for the same period, thereby revealing various teleconnective properties of the planetary wave system.

Namias (1955) focused on the eastern North Pacific Ocean for the summers of 1952–1954 where a stronger-than-normal 700-hPa anticyclone was located, a major characteristic of the pattern for that period. Namias’s (1955) map shows the probability of the signs of the 700-hPa geopotential height anomaly across the entire NH when positive geopotential height anomalies are found over the Eastern Pacific (Fig. 1.8). Namias (1955) found that when positive geopotential height anomalies exist in the Eastern Pacific, there is a high probability of finding lower-than-normal geopotential heights along the West Coast, and higher-than-normal geopotential heights in the northeast region of the U.S. This result shows that a strong eastern Pacific anticyclone has a
downstream effect on the circulation over North America by reinforcing the continental anticyclone over North America, as seen in the early 1950’s.

Both summers of 1980 and 1988 saw record levels of heat in the U.S. on par with what had taken place in the 1950’s. As described by Livezey (1980), the 1980 heat wave showed remarkable consistency with previous research. Namias (1982) notes ridge amplification over the eastern North Pacific, the deepening of a trough along the West Coast, a strong ridge over the Plains, and an East Coast trough, a similar pattern to the summers of 1952–1954. Namias (1982) continued to study teleconnective relationships by showing cross correlations of 700-hPa geopotential heights built from a 26-year dataset (1947–1972). Namias (1982) shows the teleconnective behavior between anomalously positive geopotential heights at one specific point (50°N, 150°W) and all other points on a 5° x 5° grid over the NH (Fig. 1.9a). It can be seen that if positive geopotential height anomalies exist south of the Aleutian Islands, that it is very likely that positive geopotential heights anomalies will also occur over central and eastern North America, and that negative geopotential height anomalies likely will occur over the West Coast.

Using a similar procedure but a different reference point (50°N, 30°W) (Fig. 1.9b), Namias (1982) also found that a similar anomalous 700-hPa anticyclone can occur over North America when there is a positive geopotential height anomaly over the North Atlantic. These results agree with Namias’ (1955) previous results on the behavior of anomalous midlevel circulation patterns associated with heat waves. Namias (1982) indicates yet again that persistent
long-lived heat wave patterns are a product of quasi-stationary waves in the westerlies. These results match the earlier work done in the 1950's and the pattern found with the 1980 heat wave by Livezey (1980) and Namias (1982). Lyon and Dole (1995) identified the source of anomalous stationary wave activity to be a location south of the Aleutian Islands, which is in agreement with the findings of Livezey (1980) and Namias (1982), and with what Namias characterized as the rearrangement of vorticity from a rapidly developing ridge in the east Pacific.

The 1988 case varies slightly from the standard pattern found in the westerlies during the aforementioned heat waves. Although an unusually strong 700-hPa anticyclone exists over central North America and troughs are located over each coast, 700-hPa geopotential height anomalies over the Atlantic Ocean are weakly positive, and the strong 700-hPa anticyclone normally located in the eastern North Pacific has been displaced southward. The conclusions of Namias (1991), Lyon and Dole (1995), and Chen and Newman (1998), are that, although slightly different, the 1988 circulation does present characteristics similar to previous heat wave events: a circulation that has been configured into the classic ridge–trough–ridge pattern over North America by upstream energy, the origin of which remains a matter of speculation.

Other relevant heat wave climatologies include the work of Erickson (1983), who set out to study hemispheric anomalies and their association with mean summer temperatures over the United States; Namias (1983), who studied four abnormally warm summers, 1952–1954, and 1980, and found, to a first-
order approximation, similarities to the standard three cell conceptual model discovered years earlier; and Chang and Wallace (1987), who analyzed the meteorological conditions prevalent during heat wave–droughts over the U.S. As Klein (1952a) had done previously, Chang and Wallace (1987) attempted to correlate meteorological conditions with summer (June, July, August) surface temperatures at Kansas City, MO, in order to identify circulation patterns associated with above normal warmth in the Great Plains. Chang and Wallace (1987) discovered that those months in Kansas City associated with above normal temperatures are generally characterized by upper-level anticyclones positioned over the central U.S. Chang and Wallace (1987) also showed that above-normal monthly mean surface temperatures at Kansas City exhibit strong positive correlations with anomalously positive 500-hPa geopotential heights and anomalously large 1000–500-hPa thicknesses over the U.S. (Figs. 1.10a,b). These large positive correlations are part of a wave-like pattern in 500-hPa geopotential height pattern, also reflected in the thickness field, across the NH, with strong positive correlations additionally located over the eastern North Pacific and the North Atlantic while strong negative correlations are found over the west coast of North America. In fact, Chang and Wallace (1987) note that the center areas, or largest positive and negative correlations, are similar in location to those areas of positive and negative geopotential height anomalies found by Klein (1952) and Namias (1982) in their 700-hPa teleconnection studies.
1.2.3 Short-lived heat waves

Research done on the topic of short-lived, synoptic-scale heat waves has been limited to date. McQueen and Shellum (1956) were one of the first to publish a paper on a heat wave of short duration with their work on the heat wave of 9–13 June 1956 in the Northern Plains region of the U.S. McQueen and Shellum (1956) determined that the heat wave showed remarkable consistency with those that can persist over an entire season, with only the “stagnation of…features [being] essential for…long-period heat waves.” The trigger for this event appeared to be a case of downstream development due to the dispersion of energy from an amplifying ridge over the eastern Pacific just south of the western Aleutian Islands. This development resulted in the deepening of a trough off of the West Coast, the amplification of the continental ridge over the Northern Plains and Central Canada, and the deepening of a trough along the East Coast. McQueen and Shellum (1956) documented this energy dispersion clearly in a Hovmöller diagram (Fig. 1.11). They also noted that the location of the extreme heat was farther north and west of where it was located during the 1952–54 heat waves. McQueen and Shellum (1956) concluded that the long-wave trough positioned west of the Pacific Coast resulted in the displacement of the long-wave ridge further east, advecting warm air into the Northern Plains and central Canada. The work done by McQueen and Shellum (1956) not only bridged the gap between long-lived and short-lived heat waves, but also between heat waves that occur in different regions of the U.S. as well, showing that the
“Klein model” was still relevant. Green (1956), in a *MWR* summary of June 1956 weather, also noted the strong resemblance between the composite map created by Klein (1952a) (the Kansas City method) and the 700-hPa mean geopotential height field for the first half of June 1956.

Livezey and Tinker (1996) detailed aspects of the mid-July 1995 heat wave across the Midwest and northeastern portions of the U.S. The part of Livezey and Tinker (1996) relevant to this study is the NH long-wave analysis which shows that this heat wave, manifested once again by an intense upper-level ridge and 500-hPa closed anticyclone over the U.S., is a byproduct of a large-scale downstream flow adjustment in the westerlies. Five-day mean 500-hPa geopotential height maps prepared by Livezey and Tinker (1996) show the progression of the long-wave pattern. Livezey and Tinker (1996) determined that intense convection off of the east Asian coastline, found from outgoing longwave radiation (OLR) anomalies, considerably raised geopotential heights aloft, creating a perturbation and wave train that produced the teleconnective ridge–trough–ridge pattern across the eastern Pacific and North America that is so closely associated with heat waves. Although long-lived and short-lived heat waves differ in duration, the triggers for both appear to be similar with the principal difference being the persistence of the large-scale flow pattern.

An examination of heat waves in southeastern Europe by Brikas et al. (2006), that included a case study of the 5–9 July 1988 Balkan heat wave, illustrated the synoptic-dynamic features associated with short-term heat waves. Brikas et al. (2006) determined that the hot air associated with heat waves in
southeastern Europe could be attributed to adiabatic heating from tropospheric subsidence due to 200-hPa mass convergence along the equatorward exit region of an anticyclonically curved jet streak. Brikas et al. (2006) concluded that conditions favorable for intense heat waves in southeastern Europe occur when the subtropical jet stream is displaced northward, is curved anticyclonically, and has a jet streak embedded upstream.

Galarneau et al. (2008) presented case studies alongside their global closed anticyclone climatology so as to illustrate a subset of CAs that can generate significant heat waves. These case studies illustrate that heat waves can occur when cT air masses are advected from their source regions by anomalously strong westerlies subsequent to additional warming from enhanced, dynamically driven subsidence due to upper-level ridge amplification.

Work by Galarneau et al. (2008) and Galarneau and Bosart (2006) found common signatures between the 10–15 July 1995 U.S. heat wave and the 1–22 February 2004 Australian heat wave. Galarneau et al. (2008) attributed the 10–15 July 1995 U.S. heat wave to the amplification and eastward progression of a ridge over the southwest U.S. in response to a deepening trough upstream over the West Coast. Galarneau et al. (2008) showed that the initial ridge amplification led to the development of a closed 500-hPa anticyclone that, in turn, intensified a cT air mass located over the Rocky Mountains and Mexican Plateau that was then advected northward. Later, eastward advection of the cT air occurred as the amplified long-wave pattern evolved into a more zonal, progressive long-wave pattern. Backward trajectories, calculated using
reanalysis data, illustrate nicely how cT air was advected from the Rocky Mountains towards the northern Great Plains. Galarneau et al. (2008) and Galarneau and Bosart (2006) showed that an upper-level jet streak developed between the strong ridge over the west-central U.S. and a strong trough over Hudson Bay, placing the northern Great Plains and Great Lakes region underneath the equatorward jet exit region, a favorable region for subsidence and a likely contributor to the further heating of the cT air mass.

Galarneau et al. (2008) and Galarneau and Bosart (2006) found the 1–22 February 2004 Australian heat wave to be comparable to the 10–15 July 1995 U.S. heat wave. Galarneau et al. (2008) showed that an upper-level ridge had amplified downstream of an anomalously deep trough over western Australia. The amplified ridge, with a closed anticyclone at 500 hPa, progressed eastward as the longwave pattern shifted. The strong upper-level ridge coupled with a deep trough strengthened the jet stream and positioned it such that eastern Australia was located underneath the equatorward jet exit region. The supposition by Galarneau et al. (2008) was that warming due to subsidence on the anticyclonic shear side of the strong jet enhanced the already hot cT air mass near the surface that had been produced from the sensible heating and drying of maritime tropical (mT) air that has its origins from over the Coral Sea.

Galarneau et al. (2008) and Galarneau and Bosart (2006) concluded that both the 10–15 July 1995 U.S. and 1–22 February 2004 Australian heat waves exhibited the following characteristics: the intensification of cT air masses in response to dynamically driven ridge building aloft over elevated arid and
semiarid landscapes, hot cT air then advected eastward away from its source region along the equatorward flank of a strong jet stream subsequent to the initial ridge building, with additional warming due to subsidence on the equatorward side of an strengthened jet stream likely leading to further warming.

1.2.4 Trends in U.S. heat wave frequency

Research of trends in extreme heat and heat waves has increased in recent years as global warming has become a focus of interest in the atmospheric science community. Gaffen and Ross (1998, 1999) compiled a daily temperature climatology for 188 first-order U.S. weather stations. For each station, Gaffen and Ross (1998) calculated the 85th percentiles of daily maximum, minimum, average and apparent temperatures during the period 1961–1990, which were then used to establish a threshold for extreme heat events. They showed that decadal trends for the period 1949–1995 in temperature and apparent temperature, and in 3- and 4-day heat waves, (defined as three or four days of the consecutive occurrence of extreme apparent temperatures), were positive across the entire U.S. The increased number of heat waves was seen in both the eastern and western portions of the U.S., with an approximate overall 88% increase in heat wave frequency from 1949 through 1995.

Gaffen and Ross (1999) found that summer season (JJA) temperature trends over most of the U.S. were positive by several tenths of a degree Celsius
per decade. Within the contiguous U.S., they determined that the largest temperature trends were located in the Northeast, Northwest, and Southwest regions, while the smallest were found in the South Central (Central and Southern Plains) and High Plains regions. Gaffen and Ross (1998, 1999) caution that their results may be biased to some degree by instrument changes and urbanization, although results suggest region-wide consistency.

Robinson (2001) wanted to better define what a heat wave was since a generally accepted definition did not exist. The settled-upon definition established that a heat wave occurs when, in at least a 48 h period, “neither the overnight low nor the daytime high [heat index] falls below the NWS heat stress thresholds (26.7°C and 40.6°C, respectively).” Robinson (2001) then attempted to quantify trends in decadal heat wave frequency from the 1950’s to the 1980’s. Although self-described as preliminary, his investigation determined that the Great Lakes, Southwest, Atlantic, and eastern Gulf of Mexico regions all showed an increased frequency of heat waves, while the interior South and western Gulf regions showed a decrease.

DeGaetano and Allen (2002), using 95th percentile maximum temperature trends for the contiguous U.S., showed a negative trend in maximum temperatures for the period 1930–1996 followed by a positive trend in maximum temperatures for the period 1960–1996. This change in maximum temperature trend from negative to positive around 1960 is seen in both the eastern and central portions of the U.S., while the western region exhibits a generally positive trend throughout the time period 1930–1996 (Fig. 1.12a). DeGaetano and Allen
(2002) also detailed the trends seen in heat waves, as defined as three or more consecutive days of 95\textsuperscript{th} percentile maximum temperatures, for the same regions. From region to region, the heat wave trends mirror exactly what was seen in the maximum temperature trends (Fig. 1.12b).

1.3 Goals

Based upon the above literature review, it is clear that a climatology of short-lived heat waves does not exist for the northeast U.S. In addition, the frequency analysis of both extremely hot days and heat waves are recent and the definitions of both anomalously warm days and heat waves are relatively fluid. In view of these problems, and building upon previous research, it is the goal of this thesis to construct a northeast U.S. heat wave (NHW) climatology, detailing both the antecedent conditions to, and the synoptic evolution of, heat waves. It is also the goal of this thesis to perform a frequency analysis of anomalously hot days and heat waves for the entire U.S. in order to verify previous research.

This thesis is arranged in the following manner: Chapter 2 will detail the sources of the data as well as the analysis methodology. Chapter 3 will present the results of the statistical analysis and the synoptic climatology. Chapter 4 consists of discussion and interpretation of the results from the previous chapters. Lastly, Chapter 5 will conclude the thesis and propose topics for future research.
FIG. 1.1. Percentage of (a) June, (b) July, and (c) August days in which 500-hPa anticyclones were observed in a given 10° x 10° quadrangle, 1950–1985 [from Parker et al. (1989)].
FIG. 1.2. Total number of analysis periods (twice daily) in which distinct closed anticyclone centers at 500 hPa are located in a given $2^\circ \times 5^\circ$ grid box for meteorological summer. Contour interval is 6, except 3 where dashed [from Bell and Bosart (1989)].
FIG. 1.3. Total number of (a) 200-hPa closed anticyclone events 1236 dam or greater (shaded according to the color bar) and selected mean 200-hPa geopotential height contours (solid; 1188, 1212, and 1236 dam) and (b) 500-hPa closed anticyclone events 588 dam or greater (shaded according to the color bar) and selected mean 500-hPa geopotential height contours (solid; 564, 576, and 588 dam) in the Northern Hemisphere during the period 1950–2003. [From Galarneau et al. (2006).]
FIG. 1.4. Monthly mean 700-hPa geopotential height (dam) for (a) June, (b) July, and (c) August; monthly temperature departures from long-term (1895–1980) mean in the upper five percentile (dark shading) and in the upper five to ten percentile (light shading) for (d) June, (e) July, and (f) August [from Karl and Quayle (1981)].
FIG. 1.5. Mean 700 hPa chart for the period June 1–29, 1952. Contours at 200-ft intervals are shown by solid lines, intermediate contours by lines with long dashes, and 700-hPa geopotential height departures from normal at 100-ft intervals by line with short dashes. Anomaly centers and contours are labeled in tens of feet. Areas with anomalies in excess (less than) +(-) 100 ft are hatched (stippled) [from Klein (1952a)].
FIG. 1.6. Mean 700-hPa chart for the ten 5-day periods with largest positive surface temperature anomalies at Kansas City, MO, during the summer 1947–1951. Contours at 200-ft intervals are shown by solid lines, intermediate contours by lines with long dashes, and 700-hPa geopotential height departures from normal at 100-ft intervals by line with short dashes. Anomaly centers and contours are labeled in tens of feet. Areas with anomalies in excess (less than) +(-) 100 ft are hatched (stippled) [from Klein (1952a)].
FIG. 1.7. Simple linear correlation coefficient between 5-day mean surface temperature anomalies at Kansas City, MO, in summer and 5-day mean 700-hPa geopotential height anomalies. Contoured at 0.20 intervals [from Klein (1952a).]
FIG. 1.8. The probability of the sign of 700-hPa geopotential height anomalies when a strong positive anomaly lies within the vicinity of the hatched region of the Pacific Ocean for the meteorological summer [from Namias (1955)].
FIG. 1.9. Cross correlations between 700-hPa geopotential height and a given point: (a) 50°N, 150°W; (b) 50°N, 30°W.
FIG. 1.10. Monthly mean surface air temperature at Kansas City, MO, correlated with (a) 500-hPa geopotential height, (b) 1000–500-hPa thickness. Correlations based upon anomalies for 93-month dataset for meteorological summer during the period 1946–1976. Contoured at 0.1, negative contours dashed [from Chang and Wallace (1987)].
FIG. 1.11. Hovmöller diagram showing a cross-sectional view at 500 hPa for 50°N, extending from 150°E to 5°W during the period from 0300 UTC 2 June to 1500 UTC 14 June 14 1956. Height values are in hundreds of feet, ridge areas are stippled, and trough areas are hatched. Heavy black line represents trajectory of downstream energy dispersion [from McQueen and Shellum (1956)].
FIG. 1.12a. Composite extreme maximum temperature exceedance series for the contiguous United States (top), eastern (second row), central (third row), and western regions (bottom row). ‘Eastern’ refers to stations east of 95°W and ‘western’ to stations west of 110°W. Linear trendlines are given for the 1930–96 and 1960–96 periods [from DeGaetano and Allen (2002)].

