A synoptic-dynamic analysis of the structure and evolution of persistent north Pacific wintertime ridge regimes

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A SYNOPTIC-DYNAMIC ANALYSIS OF THE STRUCTURE AND EVOLUTION OF
PERSISTENT NORTH PACIFIC WINTERTIME RIDGE REGIMES

by

Tyler C. Leicht

A Thesis
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ABSTRACT

Along the west coast of North America (NA), a majority of the precipitation for the entire year falls during winter as a result of approximately a dozen cyclones or atmospheric rivers. However, persistent upper-level ridges can prevent the occurrence of wintertime precipitation events and lead to extended drought, significant water shortages, and adverse economy-wide impacts for major population centers in California. An increased understanding of the dynamical and thermodynamical processes that govern upper-level ridge formation, persistence, and dissipation during the rainy season along western NA would allow decision makers to better manage water resources and motivates this thesis.

Persistent upper-level ridges will be defined by positive 500-hPa height anomalies ≥1.0 standard deviation that last for longer than 7 days, constructed using NCEP-NCAR reanalysis from 1948–2017. This resulting climatology will be used to construct composite analyses based off four pre-defined domains to uncover dynamical and thermodynamical linkages between persistent ridge regimes in different domains. Statistical relationships with teleconnections and case study examination are included in this presentation to further explain the evolution of persistent ridges.

Composite analysis of the persistent ridge lifecycle shows different dynamical processes are important for persistent ridges in each domain, but nearly all composites highlight important interactions with upstream cyclones throughout the ridge lifecycle. In the statistical relationships, it was found that the phase of the North Pacific jet stream had the most statistically-significant frequency anomalies up to 10 days prior to the start of a persistent ridge, while the Arctic Oscillation had a much weaker relationship for most of the persistent ridges. Finally, the case
study further demonstrates some of the dynamical and thermodynamical processes found through the composite analysis, with the biggest emphasis on repeated cyclogenesis aiding in the maintenance of a quasi-stationary ridge.
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1. Introduction

1.1 Motivation

A blocking pattern is defined as an upper-level anticyclone that remains quasi-stationary and diverts the jet stream around the upper-level anticyclone for around one week (Rex 1950, Huseke 1959, American Meteorological Society 2021). Blocking patterns have been studied for many decades because of their association with extended periods of anomalous weather conditions (Namais 1947). Fig.10 from Rex (1950) offers a good example of early studies on blocking patterns, using initial radiosonde data to analyze the middle and upper troposphere to diagnose large-scale flow patterns. In later decades, analytic studies such as Charney and DeVore (1979) helped to improve the theoretical understanding of blocking formation, maintenance, and dissipation. This study, and many others, helped to distinguish the importance of barotropic and baroclinic instabilities as they related to blocking patterns. Later studies (Mullen 1986, 1987) used more complex general circulation models to study blocking patterns, yet these models were still low resolution due to limitations of computational expense at the time. While these studies established our foundational understanding of blocking patterns, there are opportunities to revisit these early papers with longer periods of reanalysis data, improved computational resources, and a better dynamical understanding of blocking patterns.

As mentioned above, one of the main reasons for the extensive study of blocking patterns is because of the surface weather impacts of blocking patterns and how impactful those can be on a societal level. One region where knowledge of blocking patterns and their surface impacts becomes a critical forecast problem is western North America. This region has the highest
interannual variability in annual precipitation in the contiguous United States (CONUS), much of which occurs in the cool season from November through March (Dettinger 2011). During the short rainy season from November through March, much of that precipitation is associated with around a dozen or so atmospheric rivers (ARs) (AMS 2021). In this region, water managers and other agencies rely on longer-term forecasts on the subseasonal-to-seasonal (S2S) timescale in order to account for this high level of variability in precipitation timing and amounts (Dettinger 2013). Relevant forecasts made today typically rely on specific time periods known as “forecasts of opportunity” when specific combinations of various atmospheric and oceanic teleconnections are met (e.g. Pacific-North American pattern or PNA, Quasi-Biennial Oscillation or QBO, El Niño Southern Oscillation or ENSO, and the Madden-Julian Oscillation or MJO) (DeFlorio et al. 2018, Mundhenk et al. 2018). While valuable, these forecasts of opportunity are limited to specific periods where the requisite criteria are met, leaving large periods of the cool season without effective guidance on rainfall patterns across western North America. This thesis is motivated by the opportunity to examine blocking patterns and their associated weather regimes over the North Pacific Ocean and western North America from a synoptic-dynamic perspective to compliment the large body of research that already exists.

1.2 Literature Review

1.2.1 Weather Regimes during Winter

The phenomenon of atmospheric blocking is one of the most typical and well-studied weather regime that can impact surface weather in the midlatitudes. Multiple methods for diagnosing and identifying blocking patterns have been developed, but one of the simplest
methods was proposed by Dole and Gordon (1983). Instead of explicitly mentioning blocking, this study uses the terminology of persistent anomalies, examining both positive (ridge) and negative (trough) 500-hPa geopotential height anomalies that last for 5 or more days, without any requirements of flow diversion. This simple definition allowed for the detection of regions most prone to these persistent geopotential height anomalies. Fig. 3a from Dole and Gordon (1983) shows their climatological results using 14 years of data (1963–1977) for December, January, and February of those years. They found three hemispheric maxima of positive persistent anomalies: the North Pacific Ocean (south of the Aleutian Islands), the North Atlantic Ocean (south of Greenland), and the Ural Mountains in Siberia. The maximum of positive persistent anomalies across the North Pacific is also the largest in areal extent, suggesting that these anomalies are highly relevant to understanding downstream impacts across western North America.

Additional work (Dole 1986, 1989) helped to understand the dynamics of persistent anomalies, while linking the observed characteristics of persistent anomalies to theoretical work. Primarily, Dole (1986) found similarities in their observed North Pacific positive persistent anomalies to the results of Simmons et al. (1983). Simmonds et al. (1983) conducted idealized modeling of a barotropic model initialized with perturbations along and to the south of a climatological North Pacific jet stream. The similarities between observed persistent geopotential height anomalies and results of idealized model output forced by initial jet perturbations further motivates the need to consider the structure and variability of the polar and subtropical jet streams for understanding weather regime behavior. Subsequent work by Woollings et al. (2018) discusses barotropic perturbations along the jet stream (such as Rossby wave breaking, RWB) as a mechanism for explaining jet variability on an interannual timescale. This feedback between
the jet stream and barotropic instability is key to understanding the climatology of weather regimes on longer timescales.

On shorter timescales, Winters et al. (2019) help to explain the importance of the evolution and predictability of the North Pacific jet stream and its relation to large-scale weather regimes. Through empirical orthogonal function (EOF) analysis of the cool season North Pacific jet, they diagnose which configurations of the North Pacific jet stream (NPJ) are more or less predictable. In general, eastward extensions and poleward displacements of the NPJ at the time of forecast initialization are more predictable, while westward retractions and equatorward displacements of the NPJ at the time of forecast initialization are less predictable. Within each phase of the jet, comparison of the best versus worst forecasts show the worst forecasts are initialized with a positive 250-hPa geopotential anomalies east of the Dateline in the North Pacific basin. By day eight of these forecasts, the composite difference has evolved into a high-amplitude ridge or blocking pattern in all four phases of the NPJ. In general, these results suggest that current forecast models struggle with predicting blocking patterns and high-amplitude ridges in general.

One additional framework for understanding large-scale weather regime behavior is Rossby wave packets (RWPs). Fig. 13 from Rothlisberger et al. (2018) shows a schematic composite evolution of the formation of RWPs across the North Pacific basin. Rothlisberger et al. (2018) investigated the initiation of these RWPs, finding a distinct maximum of Northern Hemisphere RWPs from November through February. This study also found a regional maximum of RWP initiation from 150°E to 150°W and from 30°N to 45°N, highlighting the importance of the North Pacific basin as a source of RWP precursors. As these RWPs propagate eastward, they can interact with pre-existing flow perturbations such as blocking patterns to
reintensify and maintain the block; mature RWPs that approach an existing blocking pattern lead to large-scale deformation upstream of the blocking pattern that can help to reinforce the blocking pattern through warm air advection and anticyclonic vorticity advection. This reinforcement can act to prolong and maintain a blocking pattern, corroborating results by Shutts (1983) on eddy interactions with blocking pattern persistence.

Quinting and Vitart (2019) furthered this work by investigating the representation of RWPs in S2S models, given the potential predictability of these large-scale flow patterns. By comparing results from multiple operational S2S prediction models, they found a systematic underrepresentation of RWP decay and a tendency to have RWPs propagate farther downstream than in observations. However, this bias in underestimated RWP termination was also a function of model resolution, with coarser S2S models having a larger underestimation of RWP decay. Furthering the connection of RWPs and blocking patterns, Quinting and Vitart (2019) also found an overall underestimation of the frequency of blocking patterns in these same forecast models, which they attributed to a lack of RWP decay. They attribute some of the underrepresentation of RWP decay and blocking frequency to missing feedbacks between synoptic-scale eddies and pre-existing blocked flow patterns.

1.2.2 Blocking Theory and Dynamics

Rex (1950) was among the first to examine recent observations of the middle and upper atmosphere to try to understand atmospheric blocking. In the coming decades, many studies tried to address the fundamental question of scale (time and space) and type of instability (barotropic or baroclinic) that is most important to the formation and persistence of atmospheric blocking.
patterns. Charney and DeVore (1979) used a simple barotropic channel model to understand the atmospheric response to forcing from topography and low-level heating in an idealized model framework. They found that the above-mentioned barotropic forcing mechanisms produced realistic blocking patterns, with both stable and unstable solutions generated. These solutions are one of many possible stable solutions and can move between these solutions through forcing by smaller-scale baroclinic instability unaccounted for in this analysis. In contrast, Hoskins and Valdes (1990) explained fundamental theory on how persistent flow patterns such as active storm tracks can be continuously regenerated through baroclinic instability. They highlighted the importance of upstream vorticity fluxes, which act to distort the upstream potential temperature gradient, in maintaining the regional baroclinicity. However, vorticity fluxes alone are insufficient to account for the total observed baroclinity. They also found that mean diabatic heating within the cyclone itself helped to regenerate some of this baroclinicity for future cyclones.

Further studies using more complex models helped to further the debate about the dynamic mechanisms necessary for blocking formation and maintenance. Mullen (1986, 1987) focuses on the same mechanism that Hoskins and Valdes (1990) propose to maintain baroclinicity within a storm track: relative vorticity advection. Mullen (1986, 1987) used an early spectral general circulation model to evaluate vorticity and heat budgets around blocks identified in the model run. Through decomposing the vorticity and heat budget, he found that eddy vorticity fluxes in the mid-troposphere acted to help the blocking pattern propagate against the mean flow and maintain its quasi-stationary position. Fig. 16 from Mullen (1987) shows the schematic form of the interaction between various vorticity and heat fluxes necessary to maintain a blocking pattern. Just upstream of the block, the anticyclonic circulation depicted shows the
effects of upstream eddy vorticity fluxes (around ¼ of a wavelength upstream) in inducing and strengthening the anticyclonic anomaly of the block itself. Mullen (1987) also mentions that temperature advection by the mean flow, and not the eddies, acts to maintain the thermal anomaly associated with the block.

Colucci (1981, 1985) brings further insight into the role of planetary-scale waves in atmospheric blocking dynamics. Colucci (1985) compared two cases of explosive cyclogenesis that lead to very different planetary-scale wave patterns downstream. This work connects to Mullen (1986, 1987), finding temperature advection dominates low-level geopotential height changes, while vorticity advection dominates upper-level geopotential height changes. Thus, unique characteristics of the deepening cyclone will impact the growth and development of any downstream planetary-scale waves. Colucci (1981) conducted Fourier analysis to assess the planetary wavenumbers contributes most to blocking patterns through their lifecycle, and compared these results to idealized modeling of blocked flow in a quasi-geostrophic channel model. He found that a zonal wavenumber-3 and meridional wavenumber-2 grow in amplitude near the peak of the block, with energy growing from lower wavenumbers as the blocking pattern develops.

The importance of upstream conditions in blocking dynamics is also critical to understanding the full lifecycle of a blocking pattern. In one case analyzed in Colucci (2001), it was found that geostrophic stretching deformation preconditioned the upstream environment for block formation. Additionally, spectral decomposition of the contribution to the observed deformation found both synoptic- and planetary-scale potential vorticity (PV) advection contributed to this deformation in the days leading up to the blocking. Luo et al. (2019) found through their analytical blocking model that the upstream PV gradient was a key indicator of
blocking intensity and duration, and also allows for greater interaction with upstream eddies. The upstream PV gradient not only includes the zonal background flow, but also meridional shear of the zonal wind as well, giving more details than just the observed zonal wind. Nakamura and Huang (2018) use yet another derived quantity, local wave activity (LWA), to quantify the upstream drivers of block initiation. Fig. 1 from Nakamura and Zhu (2010) is a schematic representation of the calculation of LWA. This quantity represents the propagation of Rossby waves within the jet stream, and is able to be distinguished from meridional excursions of the jet because it is relative to the PV gradient associated within the jet stream. Results from this work show a column-averaged build up of LWA in the region just upstream of the developing block, indicative of synoptic systems slowing down from weakening upstream flow and creating a situation analogous to a “meteorological traffic jam”. This, along with the Colucci (2001) and Luo et al. (2019), help to explain how and why reduction in upstream zonal wind is so critical for blocking formation.

While there are a wide variety of proposed mechanisms and physical processes critical to blocking formation and maintenance, it can be challenging to reconcile which mechanisms are most critical and at what time in the lifecycle of a blocking pattern. One potential source of clarity comes from Nakamura et al. (1997), who created composites of the 30 strongest blocking cases in the North Atlantic basin and North Pacific basin. They assessed the relative contributions of high- and low-frequency forcing for the observed blocking formation for the composites of each basin. They found that there appeared to be clear regional differences in which processes were most important for blocking patterns. In the North Pacific, around 70% of the observed height anomalies associated with the block came from upstream synoptic storms. In contrast, this number was less than 40% over the North Atlantic, with decaying RWPs and other
low-frequency variability accounting for the rest of the observed height anomalies. This finding compliments the more recent work of Quinting and Vitart (2019) and Rothlisberger et al. (2018), who highlighted the importance of RWP termination for Atlantic/European sector blocking. It also suggests that more careful understanding of upstream synoptic conditions is critical for our dynamic understanding of blocking patterns across the North Pacific and western North America.

1.2.3 Arctic and Tropical Impacts on Midlatitude Flow

While blocking events are mostly concentrated in the midlatitudes, there are remote drivers and influences on blocking dynamics and evolution from both the tropics and the Arctic. Screen et al. (2018) ran several modeling experiments to examine the midlatitude response to sea-ice loss in the Arctic Ocean, finding a strengthening of the climatological Siberian High and weakening of the Aleutian Low. While mechanisms are proposed for this modeled surface pressure change (i.e. planetary-scale wave changes, shifting of the climatological storm track, and changes in meridionally-propagating Rossby waves), there is no unifying theory on the exact mechanisms involved. In addition, most of the variability between modeling experiments was found to be the geographic distribution of the sea-ice, meaning regional sea-ice loss could greatly influence the midlatitude flow response. Nakamura et al. (2016) wanted to further understand the linkage between sea-ice and midlatitude flow patterns by assessing the stratospheric pathway indirectly linking the Arctic and midlatitudes. By running modeling simulations where vertical wave propagation in the stratosphere is damped, they found the tropospheric response to sea-ice loss is weaker and that internal stratospheric dynamics impact the stratosphere-troposphere linkages.
Zhang et al. (2018a, 2018b) furthered investigated possible stratospheric connections to sea-ice loss, through a focus on regional variations in sea-ice. Both studies found that the upward wave forcing from reduced sea-ice in the Barents-Kara seas act in phase with a climatological ridge across Siberia, making this region even more impactful than other regions across the Arctic in forcing perturbations to the stratosphere. They also found that the typical tropospheric response directs cold air just downstream of the original perturbation, making the region of greatest sea-ice anomalies critical for determining the location of ridge building and downstream cold-air outbreaks.

Additional work highlights the two-way coupling where high-latitude ridges and/or blocking patterns can impact the state of the stratosphere and Arctic. Cohen et al. (2007, 2014) focused on the two-way coupling between radiative cooling over Siberia, ridges over Siberia, and the state of the stratosphere later on in winter. Fig.6 from Cohen et al. (2007), shows schematically how variability in autumn snow cover over Siberia can potentially perturb the polar vortex and lead to downward propagation of geopotential anomalies months later in winter. In all, the key physical mechanism in this relationship is poleward directed heat fluxes, which are proportional to the vertical component of the Eliassen-Palm (EP) flux (Eliassen and Palm 1961). Martius et al. (2009) furthered this work and examined the tropospheric conditions leading up to all sudden stratospheric warming (SSW) events, defined as negative (easterly) values of the zonal-mean zonal wind at 60°N and 10-hPa. In this study, 25 of 27 observed SSW events had an observed blocking pattern 10 to 0 days before the start date of the SSW. In addition, Martius et al. (2009) found that blocking patterns across the Pacific basin result in a splitting of the polar vortex, while blocking patterns in the Atlantic result in a displacement of the vortex off of the North Pole.
While Arctic and polar stratospheric conditions can have large impacts on midlatitude flow patterns, on the S2S timescale tropical variability like ENSO and the MJO offer the most potential predictability (Hoskins 2013). Renwick and Wallace (1996) found that the state of ENSO shifted the maxima in the 500-hPa geopotential height variance, shifting the predominant flow patterns across the Pacific on a seasonal timescale. During El Niño years, there is an overall reduced blocking frequency near the Bering Strait compared to ENSO-neutral or La Niña years, with increased ridging across western North America during El Niño years. ENSO can influence anomalies of the background state of the mean flow pattern over the North Pacific for around a year, but MJO can vary much more within an individual season. Lin and Brunet (2018) conducted a study to investigate the midlatitude response to the MJO, finding the central and eastern North Pacific basin has the strongest variance in the response to the MJO. Lin and Brunet (2018) also showed the state of the North Pacific jet stream influences the midlatitude response to an MJO event. They show that a North Pacific jet stream with stronger maximum wind speeds leads to an eastward shifted ridge-trough pattern, while an equatorward shifted jet stream produces an overall stronger midlatitude response.

Because of the documented impact the MJO can have on midlatitude flow amplification, other studies have specifically looked at the linkage between blocking pattern occurrence and the MJO. Henderson et al. (2016) examined the distribution of blocking patterns across the Northern Hemisphere at various lag times associated with the MJO, testing to assess if these blocking frequencies are significantly different from climatology. Their results show that blocking frequency increases across the western North Pacific basin after phase 6 of the MJO, across western North America after phase 7, and across Alaska after phase 8. However, there were statistically significant decreases in blocking across the North Pacific basin during the other 5
phases, suggesting that every phase of the MJO has a statistically significant impact on blocking frequency somewhere across the North Pacific basin. Attard and Lang (2019) conducted similar analysis for blocking patterns, bomb cyclone occurrence (Sanders and Gyakum 1980), and extratropical transition of tropical cyclones (Evans et al. 2017), using the state of ENSO, MJO, and QBO. Using similar methodology, the only statistically significant variation in blocking across the North Pacific broadly was after phase 5 of the MJO across the eastern North Pacific. Attard and Lang (2019) reconcile this difference between their study and Henderson et al. (2016) by noting the differences in methodology that could have influenced the two sets of results.

1.2.4 Drought across Western North America

Anomalies in wintertime precipitation over western North America can greatly impact drought conditions in a region already prone to high variance in year-to-year precipitation. Gibson et al. (2020) uses combined EOFs to identify ridges associated with daily below-average precipitation, identifying three different ridge types in and around western North America and running statistical tests to identify remote drivers influencing these ridges. Once identified, analysis of the Pearson correlation coefficient was conducted for each ridge type and various remote drivers of these ridges, such as tropical convection, ENSO, and regional sea ice concentration. Overall, there were stronger correlations (~0.4) between these remote drivers and western ridges over the eastern Pacific Ocean, rather than ridges over western North America. These results offer regional perspective on relationships between early-season boundary conditions and ridge/drought frequency, but the authors explicitly state that additional work on
the dynamics and evolution of persistent ridges is needed, especially given that little correlation existed for some of the ridge types identified.

As also discussed in Gibson et al. (2020), much of the existing literature on drought occurrence across western North America focuses on sea surface temperature (SST) anomalies. Seager et al. (2015) discuss how the impact of SST anomalies across the entire Pacific can have a positive feedback on ridges associated with drought conditions. They found that local bottom boundary conditions in the North Pacific basin can allow ridges that are able to form to persist longer than they may in other conditions, prolonging the effects of initial perturbations to the flow. Seager and Henderson (2016) state that tropical SST anomalies alone can create flow perturbations similar to those observed during periods of drought across western North America, specifically the winter of 2013–2014. However, the magnitude of the modeled flow perturbation from just SST anomalies is insufficient to properly represent observed geopotential height anomalies; midlatitude processes such as shifts in the storm track and eddy forcing for flow amplification are needed to fully account for these discrepancies. Teng and Branstator (2017) highlight two regions in the tropical northwest Pacific Ocean (near Taiwan and Papua New Guinea) where latent heat release associated with enhanced convection is most effective at leading to flow perturbations associated with drought across western North America, demonstrating again that tropical variability can greatly impact precipitation patterns over western North America.

1.2.5 Atmospheric River Impacts across Western North America
Western North America has the highest year-to-year variance in precipitation, with most of their precipitation falling in conjunction with ARs (Dettinger 2011). While well captured in most operational weather forecast models, AR forecasts on the S2S timescale rely almost entirely on atmospheric teleconnections, which in turn informs seasonal precipitation forecasts for western North America. Mundhenk et al. (2016) constructed a year-round climatology of AR occurrence across the North Pacific basin, using IVT anomalies and a feature-based detection algorithm to identify ARs. They show that El Niño conditions are typically associated with a reduced AR frequency around 45°N, while La Niña conditions favor an overall increase in ARs stretching from the tropical West Pacific to Washington and British Columbia. In addition, all 8 phases of the MJO have statistically significant changes in AR frequency associated with them, but none of these MJO phases by themselves had statistically significant anomalies in landfalling ARs across western North America. Mundhenk et al. (2018) extended this research to include the construction of joint teleconnection values of the MJO and QBO that were evaluated using the Heidke skill score (van den Dool 2007) values at various forecast lead times. These results show that there is indeed statistically significant model forecast skill at 5-week lead times, but that this significance is limited to very specific combinations of MJO and QBO, and these combinations varied as a function of latitude along western North America.

To directly link ARs to weather conditions across western North America, Payne and Magnusdottir (2014) investigated the characteristics of ARs that make landfall across western North America. Using cool-season ARs in their work, they found similar latitudinal shifts in ARs associated with each phase of ENSO and the MJO when compared to Mundhenk et al. (2016) and their year-round climatology. The unique aspect of Payne and Magnusdottir (2014) is composites of dynamics relevant to landfalling AR, such as potential vorticity, sea level pressure,
and 200-hPa wind speed. This study also stratified the strongest and weakest landfalling ARs, and this stratification shows that stronger ARs originate from further west in the North Pacific basin, have a stronger subtropical anticyclone to the south of the AR, and an increased frequency of anticyclonic Rossby wave breaking (AWB) as the AR makes landfall. These findings have implications for upstream regime impacts on landfalling ARs, and link dynamical features like the subtropical jet stream with the structure and evolution of ARs. Additional work by Gonzales et al. (2020) found that ARs with stronger low-level winds have greater precipitation efficiencies once they make landfall, compared to ARs with the same IVT but weaker winds. This finding suggests how an AR interacts with an upstream jet entrance regions modifies the regional impact of said AR.

1.3 Research Goals and Thesis Structure

1.3.1 Research Goals

There are four main goals of this research: 1) revisit and build upon previous methods of identifying persistent anomalies to construct an updated climatology of persistent ridge regimes; 2) illustrate and diagnose the relevant dynamic and thermodynamic drivers for persistent ridges throughout the ridge regime; 3) link the results of synoptic-dynamic analyses of ridge regimes to statistical relationships of ridge regimes, teleconnections, and AR frequency; and 4) provide additional understanding of downstream weather impacts through a case study analysis of a representative long-lasting ridge regime.

1.3.2 Thesis Layout
The layout for the remainder of the thesis is as follows. Data and methodology are discussed in chapter 2. Climatological results of ridge regimes, as well as composite figures describing the evolution identified ridge regimes are discussed in chapter 3. Statistical analysis linking ridge regimes to AR activity, North Pacific jet variability, and teleconnections are discussed in chapter 4. Chapter 5 discusses and explains a detailed analysis of the longest-lasting ridge regime in western North America during December 1980 and January 1981. Finally, discussion, conclusions, and future work suggestions will be discussed in chapter 6.
2. Data and Methodology

2.1 Persistent Ridge Identification Method

The National Centers for Environmental Prediction/National Center for Atmospheric Research (NCEP-NCAR) reanalysis dataset (Kalnay et al. 1996) is used for identifying persistent ridges and constructing composite analyses of said ridges. In order to identify persistent anomalies, a modified version of the criteria proposed by Dole and Gordon (1983) was implemented, using data from December, January, and February from 1948 to 2017. The original Dole and Gordon (1983) methodology requires that the 500-hPa geopotential height anomalies be at or above some height threshold (50, 100, or 150 meters) for a certain period of time (5, 10, or 15 days). Our modified version of the methodology requires that the 500-hPa geopotential height anomaly exceed one standard deviation continuously for at least 7 days. The mean and standard deviation used to create the standardized anomaly is defined using two 35-year base states: 1948–1982 and 1983–2017, to account for increasing heights due to anthropogenic climate change (Christidis and Stott 2015). This change from the original criteria was similarly used by Miller et al. (2020), allowing for persistent ridges to be identified in regions with smaller variance in the geopotential height field. Fig 2.1, a modified version of Fig. 1 from Dole and Gordon (1983), is shown to explain schematically how this identification method works.

After examining the results of the 70-year persistent anomaly identification conducted from December 1948 to February 2018, four domains were subjectively created in and around western North America, shown in Fig. 2.2 below. While subjectively derived, these domains encapsulate the overall maxima in persistent anomaly occurrence near western North America.
These domains also overlap quite well with the identified maxima in Dole and Gordon (1983), as well as Gibson et al. (2020) who used EOF analysis to create similar domains in their analysis.

One shortcoming of the Dole and Gordon (1983) methodology is that the criteria and output is only based on persistent anomalies in individual grid cells. For our research, we are interested in larger-scale persistent ridges, not just individual persistent anomalies. This required a method of linking persistent anomalies into identifiable ridges. Using the four domains described above, persistent anomalies that occur within the same domain and overlap in time with a previous persistent anomaly are combined into one persistent ridge event. As an example, if on day 5 of persistence for a given grid cell a new grid cell within the same domain became persistent for 7 days, the two instances would be combined and labeled as a persistent ridge that lasts for 12 days. This additional criterion helps to transform persistent anomalies into persistent ridge regimes relevant to synoptic-dynamic analysis, as well as allowing for distinct start and end dates of these regimes.

2.2 Composite Analysis

Having the start and end dates of the persistent ridge regimes identified allows for composite analysis of these persistent ridge regimes to be conducted. Composites of relevant synoptic-dynamic fields were composited starting with the start of ridge persistence, and continued throughout the ridge lifecycle. Because there is a large range for how long each regimes persists, intermediate composites are generated as a function of percentage of the ridge duration (e.g. 20% through the lifecycle of a persistent ridge event). This choice of a normalized composite time is motivated by Fig. 2.3, showing the cumulative distribution function (CDF) of
ridge regimes for all four domains. As shown, most persistent ridge regimes just barely meet the 7-day criteria and all four domains have at least of events last 7–10 days long. By using the normalized ridge duration as our composite method, this study avoids potentially erroneous compositing of the end of an 8-day persistent ridge regime with the midpoint of a 16-day persistent ridge regime or the beginning of a 32-day ridge regime. In addition to this composite of all persistent ridge events subset of ridges that exceed the 80th percentile in duration to assess how longer-lasting persistent ridge regimes may vary in their lifecycle compared to all ridge regimes. This is again motivated by the skewed distribution of the CDF with many persistent ridge regimes that last from 7–10 days, with a smaller subset that can last between 30 and 45 days.

Three sets of composite maps are generated for the duration of the lifecycle of the composite persistent ridge regimes, starting at the ridge start time continuing to the ridge end time at intervals of 20% of the ridge lifecycle. The first map has plotted 500-hPa geopotential height (black contours), integrated vapor transport (IVT, warm shading)(dashed line), and mean level pressure (cool shading). IVT is calculated as the integrated product of the wind and moisture content at each vertical level from 1000-hPa to 300-hPa. For the 80th percentile cases, stippling is included for regions with significantly higher low-level averaged (1000–700-hPa) moisture flux convergence (MFC, Banacos and Schultz 2005) based on bootstrap resampling. This calculation acts as a proxy for low-level latent heating similar to integrated MFC used in Torn and Hakim (2015). This calculation of layer-averaged MFC tests for where longer-lasting ridge composites have statistically higher amounts of latent heating in proximity to the persistent ridge. The second map has 1000–500-hPa thickness (red/blue lines), 700-hPa relative vorticity (shading), and integrated vapor transport (dashed line). The third panel has 700-hPa geopotential
height (black contour), 500-hPa Eady growth rate (EGR, shading), and 250-hPa winds (vector). The 500-hPa Eady growth rate is calculated with a modified version defined in Lindzen and Farrell (1980) and Hoskins and Valdes (1990):

$$\sigma = 0.31 f \left| \frac{\partial v}{\partial z} \right| \left( \frac{g}{\theta_e} \frac{d \theta_e}{dz} \right)^{-1/2}$$

where \( f \) is the Coriolis parameter, \( \left| \frac{\partial v}{\partial z} \right| \) is the local vertical wind shear between 450-hPa and 550-hPa, and \( \left( \frac{g}{\theta_e} \frac{d \theta_e}{dz} \right)^{-1/2} \) is the moist Brunt-Väisälä frequency at 500 hPa calculated using the equivalent potential temperature.