FIG. 1.12b. As in FIG. 1.12a above but for composite ≥ 3 day extreme maximum temperature exceedance events [from DeGaetano and Allen (2002)].
2. Data and Methodology

2.1 Data

Temperature data were obtained from the National Climatic Data Center (NCDC) high-resolution surface dataset (#3210 – U.S. Summary of the Day, First Order Data). Daily high temperatures were extracted for 54 surface stations (Tables I and II) for a 54-year period (1948–2001). Stations were selected on the basis of both dataset continuity and coverage within NCDC Standard Regions for temperature and precipitation (Fig. 2.1). In addition, sea-level pressure (SLP) as well as 1000-, 850-, 700-, 500- and 200-hPa geopotential height, wind, and temperature data from the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis dataset are used to create composite analyses. (Kalnay et al. 1996) The NCEP–NCAR reanalysis data are archived on a global $2.5^\circ \times 2.5^\circ$ latitude–longitude grid and available on 17 isobaric levels from 1000 hPa to 10 hPa starting from 1 January 1948.

2.2 Methodology

2.2.1 Initial methodology

Conventional meteorological wisdom states that temperature data are distributed normally (e.g., Harmel et al. 2002). Accordingly, the current study defined an anomalously hot day as any day the recorded high temperature was
at least two standard deviations (σ) above the 54-year mean daily maximum temperature. A 31-day centered moving average was used to smooth the 54-year mean maximum temperatures and σ for each calendar day. Leap days (29 February) were analyzed exclusively over the 14 leap years in the dataset.

By definition, 2σ represent 95% of the total population of a normally distributed dataset and data above +2σ represents 2.5% of the dataset. Firstly, a complete, normally distributed daily 54-year dataset (1948–2001) should contain 19724 total days and approximately 493 days should be anomalously hot days (≥ +2σ). Surprisingly, the number of anomalously hot days at most stations varied, often times considerably, from the expected value. Los Angeles, CA, had the highest number of anomalously hot days at 989, whereas Denver, CO, had the lowest number of anomalously hot days at 49. Frequency distributions of daily high temperatures show non-normally distributed (skewed) datasets, with Los Angeles (Fig. 2.2) being the most positively skewed station and Denver (Fig. 2.3) being most the negatively skewed station.

Secondly, in a normally distributed dataset, the number of data points above +2σ equals the number of data points below −2σ. However, the ratio of anomalously hot days to anomalously cold (≥−2σ) days at most stations was not 1:1. This ratio is clearly seen in Tables III and IV, which show the 10 most positively skewed and negatively skewed surface stations in the dataset, respectively. Thus, the +2σ threshold method is ineffective in identifying anomalously hot days because most stations have skewed data distributions and σ is not a reliable indicator of anomaly.
2.2.2 Physical explanation for large temperature anomalies

The Los Angeles and Denver high temperature datasets are significantly skewed. In order to understand the cause of the skewness, composites were created to depict the synoptic patterns associated with the greatest outlier temperature events. The 25 most positively (negatively) skewed days for Los Angeles (Denver) were identified for compositing purposes (see Table V). Composite sea level pressure (SLP) plots for these dates show distinct physical signatures for significant anomalous temperature events at each station (Figs. 2.4 and 2.5).

Los Angeles is positioned to the east of the Pacific Ocean, where a marine layer prominently influences the climate and daily high temperatures are typically suppressed. Composites of the 25 most positively skewed meteorological summer dates show that extreme high temperature events at Los Angeles are associated with downslope (Santa Ana) winds (Fig. 2.4). These winds result from the pressure gradient between a surface low centered over Baja California and a region of high pressure located over the borders of Utah, Wyoming, and Colorado. This adiabatic-warmed wind is also an offshore wind at Los Angeles and acts to scour out the marine layer there. In addition, a rudimentary streamline analysis (not shown) indicates that the winds likely originate from the Mexican plateau, a source region for hot and dry air. Santa Ana events produce weather at Los Angeles that is significantly different from climatology, resulting in daily high temperatures that are well above normal.
Denver is located on the front range of the Rocky Mountains, and its climate is influenced predominantly by westerly downsloping winds. The composite SLP plot for the 25 most anomalously cold meteorological summer dates (Fig. 2.5) shows that the pressure gradient at the surface generates easterly winds, the result of a high centered over eastern Montana and a region of low pressure located over northwestern Mexico and the southwestern U.S. Easterly surface winds at Denver are upslope winds, resulting in adiabatic cooling and likely associated with cloudiness and precipitation. Upslope events produce weather at Denver that is significantly different from climatology, resulting in daily high temperatures that fall well below normal.

2.2.3 Revised methodology

Instead of the $\sigma$ method, a percentile method was selected to identify anomalous temperature events. The 97.5-percentile threshold of a non-normally distributed dataset approximates the $2\sigma$ threshold of a normally distributed dataset, but is based upon the median, not the mean, and thus accounts for a skewed distribution. An anomalously hot day was defined as a day with a high temperature greater than or equal to the 31-day centered moving average of the calculated daily 97.5-percentile. The new technique identified 669 anomalously hot days at Los Angeles and 736 anomalously hot days at Denver. Recall that the number of anomalously hot days using the $2\sigma$ technique identified 989 anomalously hot days at Los Angeles and 49 anomalously hot days at Denver.
The number of anomalously hot days identified at the remaining stations, using the new technique, was similar and ranged from 661 days to 796 days. Results of the improved methodology can be visualized in Figs. 2.6 and 2.7. For Los Angeles (Fig. 2.6) during meteorological summer, the red line representing the 97.5-percentile threshold is above the orange line representing the $+2\sigma$ threshold, meaning that there will be fewer anomalously hot days identified using the former criterion. For Denver (Fig. 2.7) during meteorological summer, the red line representing the 97.5-percentile threshold falls below the orange line representing the $+2\sigma$ threshold, meaning that more anomalously hot days will be identified using the former criterion.

Once a definition was accepted for anomalously hot days, the next step was to define a heat wave. A *heat wave* was defined as three or more consecutive anomalously hot days at a single station, whereas a *regional heat wave* (e.g., northeast U.S. heat wave) was defined as concurrent heat waves at two or more stations within a NCDC standard region (Fig. 2.1) with a minimum of a one-day overlap between the individual heat waves.

2.2.4 Frequency analysis

An annual frequency analysis was performed to determine temporal trends in both anomalously hot days and regional heat waves for the period 1948–2001. This analysis was done for each NCDC standard region (Fig. 2.1) as well as for the contiguous U.S. In addition, a comparison between those
standard regions lying east of the Rocky Mountains versus those lying west of the Rocky Mountains was performed so as to determine possible intra-continental variability. To evaluate the statistical significance of trends in the frequency of anomalously hot days, coefficients of determination ($R^2$ values) were calculated using Microsoft Excel standard linear analysis. Only $R^2$ values exceeding 0.2 were considered to be statistically significant. The trendline analysis focused on three time periods: 1948–1959, 1960–2001, and 1948–2001. These periods were chosen based on previous research by DeGaetano and Allen (2002) who discovered the presence of a significant negative-to-positive trend changepoint in 95th percentile daily maximum temperatures across the U.S. around 1960. Additionally, Gaffen and Ross (1999) used the time series 1961–1996 in their examination of regional decadal temperature trends for meteorological summer.

2.2.5 Composite plots

All of the composite analyses used in this study were generated using the General Meteorology Package (GEMPAK 5.11.4) analysis software [updated from the original package devised by Koch et al. (1983)]. Of the 24 total northeast U.S. heat waves (NHWs) identified, 17 were chosen for compositing based upon similarities in the 500-hPa flow over North America (see Table VI). The criterion used in choosing the 17 cases was subjective: all cases consisting of a west-to-east trough–ridge pattern over the U.S. during the heat wave
midpoint were chosen. The compositing focused on two aspects of an NHW: the antecedent conditions and the evolution.

Composites showing the antecedent conditions consisted of 3-day averages of geopotential height and anomaly at 200 hPa and 700 hPa, as well as of sea level pressure overlain with 1000–500-hPa thickness and anomaly over North America and the Northern Pacific Basin. These pressure levels and variables were chosen to evaluate the conditions upstream of the U.S. and to determine how the upper-, mid-, and lower-levels of the troposphere each contributed to the evolution of the upper-level flow pattern. The 3-day averages for the composites were centered on 7 (D-7), 5 (D-5), 3 (D-3), and 1 (D-1) day(s) prior to onset. Height and thickness anomalies were calculated using a 1968–1996 climatology generated by the Climate Diagnostics Center (CDC).

Composites illustrating the structure and evolution of an NHW are also included. The upper-level flow pattern and dynamics can be seen in the composites of 200-hPa geopotential height, wind, and divergence and in the composites of 500-hPa geopotential height, wind, and absolute vorticity. The midtropospheric flow pattern and vertical motion are illustrated in the composites of 700-hPa geopotential height, vertical velocity, and relative humidity. Lower-tropospheric flow and implied thermal advection are shown in the composites of 850-hPa geopotential height, wind, and temperature. The upper-level flow pattern and surface pressure field as well as lower-tropospheric implied thermal advection can be seen in the composites of sea level pressure overlain with 1000–500-hPa thickness and 200-hPa winds. The mid- and lower-level thermal
structure of the atmosphere is depicted by the 700–500-hPa and 925–700-hPa lapse rates, 850-hPa isotherms, and 1000–500-hPa thickness.

Again, height and thickness anomalies were calculated using a 1968–1996 climatology generated by the CDC. All heat waves were standardized for length so that heat waves of various durations could be compared. These composites were generated at the standardized points of 4 (D.4), 3 (D.3), 2 (D.2), and 1 (D.1) day(s) prior to the onset of a heat wave, as well as on the first day (D0), midpoint (DM), and last day (DL) of a heat wave. In addition, a composite time series was created showing the synoptic evolution of the 18°C isotherm at 850 hPa and the 23°C and 24°C isotherms at 925 hPa over North America both prior to and during an NHW. These isotherms were used to illustrate the advection and/or development of warm air in the lower-troposphere and were chosen over other isotherms for their clarity.
<table>
<thead>
<tr>
<th>City</th>
<th>State</th>
<th>Station Name</th>
<th>Code</th>
</tr>
</thead>
<tbody>
<tr>
<td>Albany</td>
<td>NY</td>
<td>Albany International Airport</td>
<td>ALB</td>
</tr>
<tr>
<td>Albuquerque</td>
<td>NM</td>
<td>Albuquerque International Airport</td>
<td>ABQ</td>
</tr>
<tr>
<td>Amarillo</td>
<td>TX</td>
<td>Amarillo International Airport</td>
<td>AMA</td>
</tr>
<tr>
<td>Atlanta</td>
<td>GA</td>
<td>Atlanta-Hartsfield International Airport</td>
<td>ATL</td>
</tr>
<tr>
<td>Beaufort/Port Arthur</td>
<td>TX</td>
<td>Port Arthur Regional Airport</td>
<td>BPT</td>
</tr>
<tr>
<td>Billings</td>
<td>MT</td>
<td>Billings Logan International Airport</td>
<td>BIL</td>
</tr>
<tr>
<td>Bismarck</td>
<td>ND</td>
<td>Bismarck Municipal Airport</td>
<td>BIS</td>
</tr>
<tr>
<td>Boise</td>
<td>ID</td>
<td>Boise Air Terminal</td>
<td>BOI</td>
</tr>
<tr>
<td>Boston</td>
<td>MA</td>
<td>Boston-Logan International Airport</td>
<td>BOS</td>
</tr>
<tr>
<td>Caribou</td>
<td>ME</td>
<td>Caribou Municipal Airport</td>
<td>CAR</td>
</tr>
<tr>
<td>Charleston</td>
<td>SC</td>
<td>Charleston International Airport</td>
<td>CHS</td>
</tr>
<tr>
<td>Cheyenne</td>
<td>WY</td>
<td>Cheyenne Municipal Airport</td>
<td>CYS</td>
</tr>
<tr>
<td>Chicago</td>
<td>IL</td>
<td>Chicago-Midway Airport</td>
<td>MDW</td>
</tr>
<tr>
<td>Corpus Christi</td>
<td>TX</td>
<td>Corpus Christi International Airport</td>
<td>CRP</td>
</tr>
<tr>
<td>Covington</td>
<td>KY</td>
<td>Cincinnati-Northern Kentucky Airport</td>
<td>CVG</td>
</tr>
<tr>
<td>Denver</td>
<td>CO</td>
<td>Stapleton International Airport</td>
<td>DEN</td>
</tr>
<tr>
<td>Des Moines</td>
<td>IA</td>
<td>Des Moines International Airport</td>
<td>DSM</td>
</tr>
<tr>
<td>Dodge City</td>
<td>KS</td>
<td>Dodge City Regional Airport</td>
<td>DDC</td>
</tr>
<tr>
<td>El Paso</td>
<td>TX</td>
<td>El Paso International Airport</td>
<td>ELP</td>
</tr>
<tr>
<td>Ely</td>
<td>NV</td>
<td>Ely Airport/Yelland Field</td>
<td>ELY</td>
</tr>
<tr>
<td>Erie</td>
<td>PA</td>
<td>Erie International Airport</td>
<td>ERI</td>
</tr>
<tr>
<td>Great Falls</td>
<td>MT</td>
<td>Great Falls International Airport</td>
<td>GTF</td>
</tr>
<tr>
<td>International Falls</td>
<td>MN</td>
<td>International Falls International Airport</td>
<td>INL</td>
</tr>
<tr>
<td>Las Vegas</td>
<td>NV</td>
<td>Las Vegas McCarran International Airport</td>
<td>LAS</td>
</tr>
<tr>
<td>Little Rock</td>
<td>AR</td>
<td>Little Rock National Airport/Adams Field</td>
<td>LIT</td>
</tr>
<tr>
<td>Los Angeles</td>
<td>CA</td>
<td>Los Angeles International Airport</td>
<td>LAX</td>
</tr>
</tbody>
</table>
TABLE II. Surface stations in dataset (M–Z).

<table>
<thead>
<tr>
<th>City</th>
<th>State</th>
<th>Station Name</th>
<th>Code</th>
</tr>
</thead>
<tbody>
<tr>
<td>Medford</td>
<td>OR</td>
<td>Rogue Valley International Airport</td>
<td>MFR</td>
</tr>
<tr>
<td>Meridian</td>
<td>MS</td>
<td>Key Field Airport</td>
<td>MEI</td>
</tr>
<tr>
<td>Miami</td>
<td>FL</td>
<td>Miami International Airport</td>
<td>MIA</td>
</tr>
<tr>
<td>Milwaukee</td>
<td>WI</td>
<td>Milwaukee International Airport</td>
<td>MKE</td>
</tr>
<tr>
<td>Minneapolis-St. Paul</td>
<td>MN</td>
<td>Minneapolis-St. Paul International Airport</td>
<td>MSP</td>
</tr>
<tr>
<td>Nashville</td>
<td>TN</td>
<td>Nashville International Airport</td>
<td>BNA</td>
</tr>
<tr>
<td>New Orleans</td>
<td>LA</td>
<td>New Orleans International Airport</td>
<td>MSY</td>
</tr>
<tr>
<td>New Orleans (1948–54)</td>
<td>LA</td>
<td>New Orleans International Airport</td>
<td>MSY</td>
</tr>
<tr>
<td>New York City</td>
<td>NY</td>
<td>New York-La Guardia Airport</td>
<td>LGA</td>
</tr>
<tr>
<td>Norfolk</td>
<td>VA</td>
<td>Norfolk International Airport</td>
<td>ORF</td>
</tr>
<tr>
<td>North Platte</td>
<td>NE</td>
<td>North Platte Regional Airport</td>
<td>LBF</td>
</tr>
<tr>
<td>Oklahoma City</td>
<td>OK</td>
<td>Will Rodgers World Airport</td>
<td>OKC</td>
</tr>
<tr>
<td>Phoenix</td>
<td>AZ</td>
<td>Phoenix-Sky Harbor International Airport</td>
<td>PHX</td>
</tr>
<tr>
<td>Pierre</td>
<td>SD</td>
<td>Pierre Regional Airport</td>
<td>PIR</td>
</tr>
<tr>
<td>Pittsburgh</td>
<td>PA</td>
<td>Pittsburgh International Airport</td>
<td>PIT</td>
</tr>
<tr>
<td>Raleigh-Durham</td>
<td>NC</td>
<td>Raleigh-Durham International Airport</td>
<td>RDU</td>
</tr>
<tr>
<td>Reno</td>
<td>NV</td>
<td>Reno Tahoe International Airport</td>
<td>RNO</td>
</tr>
<tr>
<td>Salt Lake City</td>
<td>UT</td>
<td>Salt Lake City International Airport</td>
<td>SLC</td>
</tr>
<tr>
<td>San Angelo</td>
<td>TX</td>
<td>San Angelo Regional Airport/Mathis Field</td>
<td>SJT</td>
</tr>
<tr>
<td>San Francisco</td>
<td>CA</td>
<td>San Francisco International Airport</td>
<td>SFO</td>
</tr>
<tr>
<td>Seattle</td>
<td>WA</td>
<td>Seattle-Tacoma International Airport</td>
<td>SEA</td>
</tr>
<tr>
<td>Spokane</td>
<td>WA</td>
<td>Spokane International Airport</td>
<td>GEG</td>
</tr>
<tr>
<td>St. Louis</td>
<td>MO</td>
<td>St. Louis-Lambert International Airport</td>
<td>STL</td>
</tr>
<tr>
<td>Tallahassee</td>
<td>FL</td>
<td>Tallahassee Regional Airport</td>
<td>TLH</td>
</tr>
<tr>
<td>Tampa</td>
<td>FL</td>
<td>Tampa International Airport</td>
<td>TPA</td>
</tr>
<tr>
<td>Topeka</td>
<td>KS</td>
<td>Topeka-Philip Billard Airport</td>
<td>TOP</td>
</tr>
<tr>
<td>Waco</td>
<td>TX</td>
<td>Waco Regional Airport</td>
<td>ACT</td>
</tr>
<tr>
<td>Washington, DC</td>
<td>VA</td>
<td>Reagan-Washington National Airport</td>
<td>DCA</td>
</tr>
</tbody>
</table>
TABLE III. Ten most positively skewed surface stations (see Tables I and II for station codes).