2.3 AR Frequency Analysis

Along with the above composite maps, heat maps of AR frequency anomalies observed before, during, and after persistent ridge regimes in each domain will be created. Three panels for each domain will be created: 10 days before ridge start up to ridge start, the period between ridge start and ridge end, and ridge end up to 10 days afterward. These periods are selected because of the interest in impacts of persistent ridge regimes on AR frequency on the subseasonal-to-seasonal (S2S) timescale. AR dates and locations are selected from an AR catalog created using the NCEP-NCAR reanalysis dataset, using an algorithm outlined in Gershunov et al. (2017). Continuous areas of IVT in excess of 250 kg m\(^{-1}\) s\(^{-1}\) are identified, and landfalling ARs are categorized if they cross the coastline of North America from 20–60°N and exceed 1500 kilometers in length at the time of landfall.

Calculations of the odds ratio are performed as the metric to assess the change in AR activity (Wilks 2011, Moore et al. 2019). The odds ratio in this application is defined as:

$$\text{Odds} = \frac{P \cdot (1 - P)}{Q \cdot (1 - Q)}$$

where \( P \) is the probability of an event occurring and \( Q \) is the probability of the event not occurring.
The ratio is defined as $\frac{P(A|R)[1-P(A|N)]}{P(A|N)[1-P(A|R)]}$, where $P(A|R)$ is the probability of an atmospheric river occurring given a ridge is present in one of the four domains, and $P(A|N)$ is the probability of an atmospheric river occurring given a ridge is not present in one of the four domains. This calculation is conducted for the three periods described above to assess the AR-ridge relationship before, during, and after a persistent ridge regime. An odds ratio greater than one indicates an increased likelihood of AR occurrence given the presence of a persistent ridge; an odds ratio less than one indicates a decreased likelihood of an AR occurrence given the presence of a persistent ridge. These calculations will identify where ARs are more or less likely to occur before, during, and after persistent ridge regimes.

### 2.4 Relationships of Ridges and Atmospheric Indices

In addition to the relationship between landfalling ARs and persistent ridge regimes, linkages between persistent ridge regimes and additional atmospheric indices are assessed. This allows for a similar proxy of important features of the general circulation for time periods extending before and after a ridge regime to accompany the composite analysis. The indices used for this section of analysis include the Arctic Oscillation (AO, Thompson and Wallace 1998), the Madden–Julian Oscillation (MJO, Wheeler and Hendon 2004), and the state of the North Pacific jet stream (NPAC jet phase, Winters et al. 2019). The Real-time Multivariate MJO (RMM) index for the MJO accounts for the strength and position of tropical convection, so only dates with an active MJO (absolute value greater than 1 for the total RMM index) are considered in this
analysis. Both the AO and NPAC jet phase are derived from EOF analysis so neutral states represent climatological conditions, so all available dates are considered in this analysis.

Because the indices used in this section are a single value representing their atmospheric phenomena, analysis in this section will differ from the AR-ridge analysis. The phase of each of these indices are compiled when they co-occur within any of these five periods of a persistent ridge regime: 10 to 5 days before ridge start, 5 days before up to ridge start, between ridge start and ridge end, ridge end up to 5 days after, and 5 to 10 days after ridge end. Bar charts are created representing the frequency of each phase of the index co-occurring with a period in a persistent ridge regime lifecycle. To assess if this frequency is statistically significantly different from the cool season (November–March) climatology, bootstrap resampling is conducted (Wilks 2011, section 5.3.5). This process involves taking 10,000 random subsets with replacement of the atmospheric index dataset from November to March equal to the number of time intervals a persistent ridge regime occurs in one of the five time periods listed above. This creates a distribution of the number of times a phase of this index may occur in an equally sized sample. If the number of co-occurrences found before, during, or after the persistent ridge regime are outside the bounds of the 95th percentile (i.e. lower than the 2.5th percentile or greater than the 97.5th percentile), then that ridge-index relationship is said to be statistically significant compared to the climatology. All instances of statistical significance will be described as just significant in this section of the thesis for simplicity.

2.5 Case Study
Of the four domains outlined above, two of these domains (CAN and WUS) share their longest persistent ridge regime in common, lasting for just over a month. The persistence and geographic extent of the ridge regime is the motivation for selecting this case in late December 1980 and January 1981 for further analysis in this thesis. This particular persistent ridge regime not only lasted for around a month, but it featured noteworthy impacts on the surface weather all across the United States occurred due to the large-scale flow perturbations associated with this regime. Taubensee (1981) and Wagner (1981) both detail the anomalous surface weather conditions over the United States as a result of this persistent ridge regime. This period featured record warmth across Alaska and the mountain west, record cold along the eastern seaboard, and dry conditions for much of the US but especially in the central and eastern United States.

The European Center for Medium-Range Weather Forecasting Reanalysis 5th Generation (ERA5) is used for generating 6-hourly maps documenting the structure and evolution of this persistent ridge regime. Two sets of figures will be created in order to show the evolution of this long-lasting persistent ridge regime. The first set of figures will have mean sea level pressure (contour), 1000–500-hPa thickness (red/blue dashed contour), and 250-hPa wind speed (shaded). The second set of figures will have 500-hPa geopotential height (contour), integrated vapor transport (IVT, shaded), 250-hPa v-wind (red/blue dashed contour), and 400–600-hPa averaged ascent (green contour). The maps will be used to discuss the synoptic-to-planetary scale evolution of this persistent ridge regime. In addition to the maps described above, the state of the MJO and NPAC jet phase derived from the calculations in Winters et al. (2019) are used to link the statistical results from the previous section to this persistent ridge regime.
Fig. 2.1. Method for defining cases. A persistent positive (negative) anomaly case of duration $D$ satisfying criteria $(M, T)$ is defined at a point if the anomaly at that point exceeds (is less than) the threshold value $M$ for at least $T$ days. Examples are given for both positive anomaly cases ($M > 0$) and negative cases ($M < 0$). For this research, $M$ is one standard deviation above the mean and $T$ is seven days. [Fig. 1 and caption adapted from Dole and Gordon (1983)].

Fig 2.2. Domains created for grouping grid cell based persistent anomalies into persistent ridge events. For simplicity, domains will be subsequently labeled GOA (magenta, top left), CAN (red, top right), PAC (green, bottom left), WUS (blue, bottom right).
Fig. 2.3. The cumulative distribution function of ridge regimes for each domain, showing the percentile of cases reaching certain ridge durations.
3. Climatology and Composite Analysis of Persistent Ridges

3.1 Climatology of Events

Figure 3.1 shows the occurrence of the modified Dole and Gordon (1983) positive persistent anomaly criteria for seven non-overlapping decades within the period of interest from 1948 to 2017. The maximum number of positive anomalies in each decade (approximately 10 per decade) is roughly equivalent to the results of Dole and Gordon (1983), but each decade has a hemispheric maximum in a different geographic location. This variability in the location of most persistent anomalies is obscured in Dole and Gordon (1983) by having only one 14-year period of data. While outside the scope of this study, yearly-to-decadal variability in the climate system can impact the location of positive persistent anomalies and blocking patterns across the Northern Hemisphere.

Fig. 2.2 shows the domains used to compile individual persistent anomalies into persistent ridge regimes. Over the 70-year period, each domain has roughly equal number of cases of persistent ridge regimes. GOA and WUS (magenta and blue, respectively) have 120 cases, CAN (red) has 126 cases, and PAC (green) has 137 cases. As mentioned in chapter two, the lengths of these persistent ridge regimes are not uniformly distributed. Fig. 2.3 shows the cumulative distribution function (CDF) of the duration of persistent ridge regimes for all four domains. In all domains, most cases just barely meet the seven-day persistence criteria and at least half of all cases in each domain last between 7 to 10 days. However, there is a subset of cases that persist from 30 days up to 45 days and create a skewed distribution. The skewness of these distributions is a function of longitude, as the CDFs of the oceanic domains (GOA and PAC) have a shallower slope than the CDFs of the continental domains (CAN and WUS). The
oceanic basins are located in a region near the climatological exit region of the North Pacific jet stream, suggesting that dynamics for continued ridge maintenance in the jet exit region mentioned in chapter 1, such as repeated cyclogenesis upstream, eddy heat flux, and eddy vorticity flux, allow for more cases to persist longer in these western domains.

3.2 Composite of Events

3.2.1 Gulf of Alaska (GOA) Composites

The following section will discuss composites of each domain for all ridge regimes, as well as the subset composite for the case that exceed the 80th percentile in duration for each domain. Fig. 3.2 shows a composite cyclone (MSLP < 1000 hPa) just to the west of the Dateline (180° longitude) throughout the lifecycle of the GOA ridge regime. The composite upstream cyclone slowly weakens with time, indicating upstream heat and vorticity fluxes associated with the cyclone are important to maintaining the composite GOA ridge as discussed in chapter 1. At the start of the ridge lifecycle (Fig. 3.2a), there is a northward extension of a composite atmospheric river (AR) on the western edge of the developing ridge around 150°W as seen in the integrated vapor transport (IVT) field. This extension diminishes through 40% of the ridge lifecycle, and then the main AR extends toward the coast of North America underneath the ridge by the end of the composite (Fig. 3.2f). As for the ridge itself, it amplifies through 40% of the ridge lifecycle (Fig. 3.2a–c), then slowly becomes negatively tilted and retrogresses out of the domain (Fig. 3.2d–f).

Fig. 3.3a shows Sutcliffe-Trenberth forcing for ascent (implied cyclonic vorticity advection by the thermal wind) upstream of the ridge and near the edge of the northward
extended AR. This proxy for vertical motion near the edge of the AR continues through 40% of the ridge lifecycle (Fig. 3.3c), implying diabatic heating is helping to build the ridge upstream. From 60% of the ridge lifecycle to the end, the 700-hPa anticyclonic relative vorticity becomes closed off over Alaska and slowly retrogresses toward Siberia (Fig. 3.3d–f), matching the tilt and retrogression seen in the 500-hPa geopotential height in Figure 3.2d–f.

Figure 3.4 shows a large baroclinic zone, or higher Eady growth rate (EGR) values, over and to the north of the North Pacific jet stream (NPJ) that extends to the Dateline. This region of high EGR overlaps with the poleward exit region of the NPJ, an area favorable for cyclogenesis. Along with the cyclones is the diffluence within the jet exit region that allows for anticyclonic eddy vorticity to be advected toward the western edge of the GOA persistent ridge. The signature and forcing for upstream cyclones is shown through all three sets of composites, further reiterating the importance of upstream cyclones in helping to perturb the flow and reinforce the ridge. Starting around 60% through the ridge lifecycle, the NPJ begins to undercut the ridge as the ridge becomes negatively tilted and retrogresses (Fig. 3.2). This separation of the ridge from the NPJ is coincident with a reduction in the EGR in the western North Pacific Ocean (WPAC), further signifying diminished instability for upstream cyclones toward the end of the ridge lifecycle.

Together, these composites reinforce a common evolution of the composite-mean ridge in the GOA domain: upstream cyclones form in the baroclinically-unstable NPJ poleward exit region; processes associated with the composite upstream cyclone and associated diffluent jet exit region (e.g. low-level warm air advection and anticyclonic eddy vorticity fluxes on the western flank of the ridge) help to build and maintain a persistent ridge regime from the start through around 40% of the lifecycle; cyclonic vorticity advection by the thermal wind near the
edge of an upstream AR implies a diabatic contribution to observed ridge maintenance; the NPJ begins to undercut the ridge as baroclinicity wanes in the WPAC; the ridge becomes negatively tilted, and eventually the ridge retrogresses toward Siberia. Many of these characteristics are common for other domains, so this description of the evolution of a persistent ridge regime will be considered the baseline for further composites. Any differing dynamics exhibited in composites for other domains will be noted when they differ from this baseline description.

In the 80th percentile cases for the GOA domain, Figure 3.5a shows two poleward-directed ARs near 150°W (similar to Fig. 3.2) and one closer to the composite cyclone near 180° (not seen in the full composite). Having two ARs present at the ridge start suggests that the upstream composite cyclone is a compositing artifact of several individual cyclones in each GOA ridge case. Having the AR near 180° also signifies that IVT is greater upstream of the longest-lasting GOA ridge regimes at the start of ridge persistence. Stippling, representing statistically higher low-level averaged (1000–700-hPa) moisture flux convergence (MFC), is concentrated on the northern and western edges of the ridge itself, most prominently at ridge start and 20% through the ridge lifecycle (Fig. 3.5a, b). Areas with stippling represent locations of greater diabatic contributions to ridge maintenance, and usually line up with implied vertical motion from cyclonic vorticity advection by the thermal wind shown in other composites. Another difference from the full composite is higher IVT values around 150°W and 30°N at 60 and 80% of the ridge lifecycle, a potential signature of cutoff low formation that will be discussed in more detail later on.

Figure 3.6 resembles Figure 3.3 in many aspects, with similar areas of cyclonic vorticity advection by the thermal wind immediately upstream of the composite persistent ridge located around 170°W. The composite relative vorticity is noisier than in the full composite because of
the smaller number of composite members (n=24) relative to the previous set of full composites (n=120). One notable difference is a local cyclonic vorticity maximum near 150°E and 30°N, matching up with the enhanced IVT mentioned in the previous paragraph. Enhanced cyclonic relative vorticity and IVT are signatures of cutoff lows, typically called Kona lows in this part of the world. Processes involved in ridge maintenance for longer-lasting GOA ridges, such as anticyclonic Rossby wave breaking (AWB), could lead to Kona low development (Caruso and Businger 2006).

Figure 3.7 shows a similar region of higher EGR values near the core of the NPJ, but more retracted than in the full composite in Fig. 3.4. In the equatorward jet exit region, there is a stronger northerly component to the wind around 180° than in the full composite, further evidence for increased likelihood of Kona lows during the intermediate stages of the ridge lifecycle (Fig. 3.6 c–e). The NPJ undercutting the composite GOA ridge toward the end of the lifecycle is weaker than in the full composite (Fig. 3.4), suggesting the longest-lasting ridges comprising this composite have more flow to the north of the ridge and less of a jet reformation underneath the ridge around 40°N.

3.2.2 Pacific (PAC) Composites

We now shift further south to examine composites of the PAC domain. Figure 3.8 shows the PAC composite persistent ridge is broader and does not amplify to the same extent as the GOA composite ridge does (Fig. 3.2). The PAC composite ridge does retrogress and shift northwest starting around 60% through the lifecycle (Fig. 3.8d), but it does not take on the same negative tilt when compared to the GOA composite ridge (Fig. 3.2d–f). Another difference from
the GOA composite is the rapid diminishing of IVT upstream of the PAC composite ridge around 40 and 60% through the ridge lifecycle (Fig. 3.8c, d). Along with weaker IVT, the composite upstream cyclone in Figure 3.8 is 5–10 hPa weaker than the corresponding composite upstream cyclone in Figure 3.2 for the entire length of the ridge lifecycle. A weaker composite upstream cyclone may suggest previously proposed mechanisms for forming and maintaining these persistent ridges, like low-level warm air advection and eddy anticyclonic vorticity fluxes, are weaker as well, meaning other mechanisms would be required to explain PAC persistent ridge formation and maintenance.

Figure 3.9 shows a pattern of cyclonic vorticity advection by the thermal wind upstream of the PAC ridge and near the edge of the previously mentioned upstream IVT corridor. Compared to Figure 3.3, the anticyclonic relative vorticity associated with the composite PAC ridge becomes weaker and less consolidated near the ridge center as the composite progresses toward the ridge end. Figure 3.10 shows a more retracted region of baroclinicity in the PAC composite than in the GOA composite, especially around 40% through the ridge lifecycle (Fig. 3.10c). There is also a less diffluent NPJ exit region near the Dateline when compared to the GOA composite, but the diffluence increases throughout the ridge lifecycle.

Figure 3.11, the first set of composites for the PAC 80th percentile subset, shows an overall similar pattern to the full composite in Figure 3.10. IVT upstream of the persistent ridge in the western North Pacific Ocean diminishes and shrinks in areal coverage overall, similar to what is shown in Fig. 3.10. Statistically significantly higher MFC values are seen to the north and west of the PAC composite persistent ridge, similar to the GOA 80th percentile composite in Figure 3.5. One notable difference from the full composite in Fig. 3.8 is the northwestern extension and retrogression of the ridge, first noticeable in Figure 3.11c. This northwest
extension as the PAC ridge evolves is more similar to the GOA composite (Fig. 3.2) than the PAC full composite (Fig. 3.8), and that the relevant dynamics for PAC ridge maintenance is more closely related to the GOA composite evolution than to the PAC composite evolution.

In Figure 3.12d, the northwestern extension and retrogression of the PAC composite ridge is evident, with two separate anticyclonic relative vorticity maxima around 50°N and 165°W (northwest extension) and 40°N and 140°W (main PAC ridge). Along with the westward shift in IVT seen in Figure 3.11, there is a corresponding westward shift of cyclonic relative vorticity to the northwest of the ridge as the ridge evolves (Fig. 3.14a–e). The westward shift of the composite cyclonic relative vorticity and IVT maxima suggests longer-lasting PAC ridges have a stronger coupling between the composite ridge and the composite upstream cyclone than in the full PAC composite. Figure 3.13 shows a consistent reduction of the maximum value of the EGR parameter upstream of the persistent ridge, as well as a westward shift of the leading edge of the EGR maximum. There is also stronger diffluence in the exit region of the NPJ in this composite of longer-lasting PAC ridges compared to the full PAC composite in Figure 3.10. The characteristics of this diffluence is also different from previous composites, as the northerly winds in the equatorward jet exit region are around 20° further west from the southerly winds in the poleward jet exit region.

3.2.3 Canada (CAN) Composites

Figure 3.14, the full composite of persistent ridge cases for the CAN domain, shows a stronger (~5 hPa lower), more longitudinally-elongated (~20° longer) composite upstream cyclone than either the GOA or PAC domains. This is a potential by-product of aliasing involved
with compositing, but could also suggest multiple cyclones are present upstream of the newly persistent ridge in some composite members (Fig. 3.14a). Case study analysis of a CAN ridge regime (shown later) also has a similarly elongated upstream cyclone, so the composite could be mostly representative of a unique flow characteristic associated with CAN composite ridges. Fig. 3.14a also shows a northward extension of the otherwise zonal IVT corridor around 140°W, a feature that persists through 40% of the ridge lifecycle (Fig. 3.14c). As the CAN ridge regime evolves, the ridge takes on a negative tilt similar to the GOA composite (Fig. 3.2), but the “in-situ” ridge building where the ridge amplifies from ridge start through 40% of the ridge lifecycle is less amplified in the CAN composite ridge (around 6 dam smaller 500-hPa geopotential height change).

Figures 3.15a–c shows that the upstream cyclonic relative vorticity is further east (around 145°W) compared to either the GOA or PAC domain. This mirrors the eastern extension of the composite upstream cyclone, and persists for the entire composite evolution of CAN ridge regimes. Despite the elongated region of cyclonic relative vorticity, the region of cyclonic vorticity advection by the thermal wind is mostly occurring along the eastern edge of the cyclonic vorticity maximum near 150°W. However, the northward extension of the IVT maximum begins to diminish around 40 and 60% through the lifecycle of the ridge (Fig. 3.17c, d), suggesting that there is less condensation, latent heating, and diabatic contribution to ridge maintenance upstream of the ridge toward the end of the ridge lifecycle. Another unique feature of the CAN domain is the anticyclonic relative vorticity associated with the ridge remains over the continent and does not retrogress as seen in Fig. 3.5 and 3.11, suggesting topography like the Rocky Mountains is helping to keep the ridge from retrogressing (Charney and Eliassen 1949).
Figure 3.16 shows the EGR maximum extends around 20° further east in the central North Pacific basin than the GPA or PAC composites (Fig. 3.4 and 3.10). The eastward extension of the EGR maximum is aligned with the eastward extension of the cyclonic relative vorticity at 700 hPa and the composite low pressure system upstream of the CAN ridge domain. The diffluent jet exit region in the eastern extent of the EGR maximum resembles the GOA full composite (Fig. 3.4) more than PAC full composite (Fig. 3.12), but shifted east along with the EGR maximum. Unlike Fig. 3.4 and 3.10, there is very little weakening or retraction of the area of maximum Eady growth rate as the CAN ridge regime evolves. Only in Fig. 3.16e does the embedded EGR maximum diminish in magnitude and retract slightly at the eastern edge. However, this composite CAN ridge regime maintains a high amount of baroclinicity in the western North Pacific basin compared to PAC and GOA.

Figure 3.17, the 80th percentile composites for CAN, shows MFC is significantly higher than the full composite to the northwest of the persistent ridge from the ridge start to around 40% of the ridge lifecycle (Fig. 3.17a–c). After 40% through the lifecycle, the significantly higher MFC pattern becomes less coherent. Features not seen in previous composites (Fig. 3.5, 3.11) are discrete ARs in succession embedded within the larger IVT corridor to the south and east of the cyclone center. These training ARs, along with pulses in the MSLP field throughout the composite, suggests that the composite cyclone around 160°W is made up of many individual cyclones throughout the ridge lifecycle. This would corroborate the findings from case study analysis that maintenance of persistent ridge regimes requires discrete ridge rebuilding events that are smoothed out in the full composite.

As mentioned for the full composite, Fig. 3.18 shows that CAN persistent ridge regimes have a less defined anticyclonic vorticity anomaly associated with ridge itself for most of the
composite evolution. Instead, anticyclonic vorticity remains closely tied to the windward side of the Rocky Mountains especially toward the end of the ridge composite (Fig. 3.18d–f), and does not retrogress throughout the ridge lifecycle as was shown in the GOA and PAC composites (Fig. 3.6 and Fig. 3.12). Fig. 3.18 also shows multiple discrete cyclonic vorticity maxima throughout the lifecycle 15° of longitude apart, further suggesting that there are multiple upstream cyclones that aid in ridge persistence. Fig. 3.18c and 3.18d shows a reduction in the implied cyclonic vorticity advection by the thermal wind between 40% and 60% through the ridge lifecycle. This matches the timing of the breakup of the significantly increased low-level MFC upstream of the ridge as shown in the previous paragraph and suggests the two processes are dynamically linked.

Figure 3.19a shows the typical corridor of higher EGR along the core of the NPJ at the beginning of the 80th percentile CAN composite. What is unique to this set of composites is the large increase to the east of the Dateline at 20% through the ridge lifecycle. The 700-hPa height gradient does not appear to change much from the ridge start, so this increase could be a result of either increased zonal wind or a decrease in static stability within the jet exit region. No matter the cause, the periodic pulses of increased baroclinicity further corroborate the idea that upstream cyclonic forcing on the ridge is periodic in nature. Toward the end of the ridge lifecycle, the EGR maximum retracts along with the jet exit region to the west of the Dateline but reintensifies in the western North Pacific basin. Along with the retraction of the jet exit region, the diffluent pattern associated with the jet exit region is more dominated by northerly flow on the anticyclonic shear side of the jet stream rather than southerly flow on the cyclonic shear side of the jet stream. The stronger northerly flow matches with the PAC composites but not the GOA
composites (Fig. 3.5 and 3.11), reinforcing the commonality of particular large-scale flow characteristics associated with each set of composites.

3.2.4 Western United States (WUS) composites

Shifting to our last domain (WUS), Figure 3.20 shows that the IVT corridor along the jet stream throughout the composite is the strongest and most extended compared to the full composites from the other three domains. Within the full composite there are embedded IVT maxima, further indication of individual progressive upstream cyclones not shown in the MSLP composite field. As the ridge regime progresses through the lifecycle, the ridge remains relatively stationary similar to the PAC composite in Fig. 3.8. Only in Fig. 3.20f does the ridge appear to become negatively tilted and shift outside of the WUS domain toward the northwest. Otherwise, there is more of a de-amplification of the main ridge within the WUS domain (Fig. 3.20d–f) while the northern extension of the ridge retrogresses.

Figure 3.21 shows strong Sutcliffe–Trenberth forcing for ascent upstream of the ridge along the coast of Alaska and British Columbia up to around 40% of the way through the ridge lifecycle. After this point (60% until the ridge end), the 1000–500-hPa thickness gradient weakens along with the magnitude of the 700-hPa relative vorticity, leading to weaker forcing for ascent in this same region. Similar to the CAN composites (Fig. 3.17), the anticyclonic vorticity associated with the ridge stays collocated with topography over western North America. The apparent impact of topography then suggests persistent ridges in the CAN and WUS domains are a superposition of the climatological ridge over western North America and an anomalous persistent ridge forced through mechanisms mentioned before.
Figure 3.22 shows a region of enhanced EGR extending past the Dateline, especially up to 40% through the lifecycle (Fig. 3.22a–c). This configuration matches very well with Fig. 3.16 and the CAN full composite, indicative of the further jet extension required for a persistent ridge to occur over western North America. Unlike Fig. 3.16, by 60% of the way through the ridge lifecycle (Fig. 3.22d), the maximum value of the EGR begins to diminish up to the end of the ridge lifecycle (Fig. 3.22f). Looking at the overall 250 hPa winds, there is weaker diffluence in the jet exit region and more of a continuous jet stream “up and over” the ridge. This is evident by the secondary EGR maximum over British Columbia at the crest of the ridge throughout the lifecycle, as stronger mid-level wind shear enhances the baroclinicity in this region. The weaker diffluence upstream of the composite WUS ridge suggests that there is also weaker eddy vorticity flux just upstream of the ridge, meaning the mechanisms required to maintain the persistent ridge throughout the lifecycle varies for each domain.

Finishing this section of analysis, we have the 80th percentile composites in figure 3.23. Unlike the other 80th percentile composites, the WUS has a coherent region of significantly higher low-level MFC to the north and west of the ridge throughout the entire ridge lifecycle. This enhanced diabatic heating near and immediately upstream of the ridge is the most extensive of all four of the 80th percentile composites, meaning diabatic heating may be contributing to ridge formation and maintenance throughout the entire lifecycle, as suggested for composites of all the longest-lasting ridge cases. However, the similar pattern of significantly higher MFC at the end of the ridge lifecycle (Fig. 3.23f) suggests that the cessation of enhanced diabatic heating is not enough to end the persistent ridge regime by itself. Along with the statistically higher low-level MFC upstream of the ridge, the magnitude and areal extent of upstream ARs is higher than the other 80th percentile cases from the beginning to around 60% through the ridge lifecycle (Fig.
As with other 80\textsuperscript{th} percentile composites, the WUS 80\textsuperscript{th} percentile composite shows several embedded IVT maxima throughout the lifecycle itself, again suggesting that multiple upstream cyclones are responsible for forming and maintain the persistent ridge regime over western North America. In addition, the WUS 80\textsuperscript{th} percentile composite mirrors the full composite (Fig. 3.20) in the lack of negative tilt of the ridge with time. Instead, the northern extension of the ridge shifts to the northwest toward Alaska while the southern component of the WUS ridge weakens (Fig. 3.23e, f). This composite shift of the ridge at the end of the lifecycle would suggest that a subset of WUS persistent ridge regimes become GOA or PAC ridge regimes, depending on the latitude. While not addressed in this thesis, persistent ridge regimes tracked into other domains could offer even longer windows of opportunity to make forecasts into the seasonal timescales.

Similar to Fig. 3.21, Fig. 3.24 shows a broad region of forcing for upward vertical motion upstream of the persistent ridge regime throughout the ridge lifecycle from cyclonic vorticity advection by the thermal wind. The upstream cyclonic relative vorticity is strongest from the ridge start to 60\% of the way through the ridge lifecycle (Fig. 3.24a–d), and diminishes by the end of the ridge lifecycle (Fig. 3.24e, f). As stated in the previous paragraph, multiple maxima of cyclonic vorticity appear in the 80\textsuperscript{th} percentile composite and add evidence for multiple cyclones being responsible for forming and maintaining the WUS persistence ridge regime. Unlike previously examined regional composites, three separate anticyclonic vorticity maxima appear over the Great Basin, the border of Washington and British Columbia, and off the coast of northern California from 40\% through 80\% of the ridge lifecycle (Fig. 3.24c–e). The local maximum in anticyclonic relative vorticity off the coast gradually shifts north with time, reflecting the separation of the northern branch of this ridge composite toward the end of the
ridge lifecycle. The interior anticyclonic vorticity maxima diminish with time while remaining in place, suggesting these structures are dynamically tied to topography in a similar way to the CAN ridge composites.

The final composite, Fig. 3.25, shows the highest values of EGR of all of the composites in the western and central North Pacific basin toward the beginning of the ridge lifecycle. This region of enhanced baroclinicity remains extended across the Dateline up until the ridge end (Fig. 3.25f), where the eastern extent of the EGR maximum shifts to just west of the Dateline and the maximum value of the EGR is reduced from the start of ridge persistence. This reduced and shifted upstream baroclinicity again suggests that the end of a persistent ridge regime often occurs when upstream cyclones are no longer able to maintain the ridge and the ridge shifts to the northwest. As mentioned for Fig. 3.22, there is a secondary EGR maxima near the crest of the ridge over western Canada, meaning any lee cyclogenesis occurring in this region could develop more quickly and result in weather impacts downstream of the persistent ridge regime. Another similarity to the full WUS ridge composite (Fig. 3.22) is the offset between the diffluent flow in the jet exit region. The northwesterly flow at 250 hPa is consistently further west than the corresponding southwesterly flow on the periphery of the ridge. The overall flow pattern around the ridge is similar to the full composite (Fig. 3.22), meaning eddy vorticity flux may be contributing less to ridge maintenance for all WUS ridges than for persistent ridges in other domains.
Fig. 3.2. Composite 500-hPa geopotential height (contour, m), integrated vapor transport (warm shading, kg m\(^{-1}\) s\(^{-1}\)) (dashed contour 250 kg m\(^{-1}\) s\(^{-1}\)), sea level pressure (cool shading, hPa) for all GOA ridge regimes at (a) ridge start, (b) 20% of ridge duration, (c) 40% of ridge duration, (d) 60% of ridge duration, (e), 80% of ridge duration, and (f) ridge end.