<table>
<thead>
<tr>
<th>Station</th>
<th>% Negative</th>
<th>% Positive</th>
</tr>
</thead>
<tbody>
<tr>
<td>LAX</td>
<td>5.7</td>
<td>94.3</td>
</tr>
<tr>
<td>SFO</td>
<td>13.8</td>
<td>86.2</td>
</tr>
<tr>
<td>ERI</td>
<td>28.8</td>
<td>71.2</td>
</tr>
<tr>
<td>BOS</td>
<td>30.7</td>
<td>69.3</td>
</tr>
<tr>
<td>SEA</td>
<td>32.2</td>
<td>67.8</td>
</tr>
<tr>
<td>LGA</td>
<td>37.7</td>
<td>62.3</td>
</tr>
<tr>
<td>MKE</td>
<td>40.5</td>
<td>59.5</td>
</tr>
<tr>
<td>CAR</td>
<td>40.8</td>
<td>59.2</td>
</tr>
<tr>
<td>ALB</td>
<td>44.5</td>
<td>55.5</td>
</tr>
<tr>
<td>ORF</td>
<td>44.5</td>
<td>55.5</td>
</tr>
</tbody>
</table>

TABLE IV. Ten most negatively skewed surface stations (see Tables I and II for station codes).

<table>
<thead>
<tr>
<th>Station</th>
<th>% Negative</th>
<th>% Positive</th>
</tr>
</thead>
<tbody>
<tr>
<td>DEN</td>
<td>94.2</td>
<td>5.8</td>
</tr>
<tr>
<td>CYS</td>
<td>91.3</td>
<td>8.7</td>
</tr>
<tr>
<td>TPA</td>
<td>90.7</td>
<td>9.3</td>
</tr>
<tr>
<td>MSY</td>
<td>88.8</td>
<td>11.2</td>
</tr>
<tr>
<td>BPT</td>
<td>87.8</td>
<td>12.2</td>
</tr>
<tr>
<td>AMA</td>
<td>87.6</td>
<td>12.4</td>
</tr>
<tr>
<td>ELY</td>
<td>87.6</td>
<td>12.4</td>
</tr>
<tr>
<td>GTF</td>
<td>86.9</td>
<td>13.1</td>
</tr>
<tr>
<td>ABQ</td>
<td>86.5</td>
<td>13.5</td>
</tr>
<tr>
<td>SJT</td>
<td>86.3</td>
<td>13.7</td>
</tr>
</tbody>
</table>
TABLE V. Twenty-five most positively and negatively skewed dates for Los Angeles, CA and Denver, CO, during meteorological summer (1948–2001) in decreasing magnitude of deviation from the mean.

<table>
<thead>
<tr>
<th>Los Angeles</th>
<th>Denver</th>
</tr>
</thead>
<tbody>
<tr>
<td>06/10/1979</td>
<td>07/03/1972</td>
</tr>
<tr>
<td>06/16/1981</td>
<td>08/10/1968</td>
</tr>
<tr>
<td>06/11/1979</td>
<td>07/29/1971</td>
</tr>
<tr>
<td>06/15/1981</td>
<td>08/03/1978</td>
</tr>
<tr>
<td>06/18/1957</td>
<td>07/04/1972</td>
</tr>
<tr>
<td>06/17/1957</td>
<td>06/25/1958</td>
</tr>
<tr>
<td>06/17/1981</td>
<td>06/27/1983</td>
</tr>
<tr>
<td>06/24/1976</td>
<td>08/24/1992</td>
</tr>
<tr>
<td>07/10/1959</td>
<td>07/12/1951</td>
</tr>
<tr>
<td>06/28/1980</td>
<td>07/12/1987</td>
</tr>
<tr>
<td>08/06/1983</td>
<td>08/14/1979</td>
</tr>
<tr>
<td>07/09/1985</td>
<td>07/10/1951</td>
</tr>
<tr>
<td>07/04/1981</td>
<td>08/02/1976</td>
</tr>
<tr>
<td>07/01/1985</td>
<td>08/03/1991</td>
</tr>
<tr>
<td>06/09/1979</td>
<td>08/09/1960</td>
</tr>
<tr>
<td>08/13/1994</td>
<td>08/10/1981</td>
</tr>
<tr>
<td>07/11/1959</td>
<td>07/21/1961</td>
</tr>
<tr>
<td>07/12/1990</td>
<td>07/05/1960</td>
</tr>
<tr>
<td>07/21/1960</td>
<td>06/29/1951</td>
</tr>
<tr>
<td>06/22/1949</td>
<td>06/24/1967</td>
</tr>
<tr>
<td>07/04/1957</td>
<td>06/18/1982</td>
</tr>
<tr>
<td>08/12/1991</td>
<td>07/31/1965</td>
</tr>
<tr>
<td>06/20/1973</td>
<td>07/31/1995</td>
</tr>
<tr>
<td>08/31/1955</td>
<td>08/14/1959</td>
</tr>
</tbody>
</table>
TABLE VI. Northeast regional heat wave composite cases.

<table>
<thead>
<tr>
<th>Dates</th>
</tr>
</thead>
<tbody>
<tr>
<td>25–29 Aug 1948</td>
</tr>
<tr>
<td>27–30 Jul 1949</td>
</tr>
<tr>
<td>27 Aug–5 Sep 1953</td>
</tr>
<tr>
<td>30 Jul–6 Aug 1955</td>
</tr>
<tr>
<td>15–18 Jun 1957</td>
</tr>
<tr>
<td>12–17 Aug 1959</td>
</tr>
<tr>
<td>25–28 Jul 1963</td>
</tr>
<tr>
<td>1–4 Jul 1966</td>
</tr>
<tr>
<td>28–31 Aug 1973</td>
</tr>
<tr>
<td>30 Jul–3 Aug 1975</td>
</tr>
<tr>
<td>19–21 Jul 1977</td>
</tr>
<tr>
<td>13–15 Jun 1988</td>
</tr>
<tr>
<td>4–10 Jul 1988</td>
</tr>
<tr>
<td>1–5 Aug 1988</td>
</tr>
<tr>
<td>23–27 Jul 1989</td>
</tr>
<tr>
<td>25–28 Aug 1993</td>
</tr>
<tr>
<td>13–15 Jul 1995</td>
</tr>
</tbody>
</table>
FIG. 2.1. NCDC standard regions for temperature and precipitation (shaded by region) and the 54 stations contained in the dataset (gray text).
FIG. 2.2. Histogram of daily high temperatures (°C) at Los Angeles, CA, for the period 1948–2001. Histogram bin sized in 3°C increments.
FIG. 2.3. As in FIG. 2.2, except for Denver, CO.
FIG. 2.4. Composite sea level pressure (hPa) plot of the 25 most positively skewed dates during meteorological summer at Los Angeles, CA (black diamond), contoured at 3-hPa intervals.
FIG. 2.5. Composite sea level pressure (hPa) plot of the 25 most negatively skewed dates during meteorological summer at Denver, CO (black diamond), contoured at 3-hPa intervals.
FIG. 2.6. Entire 54-year daily high temperature dataset (°C) at Los Angeles, CA, for the meteorological summer season in times series form, i.e., the first 54 data points represent 1 June, the next 54 data points represent 2 June, etc. (ocean blue). Also plotted are 31-day centered moving averages for the following: the 54-year daily mean high temperature (black), the daily +2 standard deviation threshold (orange), and the daily 97.5-percentile threshold (red).
FIG. 2.7. As in FIG. 2.6, except for Denver, CO.
3. Results

3.1 Statistical Analysis

3.1.1 Eastern regions

For the purpose of this research, Eastern regions are defined as those National Climatic Data Center (NCDC) standard regions located exclusively to the east of the Rocky Mountains comprising the Northeast, Southeast, Central, Southern, East North Central, and West North Central regions. In each region, as stated in section 2.2.4, the entire time series (1948–2001), as well as the periods 1948–1959 and 1960–2001, were individually scrutinized so as to allow comparison with relevant work done by Gaffen and Ross (1999) and DeGaetano and Allen (2002). Again, from section 2.2.4, trendlines were established from the data and coefficients of determination ($R^2$ values) were calculated, with only those $R^2$ values exceeding 0.2 considered to be statistically significant indicators of trend. The Northeast region is included in this work as it is the focus of this research. The Southeast region is included as it is the only Eastern region that exhibits a statistically significant trend. Composite totals for all Eastern regions are also included for comparison with other regions across the United States (U.S.)

For the Northeast region (comprising six stations), no statistically significant trends were identified in the annual frequency of anomalously hot days (Fig. 3.1a). For the period 1948–1959, a negatively sloped trendline with a statistically insignificant $R^2$ value of 0.0151 (blue line) is seen, whereas for the
period 1960–2001, a positively sloped trendline with a statistically insignificant $R^2$ value of 0.0345 (orange line) is seen (Fig. 3.1a). Overall, for the period 1948–2001, a weak, negatively sloped trendline with a statistically insignificant $R^2$ value of 0.0047 (green line) is seen (Fig. 3.1a). High frequency years for anomalously hot days are 1949, 1952–1953, 1955, 1959, 1988, 1991, 1993, and 1995. The frequency distribution of Northeast regional heat waves resembles the distribution seen in the frequency of Northeast anomalously hot days, although the lower tallies of regional heat waves do not allow for a high-confidence statistical analysis (Fig 3.1b). Peak-frequency years for Northeast regional heat waves are 1949, 1955, 1988, and 1993.

In the Southeast region (comprising eight stations), a significant trend in the frequency of anomalously hot days is seen for the period 1960–2001 (Fig. 3.2a). A positively sloped trendline with a $R^2$ value of 0.3148 (orange line) is exhibited during this period (Fig. 3.2a). Aside from this period, no other significant trends could be identified in the time series. For the period 1948–1959, a negatively sloped trendline with a statistically insignificant $R^2$ value of 0.0291 (blue line) is seen, whereas for the overall period 1948–2001 a positively sloped trendline with a statistically insignificant $R^2$ value of 0.0948 (green line) is seen (Fig. 3.2a). Peak years in the frequency of anomalously hot days are 1952, 1954, 1980, 1990, 1993, and 1998–1999. The Southeast region mirrors the Northeast region in that peak years in the frequency of anomalously hot days are also peak years in the frequency of regional heat waves (Fig. 3.2b). Relative maxima in Southeast regional heat waves occur during the late 1940’s through
the 1950's, as well as from the 1980's through the 1990's, with peaks in 1952, 1993, 1998, and 1999 (Fig. 3.2b).

Figure 3.3a illustrates the total number of anomalously hot days for all stations located in the Eastern regions (comprising 40 stations). The frequency distributions are similar to those of anomalously hot days in the Northeast and Southeast regions. Anomalously hot days are seen in the Eastern regions, with higher frequencies of anomalously hot days occurring in the early 1950's and in the 1980's through the 1990's. However, no statistically significant trends could be identified in the time series. For the period 1948–1959, a negatively sloped trendline with a statistically insignificant $R^2$ value of 0.0191 (blue line) is seen, whereas for the period 1960–2001 a positively sloped trendline with a statistically insignificant $R^2$ value of 0.0808 (orange line) is seen (Fig. 3.3a). Overall, for the period 1948–2001, a very weak, positively sloped trendline with a statistically insignificant $R^2$ value of 0.0017 (green line) is seen (Fig. 3.3a).

Figure 3.3b illustrates the total number of regional heat waves in the Eastern regions. Again, periods of increased regional heat wave activity exist during the early 1950's and during the 1980's through the 1990's. As there are more instances of regional heat waves in the Eastern regions composite than in the individual regions, a statistical analysis can be performed. For the period 1948–1959, a negatively sloped trendline with a statistically insignificant $R^2$ value of 0.1144 (blue line) is seen, whereas for the period 1960–2001 a positively sloped trendline with a statistically insignificant $R^2$ value of 0.0609 (orange line) is seen (Fig. 3.3b). Overall, for the period 1948–2001, a very weak, positively
sloped trendline is found with a statistically insignificant $R^2$ value of 0.00004 (green line) (Fig. 3.3b).

### 3.1.2 Western regions and the contiguous U.S.

Western Regions are defined as those NCDC standard regions located within and to the west of the Rocky Mountains comprising the Northwest, Southwest, and Western regions. The entire time series (1948–2001), as well as the periods 1948–1959 and 1960–2001, were individually scrutinized, so as to be consistent with the Eastern regions analysis. Also like the Eastern regions analysis, trendlines and coefficients of determination ($R^2$ values) were calculated with only those $R^2$ values exceeding 0.2 considered to be statistically significant indicators of trend. The Southwest region is included because it is representative of the Western regions as a whole, and it exhibits the most statistically significant trends. Composite totals for all Western regions are also included for comparison with other regions across the U.S.

For the Southwest region (comprising five stations), no statistically significant trends are identified in the annual frequency of anomalously hot days (Fig. 3.4a). For the period 1948–1959, a positively sloped trendline with a statistically insignificant $R^2$ value of 0.0418 (blue line) is seen, whereas for the period 1960–2001 a positively sloped trendline with a statistically insignificant $R^2$ value of 0.102 (orange line) is seen (Fig. 3.4a). Overall, for the period 1948–2001, a positively sloped trendline with a statistically insignificant $R^2$ value of
0.0296 (green line) is seen (Fig. 3.4a). Peak years in the frequency of anomalously hot days include 1960, 1979–1981, 1990, and 1994. Figure 3.4b illustrates the frequency of Southwest regional heat waves for the period 1948–2001. The frequency of regional heat waves increases over time as manifested in the fact that during the first half of the time series, there are five regional heat waves, while during the second half of the time series, there are 14 regional heat waves (Fig. 3.4b).

In the composite totals for the Western regions (comprising 14 stations), a significant trend in the frequency of anomalously hot days is seen for the period 1948–1959, as manifested in the positively sloped trendline with a $R^2$ value of 0.5715 (blue line) (Fig. 3.5a). Aside from this period, no other significant trends could be identified in the time series. For the period 1960–2001, a positively sloped trendline with a statistically insignificant $R^2$ value of 0.0145 (orange line) is seen, whereas for the overall period 1948–2001, a positively sloped trendline with a statistically insignificant $R^2$ value of 0.1199 (green line) is seen (Fig. 3.5a).

Figure 3.5b illustrates the total number of regional heat waves in the Western regions. As there are more instances of regional heat waves in the Western regions composite than in the individual regions, a statistical analysis can be performed. Although no significant trends could be identified in the time series, an increased frequency of regional heat waves is apparent, with more regional heat waves occurring during the 1980's and 1990's than in previous decades. For the period 1948–1959, a positively sloped trendline with a statistically insignificant $R^2$ value of 0.0236 (blue line) is seen, whereas for the
period 1960–2001 a positively sloped trendline with a statistically insignificant $R^2$ value of 0.005 (orange line) is seen (Fig. 3.5b). Overall, for the period 1948–2001, a positively sloped trendline with a statistically insignificant $R^2$ value of 0.0601 (green line) is seen (Fig. 3.5b).