Fig. 3.3. Composite 1000–500-hPa thickness (red/blue contour, m) and 700-hPa relative vorticity (10\(^{-5}\) s\(^{-1}\)) for all GOA ridge regimes (a) ridge start, (b) 20% of ridge duration, (c) 40% of ridge duration, (d) 60% of ridge duration, (e), 80% of ridge duration, and (f) ridge end.
Fig. 3.4. Composite 700-hPa geopotential height (contour, m), 500-hPa Eady growth rate (shading, day⁻¹), and 250-hPa wind (vector, m s⁻¹) for all GOA ridge regimes (a) ridge start, (b) 20% of ridge duration, (c) 40% of ridge duration, (d) 60% of ridge duration, (e) 80% of ridge duration, and (f) ridge end.
Fig. 3.5. As in Fig. 3.2 but for the 80th percentile in duration GOA ridge regimes. Stippling indicates where 1000–700-hPa averaged moisture flux convergence is statistically significantly higher for the 80th percentile ridge regimes than all GOA ridge regimes.
Fig. 3.6. As in Fig. 3.3 but for the 80th percentile in duration GOA ridge regimes.
Fig. 3.7. As in Fig. 3.4 but for the 80th percentile in duration GOA ridge regimes.

Fig. 3.8. As in Fig. 3.2 but for all PAC ridge regimes.
Fig. 3.9. As in Fig. 3.3 but for all PAC ridge regimes.
Fig. 3.10. As in Fig. 3.4 but for all PAC ridge regimes.

Fig. 3.11. As in Fig. 3.5 but for the 80th percentile in duration PAC ridge regimes.
Fig. 3.12 As in Fig. 3.6 but for the 80th percentile in duration PAC ridge regimes.
Fig. 3.13. As in Fig. 3.7 but for the 80th percentile in duration PAC ridge regimes.

Fig. 3.14. As in Fig. 3.2 but for all CAN ridge regimes.
Fig. 3.15. As in Fig. 3.3 but for all CAN ridge regimes.
Fig. 3.16. As in Fig. 3.4 but for all CAN ridge regimes.

Fig. 3.17. As in Fig. 3.5 but for the 80$^{th}$ percentile in duration CAN ridge regimes.
Fig 3.18. As in Fig. 3.6 but for the 80th percentile in duration CAN ridge regimes.
Fig. 3.19. As in Fig. 3.7 but for the 80th percentile in duration CAN ridge regimes.

Fig 3.20. As in Fig. 3.2 but for all WUS ridge regimes.
Fig. 3.21. As in Fig. 3.3 but for all WUS ridge regimes.
Fig. 3.22. As in Fig. 3.4 but for all WUS ridge regimes.

Fig. 3.23. As in Fig. 3.5 but for the 80th percentile in duration WUS ridge regimes.
Fig. 3.24. As in Fig. 3.6 but for the 80th percentile in duration WUS ridge regimes.
Fig. 3.25. As in Fig. 3.7 but for the 80\textsuperscript{th} percentile in duration WUS ridge.
4. Statistical Relationship of Persistent Ridges to Atmospheric Phenomena

4.1 Ridge Relationship with Atmospheric Rivers

Figure 4.1 shows the odds ratio of an AR making landfall along the coast of western North America before, during, and after a persistent ridge in the GOA domain. Figure 4.1a shows a broad area of reduced likelihood of landfalling ARs, especially between 30°N and 40°N, in the period of 10 days prior to persistent ridge onset. Figure 4.1b, representing the period during all GOA persistent ridges, shows the area most affected by AR reduction during the GOA ridge regime has shifted from 40°N to 60°N. The darkest brown shading from 50°N to 57.5°N in Fig. 4.1b, an odds ratio of 0–0.2, means ARs are more than 5 times less likely to make landfall when there is a GOA persistent ridge present than corresponding times when there is not a GOA persistent ridge present. Further south, there is a swath from 25–32.5°N where ARs are slightly more likely to make landfall when a GOA persistent ridge is occurring. While the reduction in AR landfall immediately downstream of the persistent ridge makes intuitive sense, the slight increase in odds ratio further south hints at the occurrence of AR activity along the subtropical jet stream (STJ). In the period up to 10 days after a GOA persistent ridge, the AR odds ratio is less than one for most of western North America and especially north of 45°N and south of 30°N, with values around or above 1 from 30–45°N. While only slightly above 1, the region of slightly above average likelihood of AR landfall from 30–45°N matches with the GOA composite in Figure 3.3 and the North Pacific jet and corresponding composite IVT corridor undercutting the GOA persistent ridge by the end of the ridge lifecycle.
Figure 4.2a shows a fairly uniform reduction in AR landfall across western North America in the 10 days before a PAC persistent ridge forms, with little variation along the coast of western North America. A nearly universal AR odds ratio value of 0.6 across western North America suggests that the flow patterns up to 10 days prior to a persistent ridge can impact AR landfall and reduce precipitation on the subseasonal timescale. During PAC persistent ridges, Figure 4.2b shows that the odds ratio is greatly reduced north of 35°N, with odds ratio values slightly below 1 south of 35°N. Much like Fig. 4.1b, persistent ridges in the two western domains (GOA and PAC) have a common characteristic of near climatological AR landfall over southern California and northern Mexico. Given the longitude of the composite ridge axes shown in Fig. 3.2 and 3.8, confluent upper-level flow is expected near 145°W and could allow for STJ formation during a GOA and PAC persistent ridge. In the 10 days following a PAC persistent ridge, shown in Fig. 4.2c, there is again a fairly uniform reduction in the odds ratio of a landfalling AR over western North America. Unlike Fig. 4.2a (before PAC persistent ridge onset), the AR odds ratio values north of 50°N are slightly higher than the values in the region south of 35°N.

Fig 4.3a shows two local minims in AR odds ratio along western North America in the 10 day period before a CAN persistent ridge forms: north of 50°N and south of 32.5°N. While no composite maps were constructed for before or after the persistent ridge events, the near-climatological values of AR occurrence between 32.5°N and 50°N could suggest a zonal extension of the NPJ that could allow for relatively more AR landfalls in that zone from 32.5–50°N. Fig. 4.3b, the period during a CAN persistent ridge, shows the AR odds ratio is much lower (<0.4) from 40–55°N along the western coast of North America. South of 40°N, there are AR odds ratio values mostly between 0.6 and 0.8, representing a slight reduction in AR landfalls.
In the northern edge (57.5°N) of this domain, there is a small area of odds ratio values greater than 1 along the coast of Alaska, indicative of ARs (and cyclones more broadly) being deflected “up and around” the persistent ridge over western Canada. Fig. 4.3c, representing the 10 days following a CAN persistent ridge, once again mirrors the distribution from Fig. 4.3a with greater reduction in AR odds ratio values north of 50°N and south of 30°N compared to the region from 30–50°N. When compared to Fig. 4.3a, Fig. 4.3c has greater reduction in AR odds ratio in the areas north of 50°N and south of 30°N, suggesting the period after a persistent ridge has a greater impact on AR activity than the period before a persistent ridge. Referencing the previous section, there is a coherent ridge upstream of the CAN domain at the time of ridge end in Fig. 3.14 as it retrogresses out of the domain. The period after a CAN persistent ridge may have smaller case-to-case variability for the upstream flow pattern than the corresponding period before a CAN persistent ridge, which could be an explanation for this asymmetry in AR frequency changes.

In the period 10 days prior to a WUS persistent ridge forming, Fig. 4.4a shows a broad reduction in the odds ratio of AR landfall across western North America. Similar to Fig. 4.3, the AR odds ratio is slightly higher to the north of 50°N and to the south of 30°N, while the coastline from 30–50°N has odds ratio values closer to climatology. During a WUS persistent ridge (Fig. 4.4b), AR odds ratio values are less than one at all locations along western North America from 22.5–50°N, with the greatest decrease from 27.5–40°N. North of 50°N, the AR odds ratio values are greater than one, similar to the pattern in Fig. 4.3b for CAN persistent ridges. This pattern matches with synoptic intuition of these events; storms and ARs are typically deflected by a persistent ridge leading to greater likelihood of AR landfall to the north of a persistent ridge and decreased likelihood of AR landfall within and to the south of a persistent ridge. The AR odds ratio values for the 10 days following a WUS persistent ridge, as shown in Fig. 4.4c, are below
0.6 to the north of 37.5\degree N and near climatology (~1) to the south of 37.5\degree N. The AR odds ratio values in the 10-day period following a WUS persistent ridge in Fig. 4.4c mirrors the odds ratio values from during a GOA persistent ridge in Fig. 4.1b. As stated in chapter 3, the common evolution for a WUS persistent ridge is for a northern branch of the persistent ridge to retrogress toward the Gulf of Alaska, so this could be another indicator that persistent ridges can occur in two separate domains sequentially. Further work tracking persistent ridges using an object-based definition could allow for analysis that would classify persistent ridges that exist in multiple domains over the course of a “ridge object” lifecycle.

4.2 Ridge Relationship with Atmospheric Indices

4.2.1 Ridge Relationship with the Arctic Oscillation (AO)

Figure 4.5 shows the frequency of different states of the AO (Thompson and Wallace 1998) for periods before, during, and after a GOA persistent ridge. Between 10 to 5 days before a GOA persistent ridge, positive (>0.5) and negative (<-0.5) values of the AO are around twice as likely than AO neutral conditions. However, this period has no significantly anomalous AO conditions at the 95% confidence interval. In the 5 days immediately before the GOA persistent ridge starts, positive AO conditions occur almost 10% more often than negative AO conditions. These values are also statistically significant, with the frequency of positive AO conditions being significantly higher than the cool-season climatology and the frequency of negative AO conditions being significantly lower than the cool-season climatology. During a GOA persistent ridge, all three phases of the AO are statistically significant; AO negative and AO neutral are significantly less likely to occur than climatology, and AO positive is statistically more likely to
occur than climatology. In the five days following a GOA persistent ridge, AO negative conditions are statistically more likely to occur than climatology at a frequency of around 50%. At the same time, AO positive conditions decrease to a frequency of around 30% and become statistically less likely to occur than climatology. In the period 5 to 10 days following a GOA persistent ridge, AO negative conditions are still more likely than AO positive, but none of the three AO states are significantly different from climatology.

Fig. 4.6 shows the period 10 to 5 days before a PAC persistent ridge has higher frequency of AO negative than AO positive, but with no statistical significance for the AO frequencies. In the five days prior to a PAC persistent ridge, AO positive conditions are slightly more frequent than AO negative conditions but not enough for these frequencies to be significantly higher (positive) and lower (negative) than climatology. The distribution of AO frequencies from the previous period stays the same during a PAC persistent ridge, as AO positive conditions increase to a frequency of over 50% and the frequency remains significantly higher than climatology. Both AO negative and AO neutral conditions decrease in frequency from the previous period and remain or become significantly lower in frequency than climatology. In the two five-day periods following a PAC persistent ridge, AO negative once again becomes more frequent than AO positive. However, none of the three states of the AO are significantly different from climatology.

Of the four domains, CAN persistent ridges are associated with similar AO frequencies before, during, and after the persistent ridge itself, making it an outlier from the other domains that often shift between AO positive and AO negative conditions. In Fig. 4.7, AO negative conditions occur at least 50% of the time at all stages before, during or after a CAN persistent ridge and are always significantly higher than climatology. Similarly, AO positive conditions are
always below 30% in frequency and always significantly lower than climatology. For AO neutral conditions, the only period of statistical significance is during the CAN persistent ridge itself. The relationship between CAN persistent ridges and the AO is the only one of the four domains to have statistically significant differences up to 10 days prior to ridge onset and 10 days after ridge end. Looking at the composite maps from chapter 3 (Fig. 3.14, Fig. 3.20), the elongated composite cyclone upstream of the CAN persistent ridge projects almost directly onto the positive portion of the EOF used to calculate the AO. Given the consistency of significantly higher occurrence of negative AO values, the composite upstream cyclone in the CAN composite is likely to form well before the persistent ridge forms and lingers past when the CAN persistent ridge dissipates.

Fig. 4.8, a plot of the ridge-AO relationship for WUS persistent ridges, has AO negative conditions more likely to occur than AO neutral or AO positive conditions in the period 10 to 5 days before a WUS persistent ridge but without any statistical significance for the three AO phases. In the five days preceding a WUS persistent ridge, AO neutral conditions increase in frequency and become statistically higher than climatology, but still less than either AO positive or AO negative conditions. During a WUS ridge, AO positive conditions become more frequent than AO negative, a similar shift seen in the GOA and PAC analyses as well (Fig. 4.5, 4.6). Looking at the composite maps from chapter 3, the GOA, PAC, and WUS composites have stronger composite cyclones in the North Atlantic, while the CAN domain predominantly has a stronger composite cyclone over the North Pacific. The asymmetry in the AO values across the four domains suggests significant downstream differences in the hemispheric flow pattern to these ridges, a topic worth exploring in future work. This higher AO positive frequency and lower AO negative frequency during WUS persistent ridges are significantly higher and lower
than climatology, respectively. In the five days following a WUS persistent ridge, the AO positive frequency remains significantly high and AO neutral conditions are significantly lower than climatology. For the period 5–10 days following a WUS persistent ridge, AO positive and AO negative conditions are nearly equal in frequency, but the AO positive conditions are significantly higher than climatology. As stated in chapter 3, there is evidence that some WUS persistent ridges may retrogress and remain persistent in either the GOA or PAC domains. Looking at the period after WUS persistent ridge end and before/during a GOA or PAC persistent ridge, the AO frequency distribution looks qualitatively similar. While analysis on exact ridge tracking would be needed to definitively prove that WUS persistent ridges can be connected to GOA or PAC persistent ridges, this is another line of evidence that the connection may exist.

**4.2.2 Ridge Relationship with the Madden–Julian Oscillation (MJO)**

The next set of figures show the frequency of different phases of the MJO in the same periods relative to persistent ridges in one of the four domains. Some changes are made when compared to the AO analysis, as inactive MJO periods are not considered and the frequencies are relative to all times when the MJO was outside the unit circle in the RMM index (Wheeler and Hendon 2004). Another change relative to the AO is that MJO data is only available back to 1974, so around half of all persistent ridge cases needed to be excluded in this analysis.

Fig 4.9 shows the relationship between the MJO and GOA persistent ridges throughout its lifecycle. In the two five-day periods before a GOA persistent ridge, very few of the MJO phases are significantly different from climatology. Only the frequency of phase 1 in the period
10 to 5 days prior to GOA persistent ridge onset was statistically lower than climatology. On days during a GOA persistent ridge, the distribution of MJO frequency becomes more widely statistically significant; the frequency of phases 1–3 are significantly lower than climatology while the frequency of phases 6 and 7 are significantly higher than climatology. While this study looks at MJO occurrence when a persistent ridge is present, the statistical differences during GOA persistent ridges agrees with Henderson et al. (2016). In their study, Henderson et al. (2016) finds enhanced blocking near the GOA domain during phase 7 of the MJO and reduced blocking during phases 1–3 (though not statistically significant). After the GOA persistent ridge, the MJO frequencies of each phase have limited statistical significance. In the five days immediately after a GOA persistent ridge, the frequency of phase 7 of the MJO remains significantly higher than climatology. In the 5 to 10 days following a GOA persistent ridge, phases 2 and 8 have significantly higher occurrences than climatology and the frequency of phase 5 is significantly lower than climatology. Overall, there is a coherent signal of the enhanced probability of the MJO propagating from phases 6–8 from the start of a GOA persistent ridge to 10 days after the end of a GOA persistent ridge.

In Fig. 4.10, the occurrence of the MJO in phases 6 and 7 preceding a PAC persistent ridge becomes significantly lower than climatology. The frequency of phase 6 is significantly lower than climatology starting 10 days before PAC ridge onset, while the frequency of phase 7 is significantly lower than climatology starting five days before PAC ridge onset. There is also a localized statistical increase in MJO occurrence in phase 8 from 10–5 days prior to PAC ridge onset, but this significant increase does not continue into other periods. During a PAC persistent ridge, the frequency of phases 5 and 7 are significantly higher than climatology and the frequency of phases 1 and 8 are significantly lower than climatology. Given that phases 5–7 of
the MJO are near the longitude of the jet entrance region of the NPJ, the relationship between the MJO and persistent ridges can be dynamically linked through the NPJ itself. In the five days following the end of a PAC persistent ridge, phases 5 and 6 have significantly higher frequency and phases 1 and 2 have significantly lower frequency. These significant frequency anomalies show a coherent progression through the phases of MJO, offering more certainty that this could be used as a source of predictability. However, the pattern of statistical significance shifts in the period 5 to 10 days after a PAC persistent ridge; phases 2 and 8 are now significantly higher than climatology and phase 5 is significantly lower than climatology. The pattern of statistical significance from this last time period deviates from the 5 days immediately after the PAC persistent ridge, suggesting that the MJO does not have one common phase after the end of PAC persistent ridge.

Fig. 4.11 shows the relationship between CAN persistent ridges and MJO frequency before, during, and after a persistent ridge. In the period 10–5 days prior to CAN ridge onset, phases 6 and 7 have significantly higher frequencies while phases 1 and 8 have significantly lower frequencies. Despite differences in methodology, Attard and Lang (2019) find a similar decrease in blocking over the eastern North Pacific and western North America up to two weeks after the MJO was active in phase 8, corroborating these results. For the five-day period immediately prior to CAN persistent ridge onset, the frequency of phase 7 of the MJO remains statistically higher than climatology while the frequency of phases 2 and 4 remain significantly lower than climatology. These statistically significant frequencies “propagate” in a coherent manner in this composite CAN persistent ridge analysis, a pattern not observed in Fig. 4.9 or 4.10. During a CAN persistent ridge, phases 7 and 8 are significantly more likely to occur, while phases 3–5 are significantly less likely to occur. In the five days following a CAN persistent
ridge, there are fewer statistically significant MJO phases. The frequency of phase 7 remains statistically higher than climatology and the frequency of phase 6 becomes statistically lower than climatology. The significantly higher frequency of phase 7 of the MJO before, during, and after a CAN persistent ridge is the most striking pattern of this particular analysis. In their idealized modeling of the extratropical response to the MJO, Lin and Brunet (2018) shows a persistent 500-hPa ridge over the CAN domain when their model is forced with the −MJO pattern (analogous to the RMM phase 7), corroborating our results here. In the period 5–10 days following a CAN persistent ridge, the frequency of phases 2 and 8 of the MJO are statistically higher than climatology, while the frequency of phase 6 remains statistically lower than climatology.

Fig. 4.12 shows the relationship between WUS persistent ridges and MJO frequency. During the period from 10 to 5 days prior to WUS persistent ridge onset, the frequency of MJO phase 2 is significantly lower than climatology and MJO phase 4 is significantly higher than climatology. This significant increase in MJO phase 4 agrees with results from Gibson et al. (2020) showing significant frequency increases starting 8 days before their analogous southern ridge events. As with Fig. 4.11, a coherent signal appears and propagates in time, as the frequency of phase 5 becomes significantly higher than climatology and the frequency of phase 2 remains significantly lower than climatology in the five days preceding a WUS persistent ridge. The coherent pattern continues during WUS persistent ridge events: phases 6–8 of the MJO have significantly higher frequency than climatology while phases 1 and 3 have significantly lower frequency than climatology. As mentioned for the PAC domain, MJO phases 6–8 often interact with the equatorward side of the NPJ and allow for Rossby waves to propagate from the tropics toward western North America, offering a dynamical understanding for the increased frequencies.
of these phases of the MJO during a WUS persistent ridge. In the two five day periods following
a WUS persistent ridge, this coherent pattern of propagating MJO frequency anomalies stops and
the significant frequency anomalies that are present are more scattered. For the five days
immediately following a WUS persistent ridge, phase 2 of the MJO has a significantly higher
frequency while phase 3 has a significantly lower frequency. 5–10 days following a WUS
persistent ridge, the frequency of MJO phase 7 becomes significantly higher than climatology
while the frequency of phase 4 becomes significantly lower than climatology.

Compared to Fig. 4.5–4.8 and the AO-ridge relationships, the MJO-ridge relationships
shown in Fig. 4.9–4.12 have more significant frequency differences for all domains. The number
of statistically significant MJO phases is greater for the period preceding CAN and WUS
persistent ridges than for the period preceding GOA or PAC persistent ridges, but still to a
greater extent than the AO. While this analysis is not predictive itself, it shows that there are
statistically significant relationships with the MJO at lead times approaching the S2S timescale.
After adding in the inherent predictability of a persistent ridge by definition persisting for at least
7 days, these findings could allow for some predictive skill of persistent ridges on the S2S
timescale.

4.2.3 Ridge Relationship with the North Pacific Jet Stream (NPJ)

The next set of figures show the frequency of different phases of the NPJ in the same
periods relative to persistent ridges in one of the four domains. NPJ phase is derived from
Winters et al. (2019), with four phases as well as a climatological origin to describe the
geographic location of the NPJ. Unlike the RMM index, the NPJ phase is not concerned with the
strength of the NPJ itself; analysis relating the strength of the NPJ to persistent ridges would be highly relevant but is not included in this thesis. Similar to the MJO, NPJ phase is calculated and available in Winters et al. (2019) starting in 1979, so roughly half of the ridge cases used in the composites had to be excluded for this analysis.

Fig. 4.13 shows the relationship between the phase of the NPJ and GOA persistent ridges. Compared to the relationship with GOA persistent ridges and the AO or MJO, there is more statistical significance over almost all time periods. For simplicity, descriptions in this section will be about the coherent propagating significance mentioned for the MJO. For the jet phases statistically more likely to occur, there are two broad signals of statistical significance. The first is a shift from a poleward-shifted NPJ starting 10–5 days prior to GOA persistent ridge onset, shifting through the origin in the five days prior to a GOA persistent ridge, to an equatorward-shifted NPJ during and up to five days following a GOA persistent ridge. The second, simpler pattern from this figure is a significantly increased likelihood for a retracted NPJ for the entire period before, during, and after a GOA persistent ridge except for the five days immediately following the end of a GOA persistent ridge. For the significant frequency decreases, they occur in most other phases of the NPJ to make almost all phases statistically significant. Every phase of the NPJ is statistically different from climatology during the first three periods (10 days before GOA persistent ridge onset through GOA persistent ridge dissipation) except for the poleward-shifted NPJ in the period 5 days before GOA persistent ridge onset. After GOA persistent ridge dissipation, the origin (climatological NPJ conditions) is significantly lower in frequency in both of the five days periods after the GOA persistent ridge.

Now looking at the PAC domain in Fig. 4.14, there is a coherent pattern of significant NPJ phase anomalies that propagate in time similar to the GOA analysis. The pattern is a shift
from significant frequency increases for extended NPJ conditions 10–5 days before a PAC persistent ridge starts, a significant increase in the frequency of climatological conditions in the five days prior to PAC persistent ridge formation that ends with a significant increase in the frequency of a retracted NPJ during a PAC persistent ridge. In addition to the enhanced occurrence of an extended NPJ 10–5 days before a PAC persistent ridge, there is also significantly enhanced occurrence of climatological conditions and poleward shift of the NPJ and significantly reduced occurrence of an equatorward-shifted or retracted NPJ. In the five days immediately before a PAC persistent ridge, only the origin has a significant increase in frequency while equatorward-shifted, extended, or retracted NPJ conditions have a significant decrease in frequency during this period. During a PAC persistent ridge, only the jet retraction phase of the NPJ has a significant increase in frequency, while the origin and jet extension phases have a significant decrease in frequency. In the five days after a PAC persistent ridge, the NPJ is statistically more likely to be in a poleward or equatorward shifted position, while statistically less likely to be near the origin or extended. Unlike the GOA composites (Fig. 4.13), the near equal frequency of poleward and equatorward NPJ shifts after a PAC persistent ridge demonstrates that multiple flow patterns are possible after the persistent ridge ends. By the last period (5–10 days after a PAC persistent ridge), only jet retraction has a significantly reduced frequency, with all other phases of the NPJ near their climatological frequencies.

Fig. 4.15 shows the relationship with NPJ frequency and CAN persistent ridges. While it does not propagate through the phase space, there are several persistent, significant frequency anomalies. In all periods except for the five days immediately before the start of a CAN persistent ridge, an equatorward-shifted NPJ is significantly more likely to occur. Likewise, in all periods except for the five days immediately following the end of a CAN persistent ridge, NPJ
retractions are significantly less likely to occur. A third, lesser pattern is a significant increase in the frequency of NPJ extensions from 10 days before CAN persistent ridge onset through ridge end. The climatological position of the NPJ has mixed significance, with significant decreases in frequency from 10–5 days before CAN persistent ridge onset and 0–5 days after CAN persistent ridge dissipation and significant increases in frequency 5 days prior to CAN persistent ridge onset and through the duration of the persistent ridge. The only remaining significant NPJ frequency is a significant decrease in the frequency of poleward-shifted NPJ conditions during a CAN persistent ridge. Unlike the previous two domains, there were very few patterns that propagated through the phase space in a coherent manner. Instead, significant NPJ phase anomalies are more likely to simply persist from before, during, and after a persistent ridge forms in the CAN domain. This also matches with findings from Fig. 4.7 and 4.11, where AO negative conditions and phase 7 of the MJO were significantly more likely to occur before, during, and after a CAN persistent ridge.

Fig. 4.16 shows the relationship with NPJ frequency and WUS persistent ridges. For the 10 days before WUS persistent ridge formation, there is a significant increase in the frequency of poleward-shifted NPJ conditions. There is also a significant increase in equatorward-shifted NPJ frequency from five days prior to WUS persistent ridge onset through the end of the WUS persistent ridge lifecycle. Similar to Fig. 4.14, there are two opposing large-scale flow patterns that are significantly more likely to occur immediately before a persistent ridge, highlighting that there can be competing pathways leading toward regime formation. The third phase of the NPJ with a significantly higher frequency during a WUS persistent ridge is the origin. In the 10 days after a WUS persistent ridge, only a retracted NPJ is significantly more likely to happen. Now addressing other NPJ phases, NPJ extensions are significantly less likely to happen in all periods
before and after a WUS persistent ridge. Equatorward-shifted NPJ conditions are significantly less likely to occur in the period 5–10 days before or 5–10 days after a WUS persistent ridge. Jet retractions are significantly less likely to occur from 10 days prior to WUS persistent ridge onset through to persistent ridge dissipation. The origin only has significant decrease in the five days following the end of a WUS persistent ridge, while poleward-shifted NPJ conditions are significantly less frequent during and for five days following the end of a WUS persistent ridge.

Of the three atmospheric indices used in this analysis, the NPJ—persistent ridge relationship has the most statistically significant frequency anomalies. This suggests that the NPJ could offer the most potential predictability of any of the indices discussed, bolstered by the already established synoptic-dynamic understanding of jet streams and persistent ridges/blocking patterns. Winters (2021) has found that the phase of the NPJ can be predicted in operational S2S models up to 3 weeks in advance. However, Winters (2021) also finds that the lowest-skill forecasts initialize or verify in a period of a high-latitude ridge, suggesting that the NPJ forecast skill related to persistent ridges would be less than 3 weeks. Further work could be done to test the lead-time with which there are statistically significant NPJ phase anomalies, and if those anomalies can translate into useable predictability for the forecast of persistent ridges or other flow anomalies.
Fig. 4.1: Odds ratio of atmospheric river occurrence (a) 10 days leading up to a persistent ridge regime, (b) between start and end dates of a persistent ridge regime, and (c) 10 days following a persistent ridge regime for the GOA domain.

Fig. 4.2: As in Fig. 4.1, but for the PAC domain.

Fig. 4.3: As in Fig. 4.1, but for the CAN domain.
Fig. 4.4: As in Fig. 4.1, but for the WUS domain.

Fig. 4.5: The percent frequency of the Arctic Oscillation (AO) for various stages preceding, during, and following the lifecycle of a persistent ridge regime in the GOA domain. Phases with stars indicate where that frequency of a particular phase is significantly higher (red star) or lower (blue star) at the 95% confidence interval based on bootstrap resampling.
Fig. 4.6: As in Fig. 4.5, but for the PAC domain.

Fig. 4.7: As in Fig. 4.5, but for the CAN domain.
Fig. 4.8: As in Fig. 4.5, but for the WUS domain.

Fig. 4.9: The percent frequency of the Madden-Julian Oscillation (MJO) for various stages preceding, during, and following the lifecycle of a persistent ridge regime in the GOA domain. Phases with stars indicate where that frequency of a particular phase is significantly higher (red star) or lower (blue star) at the 95% confidence interval based on bootstrap resampling.
Fig. 4.10: As in Fig. 4.9, but for the PAC domain.

Fig. 4.11: As in Fig. 4.9, but for the CAN domain.
Fig. 4.12: As in Fig. 4.9, but for the WUS domain.

Fig. 4.13: The percent frequency of the phase of the North Pacific jet stream for various stages preceding, during, and following the lifecycle of a persistent ridge regime in the GOA domain. Phases with stars indicate where that frequency of a particular phase is significantly higher (red star) or lower (blue star) at the 95% confidence interval based on bootstrap resampling.
Fig. 4.14: As in Fig. 4.13, but for the PAC domain.

Fig. 4.15: As in Fig. 4.13, but for the CAN domain.
Fig. 4.16: As in Fig. 4.13, but for the WUS domain.