Figure 3.6a illustrates the total number of anomalously hot days for the regions within the contiguous U.S. (comprising 54 stations). No statistically significant trends could be identified in the annual frequency of anomalously hot days. As with the Eastern regions, there are two periods of increased numbers of anomalously hot days during the early 1950's and the late 1980's through the 1990's, respectively. For the period 1948–1959, a negatively sloped trendline with a statistically insignificant $R^2$ value of 0.0024 (blue line) is seen, whereas for the period 1960–2001 a positively sloped trendline with a statistically insignificant $R^2$ value of 0.0898 (orange line) is seen (Fig. 3.6a). Overall, for the period 1948–2001, a positively sloped trendline with a statistically insignificant $R^2$ value of 0.0233 (green line) is seen (Fig. 3.6a). Nationwide, peak years include 1952–1954, 1980, 1988, 1990, 1995, and 1998.

Figure 3.6b illustrates the total number of regional heat waves within the contiguous U.S. Although no statistically significant trends could be identified, there is agreement between the frequencies of regional heat waves in the contiguous U.S. and in the Eastern regions. Again, periods of increased regional heat wave activity exist in the early 1950's and in the 1980's through the 1990's. For the period 1948–1959, a negatively sloped trendline with a statistically insignificant $R^2$ value of 0.1036 (blue line) is seen, whereas for the period 1960–
2001 a positively sloped trendline with a statistically insignificant $R^2$ value of 0.0642 (orange line) is seen (Fig. 3.6b). Overall, for the period 1948–2001, a weak, positively sloped trendline with a statistically insignificant $R^2$ value of 0.0087 (green line) is seen (Fig. 3.6b). Nationwide, peak years include 1952–1954, 1980, 1988, and 1998.
3.2 Northeast U.S. Heat Wave Composites

3.2.1 Upstream antecedent conditions

The purpose of this section is to describe the time series composites of the larger-scale upstream antecedent conditions that lead to a northeast U.S. heat wave (NHW). These composites nicely illustrate the evolution of the upper-level, mid-tropospheric, and lower-tropospheric flow patterns over the North Pacific Ocean Basin and North America prior to NHW initiation.

As detailed in section 2.2.5, the composites of upstream antecedent conditions are three-day averages centered on D−7, D−5, D−3, and D−1. Of the 24 NHWs identified from the National Climatic Data Center (NCDC) temperature dataset, 17 are chosen for compositing using a simple subjective criterion; namely, a west-to-east trough–ridge pattern over the U.S. at the heat wave midpoint. Composites are generated using the NCEP–NCAR reanalysis dataset (see section 2.2.5). These composites are presented in chronological order (D−7–D−1) as Figs. 3.7–3.10, with each figure consisting of 3 panels illustrating: (a) 200-hPa geopotential height and raw anomaly (1968–1996 climatology), (b) 700-hPa geopotential and raw anomaly (1968–1996 climatology), and (c) sea level pressure, 1000–500-hPa thickness, and 1000–500-hPa thickness raw anomaly (1968–1996 climatology).

Composites for D−7 show a west-to-east trough–ridge pattern at the 200-hPa level across North America (Fig. 3.7a). The trough over the west coast North America has a 3-6 dam negative geopotential height anomaly while the
ridge over the Rocky Mountains has a 3-6 dam positive geopotential height anomaly (Fig. 3.7a). This trough–ridge pattern is reflected in the 700-hPa geopotential height field over North America, although significant geopotential height anomalies are not present aside from a narrow 3-6 dam positive geopotential height anomaly over Northern Canada (Fig. 3.7b). The trough–ridge pattern over North America is also reflected in the 1000–500-hPa thickness field with the trough having a 3-6 dam negative thickness anomaly over the west coast of North America and the ridge having a 3-6 dam positive thickness anomaly over the Canadian Rocky Mountains and central Canada (Fig. 3.7c).

Across the Pacific Ocean, the 200-hPa geopotential height field is zonal with the exception of a broad, positively-tilted trough stretching from the Bering Strait southwest to the Sea of Okhotsk (Fig. 3.7a). This trough is also reflected in the 700-hPa geopotential height field with a maximum negative geopotential height anomaly of 2-4 dam located south and east of Kamchatka (Fig. 3.7b). A ridge is present in the 200-hPa geopotential height field over eastern Asia that is associated with a maximum positive geopotential height anomaly of 6-9 dam over eastern Mongolia and northeastern China (Fig. 3.7a). This ridge is reflected in the 1000–500-hPa thickness field with a maximum positive thickness anomaly of 4-6 dam also over eastern Mongolia and northeastern China. (Figs. 3.7a,b). The ridge over eastern Asia is not reflected in the 700-hPa geopotential height field (Fig. 3.7c). Instead, a weak geopotential height gradient is seen in the 700-hPa geopotential height field over eastern Asia that is likely reflecting the
elevated terrain of the eastern Tibetan Plateau and Gobi Desert where the surface is often above 700 hPa.

Composites for D₅ show that the 200-hPa trough–ridge couplet over North America has moved incrementally eastward from D₇ (Figs. 3.7a and 3.8a). The trough has deepened and now has a maximum negative geopotential height anomaly of 6-9 dam over the eastern Gulf of Alaska and British Columbia. The positive geopotential height anomaly of the ridge over North America has decreased and is likely a result of the motion and amplification of the trough over western North America and lowered geopotential heights over Hudson Bay. This change in the upper-level flow pattern over North America is also reflected in the 700-hPa geopotential height field as the trough over western North America has deepened and now has a maximum negative geopotential height anomaly of 2-4 dam over eastern Gulf of Alaska and British Columbia (Fig. 3.8b). The eastward motion of the 1000–500-hPa thickness trough over western North America reflects the 200- and 700-hPa geopotential height fields above, but the thickness anomaly of 3-6 dam is equal to the anomaly at D₇ indicating the strength of the thickness trough remains the same (Fig. 3.8c). The 1000–500-hPa thickness ridge over North America also moves eastward while the associated 3-6 dam positive thickness anomaly over North America expands in size but remains at the same strength as D₇ (Fig. 3.8c).

Upstream of the 200-hPa North American trough–ridge couplet from D₇ to D₅, two 200-hPa positive geopotential height anomalies have appeared that are associated with geopotential height rises (Fig. 3.8a). One anomaly is located
over the northern Aleutian Islands and the western Gulf of Alaska and has a positive geopotential height anomaly of 3-6 dam whereas the other anomaly is located to the southeast over the North Pacific Ocean and has a positive geopotential height anomaly of 3-6 dam. These 200-hPa geopotential height rises are associated with slight geopotential height rises at the 700-hPa level as well but are not associated with any significant geopotential height anomalies (Figs. 3.8a,b). Beneath the positive 200-hPa geopotential height anomaly over the Aleutian Islands and the Gulf of Alaska is a 2-4 dam positive thickness anomaly in the 1000–500-hPa thickness field (Fig. 3.8c).

The 200-hPa trough located just to the east of Kamchatka has deepened and is associated with a negative geopotential height anomaly of 3-6 dam over the Bering Sea (Fig. 3.8a). The 700-hPa trough beneath the 200-hPa trough has also deepened slightly; however, the accompanying negative geopotential height anomaly over the Bering Sea remains at 3-6 dam (Fig. 3.7b). Further upstream, the 200-hPa geopotential height and 1000–500-hPa thickness fields both reflect the ridge amplification taking place over eastern Asia (Figs. 3.8a,c). The 200-hPa geopotential height field has a maximum positive anomaly of 9-12 dam while the 1000–500-hPa thickness field has a maximum positive anomaly of 4-6 dam, with both anomalies over eastern Mongolia and northeastern China.

Composites for D₃ show that the 200-hPa trough–ridge pattern over North America has evolved into a trough–ridge–trough pattern (Fig. 3.9a). This is due to the appearance of lower 200-hPa geopotential heights over the eastern U.S. that are associated with a small negative geopotential height anomaly of 3-6 dam.
over the Middle Atlantic States. This trough is reflected in the 1000–500-hPa thickness field but not in the 700-hPa geopotential height field (Figs. 3.9b,c). The 200-hPa ridge has amplified and moved slightly east and now has a maximum positive geopotential height anomaly of 6-9 dam (Figs. 3.8a and 3.9a). The 200-hPa trough over western North America is stationary but deepens slightly as evidenced by the expansion of the negative geopotential height anomaly over western Canada (Fig. 3.9a). This trough amplification is strongly reflected in the 700-hPa geopotential height field as the associated maximum negative geopotential height anomaly has increased to 4-6 dam and expanded across most of northern and eastern Canada as well as the Gulf of Alaska (Fig. 3.9b). The deeper trough over western North America is also visible in the 1000–500-hPa thickness field and is associated with an increased negative thickness anomaly of 4-6 dam over the eastern Gulf of Alaska and British Columbia (Fig. 3.9c).

Upstream of North America, the two 200-hPa positive geopotential height anomalies identified at D.5 have merged over the east-central North Pacific Ocean, a consequence of a broad area of geopotential height rises (Figs. 3.8a and 3.9a). These positive geopotential height anomalies stretch from the Aleutian Islands to an area north and east of the Hawaiian Islands with two distinct maxima present with values of 6-9 dam. The rise in positive geopotential heights is also seen at the 700-hPa level and is associated with a positive geopotential height anomaly of 2-4 dam beneath the 200-hPa geopotential height anomalies (Fig. 3.9b). The northern positive geopotential height anomaly seen
on D.₇ and D.₅ continues to be reflected as a 2-4 dam 1000–500-hPa thickness anomaly while the southern geopotential height anomaly appears at D.₃ also as a 2-4 dam 1000–500-hPa thickness anomaly (Figs. 3.7c, 3.8c, and 3.9c).

The trough seen at 200 hPa east of Kamchatka has entered the Bering Sea by D.₃ while maintaining its relative strength (Figs. 3.8a and 3.9a). The trough can be seen at 700 hPa but has lost its relative strength as the 2-4 dam negative geopotential anomaly previously seen over the Bering Sea on D.₇ and D.₅ is now non-existent (Figs. 3.7b, 3.8b, and 3.9b). The 200-hPa trough east of Kamchatka is also reflected in the 1000–500-hPa thickness field with no appreciable thickness anomaly (Fig. 3.9c). The ridge at 200 hPa over eastern Asia is pushing eastward and amplifying as evidenced by the motion and expansion of the positive geopotential height anomaly over eastern Mongolia and northeastern China (Figs. 3.8a and 3.9a). The behavior of the 200-hPa East Asian ridge can likely be explained by the lower-tropospheric WAA occurring ahead of the surface low over Mongolia (Fig. 3.9c). WAA is implied by the orientation of the 1000–500-hPa thickness field with the SLP field, an orientation that has been in place from D.₇–D.₃ (Figs. 3.7c, 3.8c, and 3.9c).

From D.₇–D.₃, geopotential height rises over the east-central North Pacific Ocean have taken place at the 200- and 700-hPa levels (Figs. 3.7–3.9). From D.₅–D.₃, this ridge amplification appears to accompany the amplification and anchoring of the downstream trough over western North America. The amplification and anchoring of the trough over western North America is, in turn, acting to amplify the downstream ridge over central North America.
Composites for D$_1$ show that the 200-hPa trough–ridge–trough pattern over North America has changed since D$_3$ (Fig. 3.10a). The eastern trough has almost completely exited North America. The 200-hPa ridge over east-central North America has broadened and slightly amplified and is now associated with a maximum height anomaly of 12-15 dam over Ontario and the upper Midwest. This change is also reflected in the 700-geopotential height field with a maximum positive geopotential height anomaly of 4-6 dam over Ontario and the Great Lakes region and in the 1000–500-hPa thickness field with a maximum positive thickness anomaly of 8-10 dam over Ontario and the upper Midwest (Figs. 3.10b,c). The 200-hPa trough over western North America has progressed eastward and is now associated with a maximum negative geopotential height anomaly of 9-12 dam over British Columbia (Fig. 3.10a). This 200-hPa trough is also reflected in the 700-geopotential height field with a maximum negative geopotential height anomaly of 6-8 dam located over British Columbia, Alberta, and the Northwest Territories and in the 1000–500-hPa thickness field with a maximum negative thickness anomaly of 6-8 dam located over British Columbia and Washington (Figs. 3.10b,c).

From D$_3$–D$_1$, the increases in the positive and negative geopotential height and thickness anomalies associated with the ridge and western trough over North America take place in the presence of minor upper-level flow amplification (Figs. 3.9 and 3.10). These anomalies are most likely due to the eastward displacement of these upper-level features away from their climatological positions. Climatologically, the meteorological summer circulation
over North America is characterized by a West Coast trough, a product of the cool, northerly Californian oceanic circulation, a ridge over the western U.S., a product of surface heating over the elevated semi-arid terrain of the Rocky Mountains, and a trough along the East Coast, a product of the land-sea temperature contrast.

On D₇, the 200-hPa trough over the West Coast of North America exhibits a 3-6 dam negative geopotential height anomaly over the eastern Gulf of Alaska and eastern North Pacific Ocean while the ridge over the Rocky Mountains exhibits a 3-6 dam positive geopotential height anomaly over most of Canada (Fig. 3.7a). From D₇–D₁, as the 200-hPa flow pattern over North America progresses eastward, the flow pattern weakly amplifies yet the geopotential anomalies grow significantly larger (Figs. 3.7–3.10). A trough producing a 3-6 dam negative geopotential height anomaly over its climatological position, when displaced, will produce a much larger anomaly in a region climatologically predisposed to a ridge. A ridge producing a 3-6 dam positive geopotential height anomaly over its climatological position, when displaced, will produce a much larger anomaly in a region climatologically predisposed to a trough.

The 200-hPa ridge and accompanying positive geopotential height anomaly over the east-central North Pacific Ocean has moved east and is associated with a maximum geopotential height anomaly of 6-9 dam stretching from the Gulf of Alaska south and east (Fig. 3.10a). The two maxima in the geopotential height anomaly at D₃ have merged in one broad maximum anomaly of 6-9 dam centered over the east-central North Pacific Ocean at D₁. The 200-
hPa ridge over the east-central North Pacific Ocean continues to be reflected in the 700-hPa geopotential height field, accompanied by a positive 2-4 dam geopotential height anomaly, and in the 1000–500-hPa thickness field, accompanied by a positive 2-4 dam positive thickness anomaly, with both anomalies located over the east-central North Pacific Ocean (Figs. 3.10b,c). The positively-tilted trough at 200 hPa located over the Bering Sea and western North Pacific Ocean has moved east and weakened (Fig. 3.10a). This trough is reflected in the 700-hPa geopotential height field and in the 1000–500-hPa thickness field, but without any appreciable geopotential height or thickness anomalies. Over eastern Asia, the 200-hPa ridge has moved eastward but does not appear to have amplified (Fig. 3.10a). The 200-hPa geopotential height anomaly associated with the East Asian ridge has broadened but is now quantitatively weaker, now having a maximum positive geopotential height anomaly of 6-9 dam located over eastern Mongolia, northeastern China, and southeastern Siberia. This ridge is not well reflected at the 700-hPa level, but is nicely seen in the 1000–500-hPa thickness field with a maximum positive thickness anomaly of 4-6 dam located over eastern Mongolia (Fig. 3.10c).

Based upon the composites, the initiation of an NHW appears to be associated with a perturbation in the 200-hPa upper-level flow pattern. As a result, a Rossby wave pattern seems to emanate from either the eastern tropical Pacific Ocean or from eastern Asia. Another noteworthy characteristic is the apparent transfer of “energy”, or the downstream progression of peak amplitude, at the 200- and 700-hPa levels (Figs. 3.7–3.10). This is nicely seen in the time
series composites of 200- and 700-hPa geopotential height and in 1000–500-hPa thickness from D₃–D₁. It is at D₃ that the amplitude of the ridge over the eastern North Pacific Ocean is at its highest (Figs. 3.9a-c). By D₁, the amplitude of the ridge over the eastern North Pacific Ocean has decreased just as the western trough over North America has deepened and the ridge over North America amplifies (Figs. 3.10a-c). To a first-order approximation, this would appear to be an example of a Rossby wave train and downstream development.

Lastly, the composites representing D₃ and D₁ appear to exhibit sharper synoptic features than those composites representing D₇ and D₅. As the composite time series progresses from D₇–D₁, an apparent reduction in the smearing of features takes place (Figs. 3.7–3.10). Although the atmospheric features in the analyzed domain may be sharper solely due to meteorological processes, another explanation could be that the composites representing the NHW are clearer than those composites representing upstream antecedent conditions. This is likely an artifact of the selection criteria used to composite NHWs: that of a west-to-east trough–ridge pattern over the U.S. at the NHW midpoint (Dₘ). Composites generated closer to Dₘ should theoretically be of better quality than composites generated further out from Dₘ. This is a downside to using compositing as a method of analysis.

3.2.2 Synoptic climatology
The purpose of this section is to describe the time series composite analyses of NHWs. These composites nicely illustrate the evolution of the synoptic-scale flow pattern over North America that leads to an NHW. It is hoped that these consistent and identifiable synoptic-scale patterns can be used to help improve NHW forecasts in the future.

As detailed in section 2.2.5, the NHW composites are generated on D₄, D₃, D₂, D₁, D₀, D₉, and D₇. Of the 24 NHWs identified from the NCDC temperature dataset, 17 are chosen for compositing using a simple subjective criterion; namely, a west-to-east trough–ridge pattern over the U.S. at the heat wave mid-point. Composites are generated using the NCEP–NCAR reanalysis dataset (see section 2.2.5). These composites are presented in chronological order (D₄–D₇) as Figs. 3.11–3.17, with each figure consisting of 6 panels illustrating: (a) 200-hPa geopotential height, wind, and divergence, (b) 500-hPa geopotential height, wind, and absolute vorticity, (c) 700-hPa geopotential height, vertical velocity, and relative humidity, (d) 700-hPa geopotential height and wind overlain with 500–700-hPa and 700–925-hPa lapse rates, (e) 850-hPa geopotential height, wind, and temperature, and (f) sea level pressure overlain by 1000–500-hPa thickness and 200-hPa wind. Figure 3.18 illustrates the synoptic evolution of the 18°C isotherm at 850 hPa whereas Fig. 3.19 illustrates the synoptic evolution of the 23°C and 24°C isotherms at 925 hPa.

Composites for D₄ show that the principal large-scale feature over North America is the trough–ridge–trough pattern nicely seen in the 200-, 500-, and 700-hPa geopotential height fields (Figs. 3.11a,b,d) and in the 1000–500-hPa
thickness field (Fig. 3.11f). The ridge (R in Figs. 3.11–3.17) can be seen over interior North America with the ridge axis extending from the Central Plains region north-northwest through Saskatchewan. The eastern trough is located along the east coast of North America with the trough axis extending from the Carolinas northeastward through Quebec towards Hudson Strait. The western trough (T in Figs. 3.11–3.17) is located west of North America with the trough axis extending from northwestern British Columbia southward over the eastern North Pacific Ocean. As discussed in section 3.2.1, this western trough is likely acting to anchor the ridge over interior North America.

Upstream of the ridge in the geopotential height field is a thermal ridge as evidenced by high 1000–500-hPa thickness values over the Rocky Mountains (Fig. 3.11f) and by the 850-hPa temperature maximum over the southwestern U.S. and Mexican Plateau (Fig. 3.11e), a reflection of the topographically induced, elevated heat source typical of meteorological summer. A maximum in the 700–500-hPa lapse rate field is located over the western U.S. and largely overlaps the aforementioned 850-hPa temperature maximum (Fig. 3.11d). The location of these steeper 700–500-hPa lapse rates implies that surface-based mixed layers are likely present over the semi-arid elevated terrain of the southwest U.S.

By D3 the western trough has deepened and moved eastward with the trough axis now positioned near 130°W and parallel to the west coast of North America (Figs. 3.12a,b). The amplification of the western trough is likely attributed to with weak low-level cold air advection (CAA) as implied by the 850-
hPa temperature and wind fields (Figs. 3.11e and 3.12e). The eastward motion of the trough is likely attributed to the inferred anticyclonic vorticity advection (AVA) upstream of the trough axis and to the inferred cyclonic vorticity advection (CVA) downstream of the trough axis; implied by 500-hPa winds overlain with absolute vorticity contours (Figs. 3.11b and 3.12b).

Also on D₃, the ridge over North America has amplified as evidenced by an increase in the 200-hPa geopotential height field along the U.S.-Canada border (Figs. 3.11a and 3.12a). This ridge amplification is weakly reflected at the 500- and 700-hPa levels (Figs. 3.12b,d). The geopotential height rises can likely be attributed to low-level warm air advection (WAA) into the ridge as implied by the 850-hPa temperature and wind fields (Figs. 3.11e and 3.12e) and by the orientation of the 1000–500-hPa thickness contours overlain by SLP over the northern U.S. (Figs. 3.11f and 3.12f). The eastward progression of the ridge is likely attributed to AVA as implied by 500-hPa winds and absolute vorticity contours upstream of the ridge axis (Figs. 3.11b and 3.12b). The eastward progression and amplification of the western trough and ridge has influenced the midlevel and low-level thermal structure of the troposphere as seen in the poleward transport of 850-hPa isotherms (Figs. 3.11e and 3.12e), the amplification of the 1000–500-hPa thickness ridge (Figs. 3.11f and 3.12f), and in the poleward migration of steeper 700–500-hPa lapse rates over the western U.S. (Figs. 3.11d and 3.12d). Ridge amplification coupled with lower geopotential heights over British Columbia result in a sharper geopotential height
gradient with an attendant anticyclonically-curved jet streak stretching from the Pacific Northwest across the U.S.-Canada border (Fig. 3.12a).

Additionally on D₃, pressure falls on the lee of the Rocky Mountains are evident as a wave in the 850-hPa geopotential height field (Fig. 3.12e) and as an area of reduced SLP oriented along a northwest to southeast axis running parallel to the Rocky Mountains (Fig. 3.12f). This surface wave is located beneath the axis of an amplifying 1000–500-hPa thickness ridge: a signature of a warm-core structure east of the Rocky Mountains and, consequently, of lee cyclogenesis (Fig. 3.12f).

On D₂, as seen at 200, 500, and 700 hPa, the western trough has moved slightly eastward (Figs. 3.13a,b,d). The modest eastward progression of the western trough is likely attributed to the implied AVA upstream of the trough axis and to the implied CVA downstream of the trough axis as seen at 500 hPa (Figs. 3.12b and 3.13b) coupled with weak low-level CAA as implied by the 850-hPa temperature and wind fields (Figs. 3.12e and 3.13e) over the Pacific Northwest. Downstream of the western trough, the lee trough seen at D₃ in the 850-hPa geopotential height field west of the Canadian Rocky Mountains has strengthened and is present in the 700-hPa geopotential height field on D₂ (Figs. 3.12d,e and 3.13d,e).

The eastward push of the western trough at D₂ has resulted in the further eastward progression and amplification of the ridge accompanied by a reorganization of the midlevel and low-level temperature fields (Figs. 3.12d-f and 3.13d-f). The ridge axis is now running north-south along 100°W (Fig. 3.13a)
with the majority of the ridge amplification taking place from Manitoba and western Ontario south to the U.S. upper Midwest (Figs. 3.13a,b). The ridge amplification is likely due to WAA and is occurring in two parts. The first part involves the midlevel advection of warm air off of the Rocky Mountains by the western trough and 700-hPa lee trough, as is nicely seen as a wave and a northeastward expansion of the 700–500-hPa lapse rate field just downstream of the 700-hPa lee trough (Fig. 3.13d). The second part involves the low-level advection of warm air from east of the Rocky Mountains. A surface low has developed on the lee of the Canadian Rockies downstream of the western trough (Fig. 3.13f). Dynamical support for the surface low is implied by 200-hPa divergence downstream of the 200-hPa western trough over Alberta (Fig. 3.13a), and by upward vertical motion (negative omega) at 700 hPa over the U.S.-Canada border between Alberta and Montana (Fig. 3.13c). The surface trough on the lee of the Rocky Mountains extends south and east of the surface low and has deepened underneath the 1000–500-hPa thickness ridge (Fig. 3.13f). This surface trough is likely thermodynamically supported by adiabatically-warmed downslope flow at midlevels (Fig. 3.13d), evidenced by downward vertical motion and low relative humidity at 700 hPa over the U.S. Great Plains (Fig. 3.13c). The 1000–500-hPa thickness and SLP fields and the 850-hPa temperature and wind fields imply that there is a broad area of low-level WAA across North America (between 100°W and 90°W and 40°N and 60°N) ahead of the surface low in Canada and ahead of the surface trough east of the Rocky Mountains in the U.S.
(Figs. 3.13e,f). This WAA is reflected in the northeastward progression of 850-hPa isotherms from $D_3$ to $D_2$ (Figs. 3.12e and 3.13e).

On $D_{-1}$, the reorganization of the upper-level flow pattern continues with the further eastward advance of the trough–ridge–trough pattern over North America, as seen at 200, 500, and 700 hPa (Figs. 3.14a,b,d). The western trough axis is now located over the West Coast of North America near 124°W, running north-south from British Columbia into California. The ridge axis is now located near 90°W, running north-south from Hudson Bay through Ontario and into the Minnesota/Wisconsin area. The 700-hPa lee trough west of the Rocky Mountains, now strengthened, has moved eastward and is reflected in the 200- and 500-hPa geopotential height fields (S in Figs. 3.14–3.17), taking on the appearance of a shortwave trough (shortwave) (Figs. 3.13d and 3.14a,b,d). This shortwave may be impinging upon the ridge over the U.S.-Canada border east of the Rockies and upstream of the ridge axis, as seen in 500-hPa absolute vorticity and geopotential height (Fig. 3.14b) and in 700-hPa geopotential height (Fig. 3.14d).

On $D_{-1}$, the evolution of the synoptic-scale flow pattern over North America continues to influence the midlevel and low-level temperature fields (Figs. 3.14d,e). As a result, we see a continued displacement of the warm air off the Rocky Mountains at midlevels towards the northeast U.S. This is nicely seen in the 700–500-hPa lapse rate field where the plume of steep lapse rates is being advected directly ahead of the shortwave (Fig. 3.14d). This plume of steep lapse rates is oriented along the 700-hPa mean flow with the lapse-rate axis extending
from the Rocky Mountains over the Midwest through southern Ontario and into the Northeast. The northeastward advection of higher temperatures at the 850-hPa level continues. Figures 3.12e, 3.13e, and 3.14e illustrate that the north-south oriented 850-hPa thermal ridge seen at D$_3$ has folded over and become oriented southwest to northeast along the mean flow by D$_1$. Low-level and midlevel WAA is again implied by the 850-hPa temperature and wind fields (Figs. 3.13e and 3.14e) and by the orientation of 1000–500-hPa thickness and SLP fields (Figs. 3.13f and 3.14f), respectively. The 850-hPa thermal maximum and steeper 700–500-hPa lapse rates, originally located over the Rocky Mountains, have clearly responded to the passage of the shortwave progressing through west-central Canada, as evidenced by the plume of warm air at both the mid-tropospheric (Fig. 3.14d) and 850-hPa (Fig. 3.14e) levels immediately ahead the shortwave. This WAA signature is also nicely seen in Fig. 3.14f with the 1000–500-hPa thickness ridge located just east of the surface low. An additional factor in the low-level WAA is the anomalous westward expansion of the Bermuda High into the southeast U.S. along with rises in the 700- and 850-hPa geopotential height fields over the same location (Figs. 3.14d-f). This westward expansion of the surface-level high pressure system with 700- and 850-hPa geopotential height rises, coupled with the impinging shortwave, is acting to tighten the geopotential height gradient ahead of the shortwave and surface troughs located on the lee of the Rocky Mountains (Fig. 3.14f). The tighter 850-hPa geopotential height and SLP gradients and the attendant increases in wind speed imply an increase in lower-tropospheric WAA across the U.S. Midwest (Figs. 3.14e,f).
Figures 3.15a-f illustrate the synoptic-scale conditions present at NHW initiation ($D_0$). On $D_0$, the plume of warm air at midlevels and low-levels has reached the northeast U.S. As seen in the 200- and 500-hPa geopotential height fields, the western trough has lifted slightly while moving further east with the trough axis now located just west of 120°W (Figs. 3.14a,b and 3.15a,b). The ridge continues to amplify as it moves further east as seen in 200- and 500-hPa geopotential height. The ridge axis has become more positively tilted and is now oriented south-southwest to north-northeast; stretching from the Great Lakes to northern Quebec (Figs. 3.15a,b). A shortwave continues to be seen in the 500-hPa geopotential height and absolute vorticity fields and in 700-hPa geopotential height field over northern Canada west of Hudson Bay (Figs. 3.15b,d). The enhancement of the geopotential height gradient and attendant strengthening of the 200-hPa jet continues due to the persistent ridge amplification south of the impinging shortwave (Fig. 3.15a). The eastern trough continues to make its exit from North America as the upper-level flow pattern evolves (Figs. 3.15a,b,d).

On $D_0$, the northeast U.S. is immediately downstream of the ridge axis and underneath the equatorward exit region of a broad anticyclonically-curved jet streak, an area where quasi-geostrophic (QG) forcing for descent is implied. Also, negative divergence (convergence) at 200 hPa over the northeast U.S. implies tropospheric descent by reason of mass continuity. The dynamically induced subsidence over the northeast U.S., implied by low relative humidity and a broad area of downward vertical motion (positive omega) at the 700-hPa level (Fig. 3.15c), is likely suppressing convection and allowing for higher levels of
insolation. In association with the upper-level ridge moving eastward across Canada, the Bermuda High continues to build inland across the eastern U.S. acting to further enhance the SLP gradient and implied WAA (Fig. 3.15f). At 700 and 850 hPa, geopotential height rises over the southeast U.S. are acting to enhance the geopotential height gradient and increase wind speed, potentially enhancing WAA at those levels (Figs. 3.15d,e). This WAA likely explains most of the observed movement of the 850-hPa thermal plume, the plume of steep 700–500-hPa lapse rates, and the 1000–500-hPa thickness ridge towards the northeast U.S. (Figs. 3.15d,e,f).

Figures 3.16a-f illustrate the synoptic-scale conditions present at the midpoint (Dm) of an NHW. As is seen in the 200- and 500-hPa geopotential height fields, the western trough has lifted and progressed eastward with the trough axis now located near 115°W (Figs. 3.15a,b and 3.16a,b). The ridge has amplified slightly and has moved further east, as also seen in the 200- and 500-hPa geopotential height fields. The ridge axis continues to have a slight positive tilt while stretching from the eastern Great Lakes to northern Quebec (Figs. 3.16a,b). A shortwave continues to be seen in the 500-hPa geopotential height and absolute vorticity fields and in the 700-hPa geopotential height field over eastern Hudson Bay and northern Quebec (Figs. 3.16b,d). The geopotential height gradient is much tighter at Dm due to continued ridge amplification with an impinging shortwave. As a consequence, a strong 200-hPa jet is present poleward of the northeast U.S. (Fig. 3.16a). The northeast continues to be underneath the equatorward jet exit region of this strong 200-hPa jet streak and
downstream of the ridge axis, a favorable region for QG forcing for descent (Fig. 3.16a). Convergence at the 200-hPa level over a portion of the northeast U.S. continues to imply tropospheric descent (mass continuity) manifest as low relative humidity and a area of downward vertical motion at the 700-hPa level (Figs. 3.16a,c).

Also on D_m the Bermuda high has intensified and penetrated further inland across the eastern seaboard of the U.S. as evidenced by a closed 700 hPa height anticyclonic circulation center situated over the Middle Atlantic States (Fig. 3.16f). At 700 hPa and 850 hPa, geopotential height rises continue to enhance the geopotential height gradient and wind field, acting to sustain the plume of warm air over the Northeast (Figs. 3.16d,e). The northeast U.S. is underneath the warmest 850-hPa temperatures and highest 1000–500-hPa thickness values east of the Rocky Mountains (Figs. 3.16e,f). Although the plume of steep 700–500-hPa lapse rates has eroded significantly, the northeast U.S. remains underneath the remnants of the lapse-rate maximum (Fig. 3.16d).

Figures 3.17a-f illustrate the synoptic-scale conditions present on the last day (D_L) of an NHW. As seen in the 200-, 500-, and 700-hPa geopotential height fields, the western trough has weakened and is now located over central North America (Figs. 3.16a,b,d and 3.17a,b,d). The ridge has also weakened and moved eastward and is now over the northeast U.S. and eastern Canada (Figs. 3.16a,b and 3.17a,b). A shortwave has exited Canada and is acting to deepen the eastern trough over the Labrador Sea, as seen in the 500-hPa geopotential height and absolute vorticity fields and in the 700-hPa geopotential height field
(Figs. 3.17b,d). The 200-hPa jet streak has weakened slightly and progressed eastward (Figs. 3.17a,b). As a result, the northeast U.S. is upstream of the jet maximum and the ridge axis and, therefore, no longer located underneath a favorable region for QG forcing for descent. In association with the weakening upper-level ridge, the southeast U.S. surface high has retreated towards the Atlantic Ocean while the 700-hPa and 850-hPa anticyclones have weakened. The warm air has returned to its pre-heat wave, climatological position over the Rocky Mountains as seen in the 700–500-hPa lapse rate and 850-hPa temperature fields (Figs. 3.17e,f).

D_L represents the last day of an NHW and it is the upper-level flow pattern in place upstream of the northeast U.S. that will bring the NHW to an end. The future eastward progression of the weakened western trough is likely to occur due to the implied AVA upstream of the trough axis and to the implied CVA downstream of the trough axis as seen in the 500-hPa wind and absolute vorticity fields over the Midwest (Fig. 3.17b). The progression of this upper-level trough will act to move the ridge further downstream and away from the northeast U.S. CAA is occurring upstream of the northeast U.S. as implied by the 850-hPa temperature and wind fields and the 1000-500-hPa thickness and SLP fields (Figs. 3.17e,f). As a result of this CAA, the 850-hPa thermal trough located over central North America should progress eastward toward the northeast U.S., resulting in eventual cooling.