5.1 Motivation for Case

As a final component of this thesis, an analysis of a particular persistent ridge case is presented here. The goal of this case study is to examine how the dynamical mechanisms and statistical relationships found for composite persistent ridges compare to those involved in an individual case, as well as to elucidate some of the synoptic variability averaged out through the compositing process. This particular case study from late December 1980 and January 1981 was selected because it happened to be the longest-lasting persistent ridge in both the CAN and WUS domains. Other cases (not included in the thesis) have mostly involved persistent ridges in the GOA and PAC domains, so this case will allow for a more complete understanding of persistent ridge evolution.

While not analyzed in exhaustive detail in this thesis, persistent ridge and blocking patterns are so widely studied because of the effect they can have on sensible weather both upstream and downstream of the block. Taubensee (1981) and Wagner (1981) both note the extreme temperature and precipitation anomalies that occurred during December 1980 and January 1981 when this ridge was quasi-stationary over western North America. During this period, Alaska and areas west of the Continental Divide experienced record warm temperatures during this event while the eastern two-thirds of the contiguous United States experienced record or near-record cold temperatures. According to Table 1 in Wagner (1981), Fairbanks, Alaska was 16.7°C above average for the month of January 1981, while much of Florida had its coldest
recorded January up to that point. As noted in Wagner (1981), there was a major pattern shift across eastern North America after the persistent ridge broke down in late January. During just the first 18 days of January 1981, Worcester, Massachusetts was on average 7.8°C below average. Along with the west-to-east dipole of temperature anomalies, there was also a reduction in precipitation across the entirety of North America. Much of eastern North America received less than 25mm of precipitation for the month, with many cities experiencing their driest January to date. Because of dry conditions earlier in the winter, much of the Ohio Valley and Northeast experienced severe drought by the end of January 1981, threatening water supplies for cities. These records show the temperature and precipitation anomalies that a persistent ridge can have downstream, and the importance in understanding these especially long-lived persistent ridges in more detail.

5.2 Large-Scale Conditions

Fig. 5.1 shows the 500-hPa geopotential height anomalies averaged over the duration of the persistent ridge over western North America (20 December 1980–27 January 1981). There is a remarkably well-defined hemispheric wavenumber-3 pattern in the height anomaly field, but each ridge and trough is unique in their configuration. The time-mean ridge over western North America is the most meridionally oriented but has perhaps the weakest positive height anomaly of the three ridges. The largest-scale feature is the broad trough encompassing nearly the entire North Pacific Ocean, which contains the lowest height anomalies of the Northern Hemisphere around 160°W. This North Pacific trough will be a key feature in the formation and evolution of the western North America persistent ridge. Further upstream, there is a zonally-oriented ridge
over Siberia centered around 90°E, another important feature in the synoptic evolution of the persistent ridge over western North America.

To further contextualize this weather regime, Fig. 5.2 shows the state of the NPJ in December 1980 and January 1981 based on Winters et al. (2019). In late December, the NPJ is initially equatorward shifted. As the persistent ridge over western North America begins to form around 22 December 1980, there is an abrupt shift toward a NPJ extension. Aside from a brief return to an equatorward-shifted NPJ in early January 1981, the NPJ remains in the extended phase for almost the entire duration of the persistent ridge lifecycle. This remarkable persistence of the NPJ phase highlights how stable the upstream flow pattern was during the persistent ridge across western North America. Toward the end of the persistent ridge lifecycle, the NPJ shifts from a jet extension toward a climatological jet position represented by the unit circle in this diagram. Having the NPJ in an anomalously perturbed state for around a month is itself remarkable and shows how rare it can be to sustain an anomalous jet stream for over a month. The overall NPJ evolution also shows how the atmosphere must be in a consistently anomalous state in order to maintain a persistent ridge for over a month. The shift of the NPJ toward a climatological position in late January 1981 coincides with the dissipation of the persistent ridge over western North America.

Fig. 5.3 shows the state of the MJO for the same time period of late December 1980 and January 1981 as defined by Wheeler and Hendon (1983). Given the previously mentioned literature on MJO—jet coupling on the S2S timescale, one may expect that the MJO was active and contributing to the anomalous extension and poleward shift of the NPJ. However, Fig. 5.3 shows that the MJO was inactive, represented by the unit circle, for most of the duration of the persistent ridge. While there may have been initial flow perturbations associated with the MJO
earlier in December, there is little evidence that tropical convection associated with the MJO aided in maintaining the anomalous NPJ state. An inactive MJO during the persistent ridge also suggests that rather than tropical variability, baroclinic eddies repeatedly reinforcing the upstream side of the ridge is the dominant mechanism by which the ridge persisted over western North America. The MJO emerged out of the unit circle around 15 January 1981 between phases 6 and 7, phases that both have statistically significant increases of frequency during a CAN and WUS persistent ridge as seen in Fig. 4.11 and 4.12. The active phase of the MJO propagates as the persistent ridge begins to dissipate in late January, ending in phase 8 before becoming inactive once again.

5.3 Synoptic Evolution of the Persistent Ridge Regime

5.3.1 Persistent Ridge Formation: 16–20 December 1980

The synoptic evolution of the western North America persistent ridge will be broken down into four periods, starting with the formation of the persistent ridge over the western United States. Fig. 5.4a shows a small negatively-tilted ridge along the coast of western North America on 16 December 1980. Upstream of the ridge, there is a shortwave trough around 40°N and 160°W with no associated surface cyclone, a second trough with a sub 980-hPa surface cyclone off the coast of the Kamchatka Peninsula, and a cutoff anticyclone over the Arctic Ocean near the Dateline on 16 December. From 16 to 18 December, the two troughs merge to become one larger negatively-tilted trough/surface cyclone while remaining in the same position. At the same time, the surface high over the Arctic Ocean builds by 16 hPa over 48 hours and elongates toward the southeast over Alaska by 18 December (Fig. 5.4c, h). At 500 hPa, this same process
appears as a strengthening block over the Arctic Ocean, evolving into a negatively-tilted ridge centered around 150°W. This ridge building from 16–18 December to the east of the Dateline is occurring in a region of southerly flow at 250 hPa and diabatic heating implied from mid-level ascent within an atmospheric river (Fig. 5.4f–h). In addition to the diabatic heating and southerly flow, concurrent low-level warm air advection implied from the surface pressure and 1000–500-hPa thickness contours also aids in developing the ridge around 170°W (Fig. 5.4a–c).

As ridge building occurs, the negatively-tilted trough in the western North Pacific basin begins to elongate further to the east associated with three separate local 500-hPa geopotential height minima from 140°E to 150°W by 19 December (Fig. 5.4i). These three height minima help to establish and elongate the basin-wide trough across the North Pacific Ocean in Fig. 5.1, as well as help to undercut the building ridge extending from western North America toward the Bering Strait (Fig. 5.4i). By 20 December, the high-latitude portion of the extended western North America ridge is disconnected from the flow pattern as the trough further south continues to elongate toward the western coast of North America. As a result, what was the high latitude portion of a single ridge is now a disconnected blocking anticyclone centered around the Bering Strait. At the same time, the southern half of this ridge centered over western North America begins to rebuild after the northern half of the ridge fractured and becomes the blocking high now over the Bering Strait (Fig. 5.4j). On 20 December, the ridge over western North America is now persistent within the WUS domain, beginning the lifecycle of the persistent ridge.

5.3.2 Persistent Ridge Building: 24–28 December 1980

Once the persistent ridge was established in the WUS domain on 20 December, it took almost a week for the ridge to amplify and become persistent within the CAN domain. The amplification of the persistent ridge required further linked upstream flow perturbations in order
to occur. As mentioned in the previous section, there is a zonally-oriented blocking high across the Bering Strait on 24 December (Fig. 5.5f). Because of this blocking high, cyclones forming in the western North Pacific Ocean were redirected further east toward North America along the northern edge of the anomalous NPJ rather than toward the northeast through the Bering Strait. Fig. 5.5a shows a 972 hPa cyclone in the Gulf of Alaska immediately upstream of the persistent ridge, along with an atmospheric river (AR) extending from just north of Hawaii toward the western coast of North America. By 25 December, the AR was making landfall along the west coast of North America on the northwestern flank of the persistent ridge, a common feature seen in both the composite maps (chapter 3) and the AR odds ratio calculations (chapter 4).

Associated with this AR was strong southerly flow at 250 hPa and embedded upward vertical motion from 400–600 hPa (Fig. 5.5g), as well as implied low-level warm air advection (Fig. 5.5b).

The aforementioned landfalling AR persisted along the coast of Oregon, Washington, and southern British Columbia for several days from 25–27 December (Fig. 5.5g–i). As this occurred, the blocking high over the Bering Strait began to retrogress and move toward eastern Siberia (Fig. 5.5i). This shift in the blocking high meant that the cyclone associated with the landfalling AR could start to move toward the northeast. Because of these complex flow interactions, the warm air advection, upward vertical motion within the AR, and strong southerly flow aloft immediately upstream of the persistent ridge persisted for several days. All three of these processes are associated with quasi-geostrophic geopotential height rises, and the persistent ridge amplified and extended into the CAN domain by 27 December as a result (Fig. 5.5b–d, g–i). With the broad trough across the North Pacific Ocean and the persistent ridge now in both the
WUS and CAN domains, the large-scale flow pattern on 28 December is more-or-less locked in for the duration of the persistent ridge lifecycle (Fig. 5.5e, j).

5.3.3 Persistent Ridge Maintenance: 4–8 January 1981

The following section will focus on a representative period of ridge maintenance during the persistent ridge lifecycle. From 28 December 1980 to 4 January 1981, the persistent ridge over western North America had taken on a negative tilt in the northern portion as the ridge amplified (Fig. 5.6f). The negative tilt of the ridge over western North America, like the antecedent blocking pattern mentioned in section 5.3.2, effectively cuts off any cyclones from entering the Arctic Ocean through the Bering Strait. This means any cyclones that form in the western North Pacific basin would travel across the North Pacific Ocean and recurve toward the north on the upstream side of the persistent ridge.

On 4 January, there were two cyclones around 170°E and 175°W, both with central pressures below 988 hPa (Fig. 5.6a). These cyclones merged to become one large cyclone centered along the Dateline by 5 January. A trough associated with the easternmost cyclone becomes separated from the flow pattern as the two previously mentioned cyclones merge together (Fig. 5.6b, g). This separated trough slowly dissipates over a few days as the merged cyclone dominates the flow pattern over the Central Pacific. Before this trough fully dissipated, there was a region of southerly flow, upward vertical motion, and a local IVT maximum immediately upstream of the ridge axis associated with the eastern edge of the trough. Through latent heating and warm air advection, this southerly flow helped to reinforce the persistent ridge over western North America (Fig. 5.6g, h). By 7 January, the cyclone near the Dateline deepened to below 972 hPa and had become more zonally elongated near the Aleutian Islands (Fig. 5.6d).
On the southern flank of this cyclone, the NPJ became extended to near 150°W on 7 January, reflected in a shift in Fig. 5.2. Along with the jet extension, another surge of southerly flow and corresponding AR associated with the warm sector of the cyclone in the Central Pacific approached the western flank of the persistent ridge (Fig. 5.6h, i). Once again, the same processes of warm air advection and diabatic heating contributed to the periodic reintensification of the persistent ridge over western North America. However, one key feature of this ridge intensification process is that the ridge axis on 6 January was further east than it was on 8 January after the period of ridge rebuilding (Fig. 5.6 h, j). This periodic reinforcement of the persistent ridge on synoptic timescales resulted in the ridge effectively moving westward in short, discrete intervals. This process is best described by the phrase “discontinuous ridge retrogression” or DRR, and can help to explain how the persistent ridge is able to stay quasi-stationary for an extended period of time.

5.3.4 Persistent ridge dissipation: 17–23 January 1981

This final section will illustrate the dissipation of the persistent ridge over western North America and the return to a more progressive flow pattern. On 17 January, Fig. 5.7a shows the NPJ was anomalously extended across the North Pacific basin, with the core of the jet stream exceeding 100 m s\(^{-1}\). In addition, there was a zonally-elongated ridge centered around 160°E and 35°N, seen in both the 250-hPa winds and 500-hPa geopotential heights (Fig. 5.7a, h). The development of this new ridge coincided with the emergence of the MJO from the unit circle into phases 6 and 7. The diabatic heating and convective outflow aloft associated with the MJO occurred just south of the zonally-elongated ridge, aiding in building this ridge and perturbing the basin-wide trough that has persisted since late December 1980. This initial perturbation to the
regional flow pattern was the first contributor to the eventual breakdown of the persistent ridge over western North America.

As this low-latitude ridge developed in association with the MJO-related convective outflow, the western North America ridge began to change in its characteristics. On 19 January, a downstream trough over North America near 110°W developed concurrently with a period of DRR and resulted in a narrowing of the western North America persistent ridge (Fig. 5.7d, k). This period of DRR was also linked to rapid cyclogenesis upstream of the ridge axis on 19 and 20 January as a shortwave trough with an embedded AR interacted with the poleward jet exit region of the NPJ (Fig. 5.7c,d,j,k).

On 21 January, the zonally-elongated ridge amplified further, with the ridge crest now around 170°W and 40°N along the western edge of this now mature cyclone. Through the amplification of this lower-latitude ridge, the cyclone and associated upper-level trough became secluded from the flow pattern by 21 January, caught between the persistent ridge over western North America and the lower-latitude ridge in the Central Pacific Ocean (Fig. 5.7e, l). As this happened, the ridge over western North America began to fold over toward the east, indicative of anticyclonic wave breaking (Fig. 5.7l). On 22 January, a second cyclone developed along the Dateline before the previous cyclone in the eastern North Pacific could travel north toward the Arctic Ocean. This second cyclone deepened and moved northeast toward the Aleutian Islands, while the first cyclone immediately upstream of the persistent ridge over western North America dissipated in the eastern North Pacific basin. Because the first cyclone was still immediately upstream of the ridge axis, the second cyclone could not interact with the upstream side of the persistent ridge over western North America and help to rebuild the ridge. Instead, more shortwave troughs and ridges formed instead of the previously dominant longer wavelength
troughs and ridges. In effect, this means the persistent ridge regime began to dissipate and the overall flow pattern became progressive.
Figure 5.1: Time-mean 500-hPa geopotential height (contours, dam) and geopotential height anomalies (shaded, m) averaged over the duration of the persistent ridge lifetime (20 December 1980–27 January 1981).
Figure 5.2: Position of the North Pacific jet stream, quantified by the phase diagram introduced in Winters et al. (2019) for late December 1980 and January 1981.
Figure 5.3: Amplitude and phase of the MJO using the RMM phase for late December 1980 and January 1981.
Figure 5.4: Mean sea level pressure (contour, hPa), 1000–500-hPa thickness (red/blue dashed contour, dam), and 250-hPa wind speed (shaded, m s⁻¹) at 0000 UTC for (a) 16 December 1980, (b) 17 December 1980, (c) 18 December 1980, (d) 19 December 1980, and (e) 20 December 1980. 500-hPa geopotential height (contour, gpm), integrated vapor transport (IVT, shaded, kg m⁻¹ s⁻¹), 250-hPa v-wind (red/blue dashed contour, m s⁻¹), 400–600-hPa averaged ascent (green contour, 5x10⁻³ hPa s⁻¹) at 0000 UTC for (f) 16 December

Figure 5.5: As in Fig. 5.3 at 0000 UTC for (a), (f) 24 December 1980, (b), (g) 25 December 1980, (c), (h) 26 December 1980, (d), (i) 27 December 1980, and (e), (j) 28 December 1980.
Figure 5.6: As in Fig. 5.3 at 0000 UTC for (a), (f) 4 January 1981, (b), (g) 5 January 1981, (c), (h) 6 January 1981, (d), (i) 7 January 1981, and (e), (j) 8 January 1981.
Figure 5.7: As in Fig. 5.3 at 0000 UTC for (a), (h) 17 January 1981, (b), (i) 18 January 1981, (c), (j) 19 January 1981, (d), (k) 20 January 1981, (e), (l) 21 January 1981, (f), (m) 22 January 1981, and (g), (n) 23 January 1981.
6. Discussion, Conclusions, and Suggestions for Future Work

6.1 Discussion and Conclusions

The overarching goal of this thesis was to further understand the mechanisms behind the formation, maintenance, and dissipation of persistent ridges over the eastern North Pacific Ocean (NPAC) and western North America. By using an updated version of the persistent anomaly criteria defined by Dole and Gordon (1983), and one nearly identical to Miller et al. (2020), a 70-year climatology of persistent anomalies was constructed. From the identified persistent anomalies, ridges were defined by continuous persistent anomalies within specific domains over the eastern NPAC and western North America. With the updated climatology created and verified against previous work, persistent ridges were further studied through composite analyses, statistical relationships to atmospheric indices, and case study analysis.