A characteristic of an NHW is the transport of warm air from the southwest U.S. towards the northeast U.S. This was implied in the time series composites
of 1000-500-hPa thickness overlain with SLP and 850-hPa isotherms overlain with geopotential height and winds (Figs. 3.11–3.17). For the period D.4–D.₇, composite time series of 850- and 925-hPa isotherms are used to further illustrate the evolution of the thermal structure of the lower troposphere across North America (Figs. 3.18 and 3.19). Figures 3.18a-e show the progression of 850-hPa isotherms from D.₄–D.₇ using the 18°C isotherm as a marker whereas Figs. 3.19a-e show the progression of 925 hPa isotherms from D.₄–D.₇ using the 23°C and 24°C isotherms as markers.

Initially, the warmest air at 850 hPa and 925 hPa is found over the western half of the U.S. along the elevated heat source of the Rocky Mountains (Figs. 3.18a and 3.19a). As the flow pattern evolves from D.₄–D.₂, the plume of warmest air in the west is perturbed and is displaced northward (Figs. 3.18a,b and 3.19a,b). From D.₂–D.₀, this warm plume is advected eastward toward the northeast U.S. (Figs. 3.18c,d and 3.19c,d) along the lower-tropospheric mean flow (Figs. 3.13e,f, 3.14e,f, and 3.15e,f). During this period, the warm plume is transported eastward on the equatorward side of the 200-hPa jet streak (Figs. 3.13a, 3.14a, and 3.15a). During the life of an NHW (D₀–D₇), the warm plume is located over the Northeast U.S. (Figs. 3.18d,e and 3.19d,e).

In situ development of warm air may be occurring between D₀ and D₇ as seen in the appearance of the 24°C isotherm over the Midwest at 925 hPa (Figs. 3.19c,d). The appearance of the 24°C isotherm is collocated with the steepest 700–925-hPa lapse rate east of the Rocky Mountains (Figs. 3.15d and 3.16d). However, it appears that the dominant signature is of warm air becoming
advected across North America from the semi-arid elevated source regions for the warm air over the western and southwestern portions of the U.S.

As discussed in section 3.2.1, the composites representing \( D_{4} - D_{1} \) appear to exhibit sharper synoptic features than those composites representing \( D_{0} - D_{L} \). As the composite time series progresses, an apparent reduction in the smearing of features takes place (Figs. 3.11–3.17). Again, this is likely an artifact of the selection criteria used to composite NHWs: that of a west-to-east trough–ridge pattern over the U.S. at the NHW mid-point (\( D_{M} \)). Composites generated closer to \( D_{M} \) should theoretically be of better quality than composites generated ahead of \( D_{M} \).
FIG. 3.1. Time series distribution of (a) anomalously hot days and (b) regional heat waves in the Northeast region (six stations) for the period 1948–2001. Included in (a) are linear regressions and regression equations for the periods 1948–1959 (blue; $R^2=0.0151$), 1960–2001 (orange; $R^2=0.0345$), and 1948–2001 (green; $R^2=0.0047$).
FIG. 3.2. Time series distribution of (a) anomalously hot days and (b) regional heat waves in the Southeast region (eight stations) for the period 1948–2001. Included in (a) are linear regressions and regression equations for the periods 1948–1959 (blue; $R^2=0.0291$), 1960–2001 (orange; $R^2=0.3148$), and 1948–2001 (green; $R^2=0.0948$).
FIG. 3.3. Time series distribution of (a) anomalously hot days and (b) regional heat waves in all regions east of the Rocky Mountains (40 stations) for the period 1948–2001. Included in (a) are linear regressions and regression equations for the periods 1948–1959 (blue; $R^2=0.0191$), 1960–2001 (orange; $R^2=0.0808$), and 1948–2001 (green; $R^2=0.0017$); in (b) are linear regressions and regression equations for the periods 1948–1959 (blue; $R^2=0.1144$), 1960–2001 (orange; $R^2=0.0609$), and 1948–2001 (green; $R^2=0.00004$).
FIG. 3.4. Time series distribution of (a) anomalously hot days and (b) regional heat waves in the Southwest region (five stations) for the period 1948–2001. Included in (a) are linear regressions and regression equations for the periods 1948–1959 (blue; $R^2=0.0418$), 1960–2001 (orange; $R^2=0.102$), and 1948–2001 (green; $R^2=0.0296$).
FIG. 3.5. Time series distribution of (a) anomalously hot days and (b) regional heat waves in all regions west of the Rockies (14 stations) for the period 1948–2001. Included in (a) are linear regressions and regression equations for the periods 1948–1959 (blue; $R^2=0.5715$), 1960–2001 (orange; $R^2=0.0145$), and 1948–2001 (green; $R^2=0.1199$); in (b) are linear regressions and regression equations for the periods 1948–1959 (blue; $R^2=0.0236$), 1960–2001 (orange; $R^2=0.005$), and 1948–2001 (green; $R^2=0.0601$).
FIG. 3.6. Time series distribution of (a) anomalously hot days and (b) regional heat waves across the U.S. (54 stations) for the period 1948–2001. Included in (a) are linear regressions and regression equations for the periods 1948–1959 (blue; $R^2=0.0024$), 1960–2001 (orange; $R^2=0.0898$), and 1948–2001 (green; $R^2=0.0233$); in (b) are linear regressions and regression equations for the periods 1948–1959 (blue; $R^2=0.1036$), 1960–2001 (orange; $R^2=0.0642$), and 1948–2001 (green; $R^2=0.0087$).
FIG. 3.7. Three-day average composite analyses centered on seven days prior to heat wave initiation. Analyses show (a) 200-hPa geopotential height (black; every 12 dam) and raw anomaly (1968–1996 climatology) (shaded; every 3 dam according to the color bar); (b) 700-hPa geopotential height (black; every 6 dam) and raw anomaly (1968–1996 climatology) (shaded; every 2 dam according to the color bar); and (c) sea level pressure (black; every 2 hPa), 1000–500-hPa thickness (red; every 3 dam), and 1000–500-hPa thickness raw anomaly (1968–1996 climatology) (shaded; every 2 dam according to the color bar). [N=17]
FIG. 3.8. As in Fig. 3.7 except centered on five days prior to heat wave initiation.
FIG. 3.9. As in Fig. 3.7 except centered on three days prior to heat wave initiation.
FIG. 3.10. As in Fig. 3.7 except centered on one day prior to heat wave initiation.
FIG. 3.11. (cont.)
FIG. 3.11. (cont.)
FIG. 3.11. (cont.)
FIG. 3.11. Composite analyses at four days prior to heat wave initiation. Analyses show (a) 200-hPa geopotential height (black; every 12 dam), wind (barbs; kt), wind speed (shaded; every 5 m s$^{-1}$ according to the color bar), and divergence [red; every 1x10$^{-6}$ s$^{-1}$, zero line omitted, where solid (dashed) lines denote divergence (convergence)]; (b) 500-hPa geopotential height (black; every 6 dam), wind (barbs; kt), and absolute vorticity (red; every 1x10$^{-5}$ s$^{-1}$); (c) 700-hPa geopotential height (black; every 6 dam), relative humidity (shaded; every 10% according to the color bar), and omega [blue; every 1 x 10$^{-3}$ µb s$^{-1}$, zero line omitted, where solid (dashed) lines denote descent (ascent)]; (d) 700-hPa geopotential height (black; every 3 dam), wind (barbs; kt), 700–500-hPa lapse rate (red; every 0.5°C km$^{-1}$, starting at 6°C km$^{-1}$), and 925–700-hPa lapse rate (blue; every 0.5°C km$^{-1}$, starting at 6°C km$^{-1}$); (e) 850-hPa geopotential height (black; every 3 dam), wind (barbs; kt), and temperature (red; every 3°C); and (f) sea level pressure (black; every 2 hPa), 1000–500-hPa thickness (red; every 3 dam), and 200-hPa wind speed (shaded; every 5 m s$^{-1}$ according to the color bar). The T and R in panel (b) represent the western trough and North American ridge, respectively. [N=17]
FIG. 3.12. (cont.)
FIG. 3.12. (cont.)
FIG. 3.12. As in Fig. 3.11 except at three days prior to heat wave initiation.
FIG. 3.13. (cont.)
FIG. 3.13. (cont.)
FIG. 3.13. As in Fig. 3.11 except at two days prior to heat wave initiation.
FIG. 3.14. (cont.)
FIG. 3.14. As in Fig. 3.11 except at one day prior to heat wave initiation. The S in panel (b) represents a shortwave trough.
FIG. 3.15. (cont.)
FIG. 3.15. (cont.)
FIG. 3.15. As in Fig. 3.14 except at heat wave initiation.
FIG. 3.16. (cont.)
FIG. 3.16. (cont.)
FIG. 3.16. As in Fig. 3.14 except at mid-point of heat wave.
FIG. 3.17. (cont.)
FIG. 3.17. (cont.)
FIG. 3.17. As in Fig. 3.14 except at last day of heat wave.
FIG. 3.18. Composite analyses of the 18°C isotherm at the 850-hPa level (red). Analyses shown at (a) four days before heat wave initiation, (b) two days before heat wave initiation, (c) initiation, (d) mid-point, and (e) last day of heat wave. [N=17]
FIG. 3.19. Composite analyses of the 23°C and 24°C isotherms at the 925-hPa level (red). Analyses shown at (a) four days before heat wave initiation, (b) two days before heat wave initiation, (c) initiation, (d) mid-point, and (e) last day of heat wave. [N=17]
4. Discussion

4.1 Statistical Analysis

For this thesis, an attempt was made to determine temporal trends in both anomalous hot days and regional heat waves by performing annual frequency analyses (see section 2.2.4) on National Climatic Data Center (NCDC) standard regions (Fig. 2.1) and on composites (sum aggregates) of all regions lying east of the Rocky Mountains (eastern regions), west of the Rocky Mountains (western regions), and across the contiguous United States (U.S.) Low tallies of regional heat waves in individual regions did not allow for high-confidence trendline analyses, but analyses could be performed on the regional and contiguous U.S. composites. To evaluate the statistical significance of trends in the annual frequency of anomalous hot days, coefficients of determination (R² values) were calculated with only those R² values exceeding 0.2 considered to be statistically significant indicators of trend. The trendline analyses focused on the periods 1948–1959, 1960–2001, and 1948–2001, based the results of DeGaetano and Allen (2002) and closely matching the methodology of Gaffen and Ross (1998, 1999).

Trendline analyses of the eastern regions showed that most of the trends were not statistically significant. For 1948–1959, negatively sloped trendlines are associated with the annual frequency of anomalous hot days and regional heat waves in the eastern regions composites (Figs. 3.3a,b). These negatively sloped trendlines, while not statistically significant, are generally consistent with the
negative trends in maximum temperature and $\geq 3$ day extreme maximum temperature exceedance events identified by DeGaetano and Allen (2002, their Figs. 8 and 12) in their central and eastern regions (defined as those stations located east of 110°W and 95°W, respectively) for the same period. Negatively sloped trendlines are also associated with the annual frequency of anomalous hot days in the Northeast and Southeast regions for 1948–1959 (Figs. 3.1a and 3.2a). These negatively sloped trendlines, despite not being statistically significant, are supported by the negative maximum temperature trend identified by DeGaetano and Allen (2002, their Fig. 8) in their eastern region for the same period.

Although not statistically significant, positively sloped trendlines for 1960–2001 correspond with the annual frequency of anomalous hot days and regional heat waves in the eastern regions composites (Figs. 3.3a,b). These positively sloped trendlines are generally consistent with the positive trends in maximum temperature and $\geq 3$ day extreme maximum temperature exceedance events identified by DeGaetano and Allen (2002, their Figs. 8 and 12) in their central and eastern regions for the same period. The annual frequency of anomalous hot days in the Northeast region is associated with a positively sloped trendline, although this trend is not statistically significant (Fig. 3.1a). A positively sloped trendline is also associated with the annual frequency of anomalous hot days in the Southeast region that represents a statistically significant trend ($R^2 = 0.3148$) (Fig. 3.2a). The positively sloped trendlines associated with the annual frequency of anomalously hot days in the Northeast and Southeast regions for
1960–2001 generally agree with DeGaetano and Allen (2002, their Fig. 8) and the positive maximum temperature trend they identified in their eastern region.

For 1948–2001 statistically insignificant, positively sloped trendlines are seen in the annual frequency of anomalous hot days and regional heat waves in the eastern regions composites (Figs. 3.3a,b). Positively sloped trendlines in the annual frequency of anomalous hot days are also seen in the Northeast and Southeast regions, but again these trends are not statistically significant (Figs. 3.1a and 3.2a). These results overlap with those of Gaffen and Ross (1998) and Robinson (2001). Gaffen and Ross (1998) found positive decadal trends for 1949–1995 in temperature and extreme heat-stress events (defined as three or four days of the consecutive occurrence of 85th percentile apparent temperatures) during meteorological summer across their Northeast and Southeast regions. Robinson (2001) found that for 1951–1990 the eastern contiguous U.S. showed an increased frequency of heat waves (defined as when, in a \( \geq 48 \) h period, neither the overnight low nor the daytime heat index fell below the NWS thresholds of 26.7°C and 40.6°C, respectively).

Subjectively, the eastern regions appear to exhibit a negative trend in both anomalous hot days and regional heat waves for 1948–1959 (Figs. 3.1a,b, 3.2a,b, and 3.3a,b). The trend appears to reverse itself for 1960–2001 where, subjectively, there appears to be a positive trend in both anomalous hot days and regional heat waves. This analysis is based on the annual frequencies of anomalous hot days and region heat waves where clearly the higher-frequency periods in the eastern regions are in the 1950’s, 1980’s and 1990’s. This is
consistent with the trendlines seen in the Northeast region, Southeast region, and in the eastern regions composite, although most are not statistically significant trends.

Trendline analyses of the western regions showed that most of the trends were statistically insignificant. For 1948–1959 positively sloped trendlines are seen in the annual frequency of anomalous hot days and regional heat waves in the western regions composites (Figs. 3.5a,b). The positive trend in the annual frequency of anomalous hot days is statistically significant ($R^2=0.5715$) whereas the positive trend in regional heat waves is not statistically significant. Both positively sloped trendlines are by and large consistent with DeGaetano and Allen (2002, their Figs. 8 and 12) and the positive trends for 1948–1959 in maximum temperature and $\geq 3$ day extreme maximum temperature exceedance events identified in their western region (defined as those stations located west of 110°W). A positively sloped trendline is also associated with the annual frequency of anomalous hot days in the Southwest region for 1948–1959 although this is not considered statistically significant (Fig. 3.4a). This positively sloped trendline is also by and large consistent with the positive trend in maximum temperature for the same period as recognized by DeGaetano and Allen (2002, their Fig. 8) in their western region.

For 1960–2001 the western regions composites exhibit positively sloped trendlines with the annual frequency of anomalous hot days and regional heat waves (Figs. 3.5a,b). Though not statistically significant, these positively sloped trendlines agree with the positive trends in maximum temperature and $\geq 3$ day
extreme maximum temperature exceedance events seen by DeGaetano and Allen (2002, their Figs. 8 and 12) in their western region for the same period. A positively sloped trendline is also linked to the annual frequency of anomalous hot days in the Southwest region for the period 1960–2001, although this trend is statistically insignificant (Fig. 3.4a). The positively sloped trendline identified for the Southwest generally agrees with results from DeGaetano and Allen (2002, their Fig. 8) of a positive trend in maximum temperature in their western region.

For the overall period 1948–2001, positively sloped trendlines are seen in the annual frequency of anomalous hot days and regional heat waves in the western regions composites (Figs. 3.5a,b). These trends are not statistically significant. A positive trend in the annual frequency of anomalous hot days is also seen in the Southwest region (Fig. 3.4a) and is also not statistically significant. These results generally agree with Gaffen and Ross (1998) and Robinson (2001). Gaffen and Ross (1998) found positive decadal trends for 1949–1995 in temperature and extreme heat-stress events during meteorological summer across western portions of the contiguous U.S. Robinson (2001) found positive decadal trend in heat wave frequency for 1951–1990.

The trendline analysis performed on the contiguous U.S. composites did not prove to be statistically significant. Subjectively, the annual frequency of anomalous hot days and regional heat waves for the contiguous U.S. composites (Figs. 3.6a,b) mimics the annual frequency of anomalous hot days and regional heat waves for the eastern regions (Figs 3.1a,b, 3.2a,b and 3.3a,b). Similarities in the annual frequencies of anomalous hot days and regional heat waves
between the eastern regions and the contiguous U.S. composites may be due to the 40:14 ratio of eastern region stations to western region stations contributing to the contiguous U.S. composites.

For 1948–1959 negatively sloped trendlines are associated with the annual frequency of anomalous hot days and regional heat waves in the contiguous U.S. composites (Figs. 3.6a,b). These negatively sloped trendlines, while not statistically significant, are generally consistent with the negative trends in maximum temperature and ≥ 3 day extreme maximum temperature exceedance events identified by DeGaetano and Allen (2002, their Figs. 8 and 12) in the contiguous U.S. for the same period. For 1960–2001 positively sloped trendlines are associated with the annual frequency of anomalous hot days and regional heat waves in the contiguous U.S. composites (Figs. 3.6a,b). These positively sloped trendlines are again not statistically significant. They do, however, generally agree with the positive trends in maximum temperature and ≥ 3 day extreme maximum temperature exceedance events identified by DeGaetano and Allen (2002, their Figs. 8 and 12) in the contiguous U.S. for the same period. For 1948–2001 the annual frequency of anomalous hot days and regional heat waves in the contiguous U.S. composites exhibit positively sloped trendlines that are not statistically significant (Figs. 3.6a,b). Results from the trendline analysis of the contiguous U.S. composites generally agree with Gaffen and Ross (1998, 1999) that show that for 1949–1995 positive decadal trends in temperature and extreme heat-stress events during meteorological summer exist across the contiguous U.S.
High-frequency years of anomalous hot days and regional heat waves in the eastern regions and the contiguous U.S. composites include 1952–1954, 1980, 1988, and 1995 (Figs. 3.3a,b and 3.6a,b); years that have been previously associated with significant heat wave activity in the contiguous U.S. Examples of literature on the heat wave–drought summers of 1952–1954 include *Monthly Weather Review* monthly summaries by Klein (1952a, 1952b), Winston (1953), Klein (1953), Hawkins (1954), and Winston (1954). Examples of literature on the heat wave–drought summer of 1980 include *Monthly Weather Review* monthly summaries by Dickson (1980), Livezey (1980), Taubensee (1980). A comparison of the 1980 and 1988 heat wave–drought summers was performed by Lyon and Dole (1995). Lastly, an example of an intense synoptic-scale heat wave in 1995 took place on 10–15 July 1995 across the Midwest (i.e., Livezey and Tinker 1996 and Galarneau et al. 2008). Higher-frequency years in the Western regions could not be correlated with previously published literature. The papers discussed in this paragraph will be further referenced in section 4.2 for discussion on the synoptic evolution of heat waves.

Overall, the lack of statistical significance for the vast majority of the trendlines analyses does not allow for making definitive statements about the long-term trends in either anomalous hot days or regional heat waves. However, subjectively it appears that the annual frequency of anomalous hot days and regional heat waves in the eastern regions was higher in the 1940’s and 1950’s and in the 1980’s and 1990’s. In the western regions, again subjectively, there appears to be an increase in the annual frequency of anomalous hot days and
regional heat waves throughout the analysis period. These increasing trends seen across the U.S. are generally consistent with the recent warming trends in the temperature record, which are frequently attributed to anthropogenic global warming. If these warming trends were to continue, an increase in the annual frequency of anomalous hot days and regional heat waves across the U.S. would likely be the result.
4.2 Northeast U.S. Heat Wave Composites

4.2.1 Upstream antecedent conditions

An attempt was made to describe the larger-scale upstream antecedent conditions that evolve into a northeast United States (U.S.) heat wave (NHW) using time-series composites that illustrate the evolution of the upper-level, midtropospheric, and lower-tropospheric flow patterns over the North Pacific Ocean Basin and North America. As detailed in section 2.2.5, the composites of upstream antecedent conditions are three-day averages centered on $D_{-7}$, $D_{-5}$, $D_{-3}$, and $D_{-1}$. These composites are presented in chronological order ($D_{-7}$–$D_{-1}$) as Figs. 3.7–3.10, with each figure consisting of 3 panels illustrating: (a) 200-hPa geopotential height and raw anomaly (1968–1996 climatology), (b) 700-hPa geopotential and raw anomaly (1968–1996 climatology), and (c) sea level pressure (SLP), 1000–500-hPa thickness, and 1000–500-hPa thickness raw anomaly (1968–1996 climatology).

A clear ridge-trough-ridge pattern stretching from over the east-central North Pacific Ocean to central North America is evident in the composites just prior to NHW initiation (Figs. 3.7–3.10). This ridge-trough-ridge pattern, seen at $D_{-1}$ in the 200- and 700-hPa geopotential height and 1000–500-hPa thickness fields and anomalies, appears to originate from the east-central North Pacific Ocean (Figs. 3.10a-c). The composite plots indicate that NHW initiation is associated with the amplification and eastward displacement of an anomalously strong ridge in the upper-level flow pattern towards the northeast U.S. (Figs. 3.7–
The ridge-trough-ridge pattern seen in the 200- and 700-hPa geopotential height and 1000–500-hPa thickness fields and anomalies in the composite plots implies that ridge building over the east-central North Pacific Ocean between D.7 and D.3 will cause downstream ridge amplification over North America by D.1 (Figs. 3.7, 3.8, and 3.10). The 200- and 700-hPa geopotential height rises over the east-central North Pacific Ocean between D.7 and D.3 are followed, subsequently, by 200- and 700-hPa geopotential height falls in the pre-existing, anomalously deep trough over the west coast of North America between D.5 and D.1 (Figs. 3.7–3.10). These 200- and 700-hPa geopotential height falls between D.5 and D.1 are themselves followed, subsequently, by 200- and 700-hPa geopotential height rises in the ridge over North America between D.3 and the time of NHW initiation (D.0, not shown). The composites seem to indicate two phases in the behavior of upper-level flow pattern. For D.7–D.3, the upper-level flow pattern over the eastern North Pacific Ocean and North America remains relatively stationary, whereas for the period D.3–D.1, that same upper-level flow pattern becomes more progressive as individual trough and ridges begin to move eastward.

The composite analyses appear to indicate that, prior to NHW initiation, the enhancement and later eastward progression of a pre-existing ridge over North America is coupled to the upstream development of a ridge over the east-central North Pacific Ocean (Figs. 3.7–3.10). This east-central North Pacific Ocean ridge building, found largely over the Gulf of Alaska and the Hawaiian Islands, begins approximately five days before NHW initiation, and is seen in the
pattern of 200- and 700-hPa geopotential height fields and anomalies. It appears to be related to two possible signals in the 200-hPa geopotential height field and anomaly composites on D.5 (Fig. 3.8a) and might explain why there are two maxima in the positive 200-hPa geopotential height anomaly over the east-central North Pacific Ocean at D.5 and D.3 (Figs. 3.8a and 3.9a). One signal appears to show that the 200-hPa ridge building event and associated maximum positive geopotential height anomaly at D.5 over the western Gulf of Alaska and Aleutian Islands is a part of a weak Rossby wave train (RWT) emanating from either eastern Asia or the western North Pacific Ocean (Fig 3.8a). The second signal is associated with ridge building and geopotential height rises away from the main flow at 200 hPa over the east-central equatorial Pacific Ocean. The mechanism increasing 200-hPa geopotential height values is not known at this time; however, a likely candidate for nondynamically driven ridge building aloft (i.e., 200 hPa) is the latent heat release (diabatic heating) associated with significant tropical convective activity, possibly associated with the phase and amplitude of the Madden–Julian Oscillation (e.g., Madden and Julian 1971).

Regardless of the cause, the strongest signal in the composites overall is the development of a ridge-trough-ridge pattern, as seen in the 200- and 700-hPa geopotential height fields, from D.5 though D.3 over the east-central North Pacific Ocean and North America. The ridge-trough-ridge pattern emanating from the east-central North Pacific Ocean along with the amplification of successive anomalies downstream of the initial ridge development over the east-central North Pacific Ocean is consistent with the signature of downstream baroclinic
development (e.g., Orlanski and Sheldon 1993) associated with the formation and eastward displacement of a RWT.

The findings of this analysis on the NHW upstream antecedent conditions are consistent with previous research on heat wave–droughts. Recall that McQueen and Shellum (1956), as discussed in section 1.2.3, determined that the only significant difference between long-term (seasonal) and short-term (synoptic-scale) heat waves was the persistence of the upper-level and midlevel flow patterns. Previous research has found an apparent association between positive geopotential height anomalies over the east-central North Pacific Ocean between the Gulf of Alaska and the Hawaiian Islands and heat wave–droughts over the U.S. Namias (1955, his Fig. 6; 1982 his Fig.13) found that if a strong 700-hPa positive geopotential height anomaly is in place over the east-central North Pacific Ocean (south of the Aleutian Islands) that it was highly likely both a negative geopotential height anomaly will be present over the west coast of North America and a positive geopotential height anomaly will be present over the northeast U.S.

The results of this thesis are consistent with the results Klein (1953), Winston (1953), Namias (1955, 1982), and Erickson (1983), Lyon and Dole (1995) who all found midlevel (i.e., 700 hPa) flow patterns associated with heat waves and heat wave–droughts over the contiguous U.S. Namely, that a strong ridge over the east-central North Pacific Ocean (with accompanying positive geopotential height anomalies) generates a ridge-trough-ridge pattern (an apparent RWT) that causes the amplification and eastward displacement of an
established ridge over North America that can trigger a NHW. A ~9 dam positive 700-hPa geopotential height anomaly was found by Klein (1952a, his Fig. 4) to be located over the east-central North Pacific Ocean in the monthly mean flow for June 1952, an anomalously hot month across the central U.S. Klein (1952b, his Fig. 1) also determined that a ~8 dam positive 700-hPa geopotential height anomaly was located over the east-central North Pacific in the monthly mean flow for July 1952, an anomalously hot month across the central U.S.

Examples of anomalously hot weather over the contiguous U.S. east of the Rocky Mountains that are correlated with a positive 700-hPa geopotential height anomaly (strong ridge) over the east-central North Pacific Ocean, a relationship discovered in this thesis, include: Winston (1953, his Fig. 4) who found a ~10 dam positive 700-hPa geopotential height anomaly in the monthly mean flow for June 1953; Klein (1953, his Fig. 10) who found a ~7 dam positive 700-hPa geopotential height anomaly in the monthly mean flow for August 1953; Hawkins (1954, his Fig. 1) who found a ~7 dam 700-hPa positive geopotential height anomaly in the monthly mean flow for July 1954; Winston (1954, his Fig. 2) who found a ~8.5 dam 700-hPa positive geopotential height anomaly in the monthly mean flow for August 1954; Dickson (1980, his Fig. 2) who found a 11.4 dam 700-hPa positive geopotential height anomaly in the monthly mean flow for June 1980; Livezey (1980, his Fig. 3) who found a 3.9 dam 700-hPa positive geopotential height anomaly in the monthly mean flow for July 1980; and Taubensee (1980, his Fig. 2) who found a 4.4 dam 700-hPa positive geopotential height anomaly in the monthly mean flow for September 1980.
The composites showing NHW upstream antecedent conditions indicate a possible association between RWTs generated from either the western Pacific Ocean or eastern equatorial Pacific Ocean and the eventual amplification and eastward progression of an existing ridge over the U.S. (e.g., a NHW) (Figs. 3.7–3.10). This finding is consistent with McQueen and Shellum (1956, their Fig. 3) who found that energy dispersion (i.e., a RWT) from near Siberia approximately five days earlier was partially responsible for the 9–13 June 1956 heat wave over the Great Lakes area. Livezey and Tinker (1996) concluded that the 10–15 July 1995 heat wave across the Midwest and northeastern portions of the U.S. had a possible connection to a RWT generated near the East Asian coastline from convective activity approximately seven days earlier. A time–longitude analysis by Galarneau et al. (2008, their Fig. A4) shows this RWT forming near East Asia, propagating across the North Pacific Ocean, and reaching the U.S. approximately seven days later at heat wave initiation. The RWT is also seen in the mean 500-hPa geopotential height analysis generated by Galarneau et al. (2008, their Fig. A1) for 6–10 July 1995.

4.2.2 Synoptic climatology

An analysis of the synoptic-scale conditions over North America prior to and during a northeast U.S. heat wave (NHW) was performed using time-series composites. These composites illustrate the structure and evolution of the upper-level, midtropospheric, and lower-tropospheric flow patterns over North
America that lead to a NHW. As detailed in section 2.2.5, the NHW composites are generated on $D_{-4}$, $D_{-3}$, $D_{-2}$, $D_{-1}$, $D_0$, $D_M$, and $D_L$. These composites are presented in chronological order ($D_{-4}$–$D_L$) as Figs. 3.11–3.17, with each figure consisting of 6 panels illustrating: (a) 200-hPa geopotential height, wind, and divergence, (b) 500-hPa geopotential height, wind, and absolute vorticity, (c) 700-hPa geopotential height, vertical velocity, and relative humidity, (d) 700-hPa geopotential height and wind overlain with 500–700-hPa and 700–925-hPa lapse rates, (e) 850-hPa geopotential height, wind, and temperature, and (f) sea level pressure (SLP) overlain by 1000–500-hPa thickness and 200-hPa wind. Figure 3.18 illustrates the synoptic evolution of the 18°C isotherm at 850 hPa whereas Fig. 3.19 illustrates the synoptic evolution of the 23°C and 24°C isotherms at 925 hPa. In general, the composite plots illustrate the synoptic-scale conditions and how they evolve to produce a NHW (Figs. 3.11–3.17).

At $D_{-4}$, a trough is located over the West Coast of North America and a ridge is located over central North America (Fig. 3.11a). From $D_{-4}$ to $D_0$, the upper-level flow pattern evolves such that the West Coast trough pushes into western North America causing the downstream ridge to amplify, initially ($D_{-4}$–$D_{-2}$) (Figs. 3.11a, 3.12a, and 3.13a), and then move eastward ($D_{-2}$–$D_0$) (Figs. 3.13a, 3.14a, and 3.15a). A similar pattern shift was identified by McQueen and Shellum (1956) who found that, for the 9–13 June 1956 synoptic-scale heat wave, an amplified, progressive trough impinging upon western North America also displaced a downstream ridge over central North America, subsequently advecting warm air into the Northern Plains and central Canada. A similar upper-
level flow pattern is seen for 10 June 1956 as analyzed by McQueen and Shellum (1956, their Fig. 2b) which shows a 300-hPa anomalously deep trough over the West Coast upstream of a 300-hPa amplified ridge over central North America. Another similar pattern shift was identified for the 6–10 July 1995 “amplification phase” of a ridge over North America prior to the 10–15 July 1995 Midwest heat wave where the amplification and eastward progression of a ridge over the Intermountain West was due to the presence of a strong, progressive trough upstream over the eastern North Pacific Ocean (Galarneau et al. 2008, Livezey and Tinker 1996).

At D₄ the maxima in 500–700-hPa lapse rate and 850-hPa temperature over the Intermountain West are likely a signature of the formation of surface-based mixed layers that are a result of sensible heating over semiarid elevated terrain (Figs. 3.11d,e). As these surface-based mixed layers are dragged eastward they transition into elevated mixed layers (EMLs). The ridge amplification between D₄ and D₂ over North America results in an intensification and northward expansion of the warm air beneath the ridge at 850 hPa that is also reflected in steep 500–700-hPa lapse rate (Figs. 3.11d,e, 3.12d,e, and 3.13d,e). This intensification and northward expansion of midlevel and low-level warm air (EMLs) is also consistent with the “amplification phase” and northward expansion of warm air at 850 hPa as described by Galarneau et al. (2008, their Figs. A1b, A5) just prior to the onset of the 10–15 July 1995 Midwest heat wave.

As the upper-level trough continues to push eastward across western North America between D₂ and D₀ (Figs. 3.13a, 3.14a, and 3.15a), a surface low
develops in the lee of the Canadian Rockies while a surface trough develops in the lee of the U.S. Rocky Mountains (Figs. 3.13f, 3.14f, and 3.15f). The southwesterly flow ahead of these surface features is likely advecting low-level warm air towards the north and east, as inferred by the 850-hPa wind and temperature fields and the 1000–500-hPa thickness and SLP fields over the Great Plains and central Canada (Figs. 3.13e,f, 3.14e,f, and 3.15e,f). This southwesterly flow is consistent with the southwesterly flow implied in the 850-hPa backwards trajectories calculated by Galarneau et al. (2008, their Fig. A2c) prior to the 10–15 July 1995 Midwest heat wave.

A wave in the 850-hPa temperature field and 500–700-hPa lapse rate over Canada east of the Canadian Rocky Mountains (Figs. 3.13d,e) is seen just downstream of either a lee trough or shortwave trough that has appeared in the D_2 composite of 700-hPa geopotential height (Fig. 3.13d). This event marks the beginning of the signature for D_2–D_0, that of the advection of warm air from the elevated arid and semiarid terrain in the western and southwestern portions of the U.S. towards the northeast U.S. This midlevel and low-level warm air advection (WAA) is occurring in response to the progressive long-wave trough over the west coast of North America in conjunction with the short-wave trough over Canada in the lee of the Rocky Mountains (Figs. 3.13a,e). The D_2–D_0 midlevel and low-level WAA is seen in the transport of steeper 500–700-hPa lapse rate values, higher 1000–500-hPa thickness values, and higher 850-hPa temperatures from the Intermountain West towards the northeast U.S. (Figs. 3.13d,e,f, 3.14d,e,f, and 3.15d,e,f). Additionally, a continuity analysis of the
The evolution of the 18°C isotherm at the 850-hPa level shows the advection of a plume of warm air from the Intermountain West toward the northeast U.S. (Fig 3.18). The overall advection (D$_{t}$–D$_{0}$) of this 850-hPa plume of warm air is consistent with the 21°C isotherm continuity map shown by Galarneau et al. (2008, their Fig. A5) which illustrates the transport of warm air at the 850-hPa level prior to the 10–15 July 1995 Midwest heat wave, including the initial northward transport during the ridge “amplification phase”, and the subsequent eastward transport thereafter. Galarneau et al. (2008) showed that continental tropical (cT) air masses produced over arid and semiarid source regions will warm in the presence of dynamically driven upper-level ridge amplification and accompanying subsidence in the region of anticyclonic shear along the equatorward flank of the subtropical jet (STJ). Subsequently, the downstream advection of these hot air masses is accomplished by transient disturbances propagating along strong STJs.

The midlevel and low-level WAA for (D$_{t}$–D$_{0}$) is also generally consistent with Klein (1952a) who showed that the geopotential height gradient created between an east-central U.S ridge and a trough over the west coast of North America, seen in the monthly mean flow for June 1952, likely acted to transport hot, desert-like air from the southwestern U.S. eastward. A similar pattern is also seen in the 10 June 1956 300-hPa analysis for the 9–13 June 1956 heat wave (McQueen and Shellum 1956, their Fig. 2b). The upper-level flow pattern shows an anomalously deep trough over the West Coast upstream of an amplified ridge over central North America, a pattern orientation that suggests the advection of
EMLs from the Intermountain West north and east along the equatorward flank of the STJ towards the Northern Plains and central Canada. Additionally, adiabatically-warmed downslope flow may be potentially acting to increase surface temperature downstream of the Rocky Mountains, as implied by the 500-hPa geopotential height field.

Surface-based mixed-layers become EMLs once they are advected away from their source regions. Again, the composites for D_{2–0} appear to show the advection of warm air at mid-levels from the Intermountain West towards the north and east, implying that the plume of steep 500–700-hPa lapse rate might reflect the advection of EMLs eastward. In addition, for D_{2–0}, the 200-hPa jet strengthens over the U.S.-Canada border as the geopotential height gradient has sharpened between the transient shortwave trough over northern Canada and the amplifying ridge to the south over the U.S. (Figs 3.13a, 3.14a, and 3.15a). Banacos et al. (2010) note that in order for an EML plume to persist into the northeast U.S. after advecting away from its source region, the plume must travel along a subsiding, anticyclonically curved flow and ultimately become entrained into a moderately fast westerly to northwesterly midtropospheric flow. The composites for D_{2–0} are consistent with Banacos et al. (2010) in which the aforementioned plumes of warm air at 850-hPa and steep 500–700-hPa lapse rate appear to be advected beneath the anticyclonic shear side of the existing 200-hPa anticyclonically curved jet (Figs. 3.13d,e, 3.14d,e, and 3.15d,e) and along the northern periphery of the strengthening 700-hPa anticyclone and attendant geopotential height gradient by D_{0} (Fig. 3.15d). The composites for
D₂–D₀ are also consistent with Galarneau et al. (2008, their Figs. A1b and A5) which show that during 11–15 July 1995 Midwest heat wave, the advection of warm air at 850 hPa, as shown in their 21°C isotherm continuity map, occurs along the anticyclonic shear side of a 500-hPa jet.

By NHW initiation (D₀), the adveceted plumes of warm air at 850 hPa and steep 500–700-hPa lapse rate have reached the northeast U.S. The relationship between EMLs and elevated surface temperatures in the central U.S. is established in the meteorological literature (i.e., Galarneau et al. 2009), but Banacos et al. (2010) and Cordeira et al. (2010) found that above normal surface temperature in the northeast U.S. can also be associated with steep lapse rate events (i.e., EMLs) as well. Cordeira et al. (2010) identified two steep lapse rate events, 4–7 April 2010 and 22–27 May 2010, over the northeast U.S. that were associated with record high surface temperatures. The composite plots show that, by NHW initiation, a plume of steep lapse rates is oriented along the 700-hPa mean flow and extend from the Intermountain West over the Midwest through southern Ontario and into the Northeast (Fig. 3.15d). This transport of steep lapse rate air is consistent with Banacos et al. (2010) and appears to be a critical component of NHW initiation.

Between D₀ and Dₗ, a strong 700-hPa closed anticyclone is present over the eastern U.S. that is encroaching northward into the northeast U.S. (Figs. 3.15d, 3.16d, and 3.17d). This is consistent with the results of Reed (1933, 1937) who found a strong correlation between the position of the North American anticyclone (at 700 hPa) and anomalously high surface temperatures. At the
surface and beneath the 700-hPa anticyclone is the Bermuda high that is also present over the eastern U.S. during the NHW (Figs. 3.15d, 3.16d, and 3.17d). This fits with Reed's (1937) conclusion that above normal surface temperatures exist when the Bermuda high was co-located beneath the midlevel North American anticyclone (at 700 hPa). Karl and Quayle (1981), in their work on the 1980 summer-season heat wave, also found anomalously warm surface temperature beneath areas of strongly positive 700-hPa geopotential height anomalies.

At $D_M$, the midpoint of a NHW, the Northeast U.S. is positioned beneath a favorable region for QG forcing for descent in the upper-level flow pattern as it is located below the anticyclonic shear side of the 200 hPa STJ exit region and downstream of a ridge axis. The 200-hPa STJ is strong due to the tight geopotential height gradient found between the shortwave trough over Northern Canada and the strong ridge to the south over the U.S. In their study of the 10–15 July 1995 Midwest heat wave, Galarneau et al. (2008, their Fig. A1b) presented the mean 500-hPa flow pattern so as to illustrate the synoptic conditions in place during the peak of the heat wave. Their 500-hPa mean flow is generally consistent with the 200- and 500-hPa flow patterns exhibited in the composite plots: that the Midwest was located just downstream of a ridge axis, on the equatorward flank of a strong STJ, and beneath the equatorward side of the STJ exit region. Galarneau et al. (2008, their Fig. A1b) show that this strong jet was the product an enhanced geopotential height gradient between an anomalously strong trough over Northern Canada and an anomalously strong
ridge south of the trough over the eastern U.S. The only major difference between the composite plots and the 500-hPa flow pattern seen in Galarneau et al. (2008, their Fig. A1b) is that the 500-hPa pattern for the 10–15 July heat wave is anchored further west, which is consistent with the fact that this heat wave was located upstream of the northeast U.S. over the Midwest.

Galarneau et al. (2008) and Galarneau and Bosart (2006) found that both the 10–15 July 1995 U.S. Midwest and 1–22 February 2004 Australian heat waves exhibited the following progression: the intensification of cT air masses in response to dynamically driven ridge building aloft over elevated arid and semiarid landscapes, then the transport of hot cT air then advected eastward away from its source region along the equatorward flank of a strong STJ subsequent to the initial ridge building. Additional warming due to subsidence on the equatorward flank of a strengthened STJ likely leads to further warming. This description closely matches the composite plots generated for this work showing the standard model for NHW development, thus helping to confirm this usefulness of this model.
5. Conclusions and Research Opportunities

5.1 Conclusions

The results of a statistical analysis in the annual frequency of anomalously hot days and regional heat waves across the contiguous U.S. is presented along with a synoptic climatology of northeast U.S. heat waves (NHWs) for the 54-year period 1948–2001. Heat waves were identified using a 54-station dataset of daily high temperatures retrieved from the National Climatic Data Center (NCDC) Daily Surface Dataset. The 54 stations selected were approximately equally distributed across the contiguous U.S. and were separated for analysis purposes into the nine standard climate regions as defined by the NCDC. Anomalously hot days were defined as those days where the daily high temperature exceeded the climatological 97.5 percentile temperature (approximating a plus two standard deviation anomaly). If a station experienced anomalously hot days for at least three consecutive days that station met the definition of a heat wave. Heat waves were defined as regional heat waves (i.e., NHWs) when they occurred simultaneously at two or more cities within an NCDC region.

Although high temperature data is widely assumed to be normally distributed, percentiles had to be used in place of standard deviation anomalies because it was found that high temperatures in most cities are actually not normally distributed. A statistical analysis shows that non-normal high temperature distributions are, in fact, common. For most cities (e.g., Denver), daily high temperatures more than two standard deviations below normal are
more frequent than daily high temperatures more than two standard deviations above normal, i.e., the high temperature distribution is skewed toward lower values. At a few stations, such as Los Angeles, the high temperature distribution is skewed toward higher values because daily high temperatures more than two standard deviations above normal are more frequent than daily high temperatures more than two standard deviations below normal. The significant outliers of daily high temperature skewing these datasets are consistently seen throughout the year regardless of season.

Denver was chosen to represent stations with a negatively skewed daily high temperature dataset whereas Los Angeles was chosen to represent stations with a positively skewed daily high temperature dataset. Although skewed distributions of daily high temperatures are seen in all four seasons, only the skewness associated with daily high temperatures during meteorological summer was chosen for analysis since summertime heat waves are the focus of this study. A composite of sea-level pressure (SLP) for the 25 most negatively skewed meteorological summer dates in Denver shows that the pressure gradient at the surface (anticyclone to the north and to the east of the Rockies) generates easterly surface upslope winds and adiabatic cooling. Upslope wind events at Denver produce weather that is significantly different from climatology, resulting in daily high temperatures that fall well below normal, thus skewing the dataset toward negative values. A composite of SLP for the 25 most positively skewed meteorological summer dates in Los Angeles shows that the pressure gradient at the surface generates adiabatically warmed gusty, downslope
easterly surface winds in response to higher SLP over the Great Basin than along the California coast (i.e., Santa Ana winds). Santa Ana winds at Los Angeles produce weather that is significantly different from climatology, resulting in daily high temperatures that are well above normal, thus skewing the dataset toward positive values.

In order to identify temporal trends in annual frequency of anomalously hot days and regional heat waves, time series of both metrics were prepared for the nine NCDC standard climate regions, those regions lying east of the Rocky Mountains (eastern regions), those regions lying west of the Rocky Mountains (western regions), and the contiguous U.S. as a whole (all regions). The trendline analyses focused on three time periods: 1948–1959, 1960–2001, and 1948–2001. The statistical analyses identified very few statistically significant trends in any of the time periods: the two exceptions are an increasing trend in the annual frequency of anomalously hot days seen in the Southeast region composite for 1960–2001 and an increasing trend in the annual frequency of anomalously hot days seen in the western regions composite for 1948–1959. A subjective analysis suggests that the annual frequency of anomalously hot days and regional heat waves in the eastern regions and the contiguous U.S. is higher in the 1940’s through 1950’s and in the 1980’s through 1990’s, while the western regions show a gradual increase throughout all the analysis periods.

Composite analyses are presented to examine the upstream antecedent conditions for NHWs. The composite analyses indicate that prior to NHW initiation an anomalous ridge seen in the upper-level flow pattern over west-
central North America amplifies and progresses eastward. The behavior of the ridge over North America is coupled to downstream development associated with the formation and eastward displacement of a Rossby wave train (RWT) (a ridge–trough–ridge configuration) in the upper-level flow pattern. The RWT appears to have formed in response to upper-level geopotential height rises over the east-central North Pacific Ocean that are found to occur approximately five days prior to NHW initiation.

The cause of the ridge amplification over the east-central North Pacific Ocean is potentially related to two distinct signals in the composite plots of upper-level geopotential height. One signal is identified over the western Gulf of Alaska and Aleutian Islands and appears to be part of a weak RWT emanating from either eastern Asia or the western North Pacific Ocean. The second signal is connected to ridge building and geopotential height rises away from the main upper-level flow over the east-central equatorial Pacific Ocean and is potentially linked to tropical convective activity associated with the Madden–Julian Oscillation.

Composite analyses are also presented to examine the synoptic-scale evolution of NHWs. The synoptic-evolution of NHWs can be roughly divided into three phases: the amplification phase, and advection phase, and the NHW phase. The amplification phase occurs approximately four to two days before NHW initiation, and involves the amplification of an existing upper-level trough–ridge pattern over western North America, the latter two components of the aforementioned ridge–trough–ridge pattern identified in the upstream antecedent
conditions. The development of the upper-level ridge over the east-central North Pacific Ocean is likely associated with the downstream amplification of both the West Coast trough and the ridge over the Intermountain West. This ridge amplification results in the intensification and northward transport of midlevel surface-based mixed-layers and low-level continental tropical (cT) air masses that formed over the arid and semiarid elevated terrain of the Intermountain West.

The advection phase begins approximately two days before NHW initiation with the trough–ridge pattern over western North America becoming progressive just as a short-wave trough appears over west-central Canada in the lee of the Rocky Mountains. The combination of these synoptic features acts to advect low-level warm air and midlevel surface-based mixed-layers, now termed elevated mixed-layer (EMLs) when advected away from their source regions, eastward away from the Intermountain West and towards the northeast U.S. The plumes of low-level warm air and EMLs are advected eastward along the equatorward flank of a strong upper-level subtropical jet (STJ) and on the northern periphery of a strong midlevel anticyclonically curved anticyclone over the southeast U.S. The midtropospheric warming associated with the subsidence found on the equatorward flank of an anticyclonically curved STJ coupled with relatively fast midlevel flow is likely important in the eastward advection and maintenance of regions of steep midlevel lapse rates (i.e., EMLs) toward the northeast U.S.
A NHW will occur when the low-level warm air and steep midlevel lapse rate air (i.e., an EML) advected from the Intermountain West reaches the northeast U.S. At the same time, the upper-level flow pattern is positioned so that northeast U.S. is located beneath the equatorward exit region of a strong STJ and downstream of a ridge axis, a favorable region for quasi-geostrophic (QG) forcing for descent. This midtropospheric descent and attendant warming is likely acting to maintain the steep midlevel lapse rates and accompanying low-level strong static stability, an arrangement consistent with above normal surface temperatures.

5.2 Research Opportunities

The analyses in this thesis used 54-year (1948–2001) daily high temperature datasets for 54-stations located across the contiguous U.S. Future work should extend the analysis to include daily high temperature data for 2001–2011. Additional daily high temperature data could be gathered from the National Oceanic and Atmospheric Administration’s (NOAA) Cooperative Observer Program (COOP) stations so as to eliminate spatial gaps between first-order stations in the National Climatic Data Center’s (NCDC) standard climate regions and to help identify regional heat waves trends by increasing the robustness of the frequency data. For the statistical analyses, the addition of these data would likely influence the statistical significance of the trendline analyses performed on the data and potentially bolster previously identified
trends in the data, as well as potentially identify new, more recent regional heat wave events. Future statistical analyses could focus on overnight low temperatures or NHW duration and attempt to identify any trends in either. Another possibility would be to use nonlinear trend analyses to see if better-fit trendlines with higher coefficients of determination ($R^2$ values) could be found.

Future investigations of the upstream antecedent conditions and the synoptic evolution of NHWs should include the use of case studies. While the composite plots are helpful in creating a conceptual model of how a NHW is initiated, case studies should be performed to help identify potentially important but subtle features that are likely smoothed out by the compositing procedure. Case studies would likely determine if the signatures of RWT formation that were seen in the composites of the upstream antecedent conditions in the western North Pacific Ocean and in the eastern equatorial Pacific Ocean exist in real meteorological situations. Analyses of outgoing long-wave radiation (OLR) data and the advection of potential vorticity in the upper troposphere by the irrotational wind could help to determine if RWT formation was associated with convective activity and if that convective activity was associated with the phase and amplitude of the Madden–Julian Oscillation. Hovmöller diagrams could be employed with these cases to follow the progression of RWTs and downstream development prior to NHW initiation.

For future composites, more recent NHW cases could be added to the composite analyses to help improve the signatures found in the present composites, assuming they fit the original selection criteria. Statistical
significance and correlation testing could be performed on the composites of the upstream antecedent conditions so as to determine if or any synoptic features are statistically significant to the later development of NHWs. Composite soundings could be generated at key locations such as along the path that the plumes of warm air and steep lapse rates take on their way towards the northeast U.S.

For the synoptic evolution, case studies could include an analysis of vertical cross-sections along the equatorward flank and the equatorward jet exit region of the STJ would help to determine the sense of vertical circulations flanking the jet and the role of the secondary circulations near the jet exit region in maintaining steep lapse rate air or elevated mixed-layers. Backwards trajectories could be generated for these case studies to determine source regions for warm air at various pressure levels. Additionally, these case studies would help determine if the short-wave/lee trough found over west-central Canada approximately two days before NHW initiation actually exists or if it is an artifact in the composites plots.

Lastly, the next step for future investigations into heat waves would be to take the established analysis techniques developed for NHWs and apply them to other NCDC standard climate regions to help assess the direct and indirect causes of variability in regional heat waves. This analysis would be done to create synoptic climatologies of heat waves for other areas within the U.S. and to determine the upstream antecedent conditions that lead to heat waves in these other regions. A comparison of the antecedent conditions necessary for heat
wave development and how these heat waves evolve across various regions of the U.S. would be a natural extension of this thesis.
REFERENCES