6.1.1 Climatology and Composite Analysis of Persistent Ridges

6.1.1.1 Climatology of Events

When compared to the results from Dole and Gordon (1983), the modified criteria used in this study based on standardized 500-hPa geopotential height anomalies generated similar results. These results were also qualitatively similar to Miller et al. (2020), although their threshold of two standard deviations reduced the number of cases in their sample from our calculations. Our climatology of persistent anomalies also benefits from the length of record in the NCEP-NCAR reanalysis dataset, allowing us to create a 70-year climatology. When breaking up this climatology into seven 10-year subsets, decadal variability in the number of persistent
anomalies is apparent. Further work can and should be performed to investigate the role of shifting “hotspots” for persistent anomalies across the Northern Hemisphere.

Persistent anomalies were aggregated within pre-defined domains (Fig. 2.2) into persistent ridges, allowing for discrete start and end dates of the persistent ridges. Figure 2.3 shows the CDF of the duration of persistent ridges for each of these four domains. While most persistent ridges persist for 7–10 days, a small subset of persistent ridges within each domain can persist for up to 45 days. Among the four domains, there is a strong longitudinal variation in the slope of the CDF curve. The western domains centered over the NPAC (GOA and PAC) have a shallower slope and thus more cases that persist for between 14–21 days. The eastern domains centered mostly over western North America (CAN and WUS) have a steeper slope and very few cases between the 7–10-day persistent ridge and the 30+-day persistent ridge. Results from chapters 3–5 suggest through multiple lines of evidence that the inability of the North Pacific jet stream (NPJ) to remain extended during CAN and WUS ridge events limits the number of longer-duration persistent ridges when compared to the GOA and PAC domains.

6.1.1.2 Composite of Events

6.1.1.2.1 Gulf of Alaska (GOA) Composites

Many of the results from the composite analysis of the persistent ridge events agree with previous studies. The GOA composite was considered the “baseline”, as it highlighted many of the dynamical and thermodynamical processes found to be important for persistent ridge formation and maintenance. The overall GOA persistent ridge lifecycle featured a slowly weakening cyclone on the western edge of the persistent ridge, a northward extending AR along
the western edge of the persistent ridge, a diffluent upper-level jet exit region to the southwest of
the persistent ridge, and a region of baroclinicity upstream along the NPJ that weakened
throughout the ridge lifecycle. The presence of cyclones and ARs upstream of a blocking pattern
all agree with previous studies about the role of synoptic-scale cyclones in forming and
The role of cyclones is further corroborated by the reduction in EGR throughout the ridge
lifecycle, indicating that a reduction in baroclinic instability in the western NPAC is a key
component in the dissipation of persistent ridges.

Composites of the 80th percentile cases generally match with the full composite for each
domain, but some key differences offer insight into slightly different dynamical processes. The
GOA 80th percentile composites show two ARs upstream of the persistent ridge separated by
around 30° of longitude. A series of ARs like this indicates several cyclones in succession,
iluminating relevant dynamical features obscured in the full composite. The longest-lasting
GOA persistent ridges had stronger low-level MFC along the northern and western edges of the
persistent ridge, indicating a statistically significantly stronger diabatic contribution to ridge
formation and maintenance. The 80th percentile GOA composites also suggested that cutoff low
formation in the central NPAC could be a common occurrence for longer-lasting persistent
ridges. Whether or not this is another example of relevant dynamics observed with a smaller
number of composite members or a distinct dynamical difference of the longer-lasting persistent
ridges is unclear and could be an avenue for future work.

6.1.1.2.2 Pacific (PAC) Composites
Of the four domains, the PAC composites were the most difficult to assess and offer a fair amount of future work to further understand them. In general, the full PAC composites had weaker upstream cyclones, weaker IVT, and weaker upper-level diffluence within the jet exit region. Because of the weaker upstream cyclone and jet exit region, there is the least amount of evidence for baroclinic instability aiding in the formation and maintenance of persistent ridges in the PAC domain. An additional layer of complexity is introduced from the 80th percentile composite of the PAC persistent ridges, where the overall lifecycle of the persistent ridges more closely matches the GOA composite than the PAC composite. Within the 80th percentile composite, the EGR reduces more upstream of the persistent ridge and the composite persistent ridge is closer in proximity to the upstream cyclone. It is not an exact analog to the GOA composite, since the composite upstream cyclone diminishes much quicker in the PAC 80th percentile composites than in the GOA composite, but the differences between the full and 80th percentile PAC composites suggests that dynamically-different ridge types are being combined and averaged together within the PAC domain. This would require considerable additional work to parse out how best to categorize ridge types by dynamical mechanisms rather than combining all into one set of composites.

6.1.1.2.3 Canada (CAN) Composites

The CAN domain composites have the deepest and broadest upstream cyclone of all the composites, further reiterating the important role of upstream cyclones in helping to form and maintain persistent ridges. The upstream cyclone is very zonally extended, again a potential artifact of aliasing. However, the case study analysis in chapter 5 of a CAN persistent ridge case
showed similar large-scale “gyres” in the central NPAC with multiple embedded cyclones. The case selected for chapter 5 is by definition an outlier as the longest-lasting case within two domains, but it still appears to be a unique feature of composite CAN persistent ridges. Within the 80th percentile composites, multiple training cyclones are visible in the MSLP and IVT fields. As stated for the GOA composite, this difference could just be a result of fewer composite members rather than a dynamical difference of these longer events.

Another unique characteristic of the CAN composites is the lack of retrogression of the ridge, as 700-hPa anticyclonic vorticity remains closely aligned with the Canadian Rockies throughout its lifecycle. Having mountains to favor a ridge through stationary Rossby waves is another difference between the western and eastern domains in this study, potentially explaining the distribution of persistent ridge duration in Fig. 2.2. The final dynamical difference for composite CAN persistent ridges is that the EGR does not diminish throughout the ridge lifecycle. The eastern extent of the EGR maximum retracts as the composite ridge approaches the end of its lifetime, but the maximum values do not diminish. In this composite, baroclinic instability does not diminish with time, suggesting that a lack of baroclinicity is not a limiting factor in the lifecycle of CAN persistent ridges. This is further corroborated by the 80th percentile composites, as there is a remarkable retraction of the jet stream and eastern extent of the EGR maximum but a strengthening of the magnitude of EGR in the western NPAC by the end of the CAN persistent ridge lifecycle. As discussed before, it appears that inability to maintain an extended jet stream for long periods is the biggest limiting factor in the lifetime of CAN composite ridges, not the absence of baroclinic instability.

6.1.1.2.4 Western United States (WUS) Composites
The final domain for the composite analysis is for the Western United States. The most obvious difference in this domain is the higher magnitude of IVT across the western and central NPAC, especially through 60% of the ridge lifecycle. Higher IVT upstream of the developing and building persistent ridge in the WUS domain would suggest that latent heating may play a more important role in the dynamical evolution of WUS persistent ridges. In fact, low-level averaged MFC is always significantly higher for the longest-lasting WUS persistent ridges than the full climatology in the ridge evolution. This differs from other long-lasting ridges, where the coherent structure of the MFC significance dissipates around 60% of the way through the ridge lifecycle. For the 80th percentile WUS composites, MFC is statistically significantly higher to the north and west of the ridge axis even during ridge dissipation, meaning the absence of statistically significantly higher MFC is not always the leading factor in ridge dissipation.

Another feature upstream across the NPAC is the reduction in EGR quite dramatically at the end of the persistent ridge lifecycle. Having a reduced amount of baroclinic instability upstream of the persistent ridge again suggests that a lack of cyclones helping to reinforce the ridge is a key component in the dissipation of most persistent ridges. This leaves the CAN composites as the outlier amongst the other domains, where there is still sufficient baroclinic instability for further cyclones and DRR at the end of a ridge lifecycle. This leads to the general conclusion that in general, baroclinic instability and upstream cyclones are the necessary dynamical feature for the formation and maintenance of persistent ridges over the eastern NPAC and western North America. In the eastern domains of WUS and CAN, this can partially be limited by the ability of the NPJ to remain extended for sufficiently long to maintain a persistent ridge for over 10 days as can happen more easily for the GOA and PAC domains.
6.1.2 Statistical Relationship of Persistent Ridges to Atmospheric Phenomena

6.1.2.1 Ridge Relationships with Atmospheric Rivers

The following section examined statistical relationships between persistent ridges and various atmospheric indices. The first set of analysis was for persistent ridges and the odds ratio of landfalling ARs along the western coast of North America. In general, there is a slight reduction in AR frequency in the 10 days prior to the start of a persistent ridge for all four domains. All things being equal, one might expect odds ratio values near one for this period, so the slight reduction means that even before a persistent ridge forms, the overall flow pattern is less conducive for landfalling ARs across nearly all of western North America. During a persistent ridge in any domain, there is a strong reduction of landfalling ARs immediately downstream of the ridge axis. In the 10 days after a persistent ridge, there is still a residual reduction in ARs across all of western North America, indicating that flow patterns which follow persistent ridges are still not conducive for AR landfall over western North America.

For the GOA analysis, there is a unique feature of a slight increase in the AR landfall odds ratio during and after a GOA persistent ridge across southern California and northern Mexico. This indicates the possibility of subtropical jet streak formation downstream of the persistent ridge, a feature also visible in the composite analysis of chapter 3. The slight increase in AR landfall likelihood reinforces the need for synoptic understanding of persistent ridges (e.g. size, location, and shape) to properly account for downstream characteristics. In the 10 days following a PAC persistent ridge, there is the greatest reduction in AR landfall across western North America of all four domains. Because the PAC composite analysis was the least clear of
the four domains, it is unknown what dynamical features may be leading to this composite having the greatest reduction in AR landfall in the period after the end of the persistent ridge. The final unique characteristic from this analysis is the similarity between the AR odds ratio from after a WUS persistent ridge and during a GOA persistent ridge. It was speculated that rather than dissipating, persistent ridges from the WUS domain may retrogress and continue being persistent within the GOA or PAC domains. While feature tracking of these ridges would be needed to quantify this, the similarity of the impacts of these periods on AR landfall likelihood suggests a linkage that requires further examination.

6.1.2.2 Ridge Relationships with Atmospheric Indices

6.1.2.2.1 Ridge Relationships with the Arctic Oscillation (AO)

The next set of analyses will involve the relationship between various atmospheric indices and persistent ridges. The first index will be the Arctic Oscillation, a widely used index as a proxy for hemispheric flow patterns and defined by Thompson and Wallace (1998). In general, statistically-significant anomalies in the frequency of phases of the AO emerge only five days prior to the start of a persistent ridge for most domains. This limits the effectiveness of the AO as a proxy for flow patterns related to the formation of persistent ridges. During a persistent ridge, three out of four domains have a statistically higher occurrence of positive AO conditions which makes sense when ridges (or higher surface pressure) are present over the midlatitudes. The one outlier to these general trends is for the CAN domain, which has a consistent statistical significance pattern from 10 days prior to CAN persistent ridge start to 10 days following CAN persistent ridge end. Because of the EOF used for calculating the AO, the upstream cyclone seen
in the CAN composites projects directly onto the AO itself. Knowing this, the consistently negative values of the AO before, during, and after a CAN persistent ridge suggests that the composite cyclone exists upstream of the CAN domain well before the start of a persistent ridge.

6.1.2.2.2 Ridge Relationships with the Madden–Julian Oscillation (MJO)

As with the AO, the phase of the MJO was assessed for periods before, during, and after a persistent ridge in each domain. In general, the MJO had statistically-significant frequency anomalies prior to persistent ridge onset for the CAN and WUS domains but little to no significance prior to persistent ridges in the GOA or PAC domains. Previous studies (Henderson et al. 2016, Attard and Lang 2019) had differing results to this study for MJO frequency anomalies at differing lag times before ridge or block onset, suggesting the relationship to the MJO is highly sensitive to the blocking definition used. However, results of MJO frequency during a block/persistent ridge do generally agree between this study and those papers. Across all four domains, phase 6 of the MJO is significantly more frequent in occurrence during a persistent ridge, suggesting the connection between the MJO and persistent ridges may be through an intermediate source (the NPJ) rather than a direct teleconnection for each domain. Of the four domains, the WUS has the clearest signal of a propagating MJO pattern that travels from phase 4 10 days prior to persistent ridge onset to phase 7 10 days after persistent ridge dissipation. For the CAN domain, MJO phase 6–8 are in some combination always statistically significantly higher than climatology for all five time periods assessed. While this may be due to aliasing, the consistency of the AO frequency for the CAN domain suggests that stalled or slowly-propagating MJO may be common before, during, and after a CAN persistent ridge.
The final statistical relationship assessed in this thesis is between persistent ridges and the phase of the NPJ. Of the three indices used in this thesis, the NPJ has the most statistically-significant frequency anomalies across all four domains. Not only is there statistically-significant frequency anomalies for certain phases of the NPJ, but these anomalies agree with the independent assessment of a synoptic-dynamic meteorologist. In general, there are different characteristics between the northern (GOA and CAN) and southern (PAC and WUS) domains for their patterns of statistically-significant NPJ frequencies. GOA and CAN both have at least one phase of the NPJ that has a statistically-significant increase in frequency from 10 days prior to persistent ridge formation through the end of the persistent ridge lifecycle. Persistent ridges within the PAC and WUS domains lack the statistically-significant NPJ frequency increase for the period before and during the persistent ridge, instead having changing phases with statistically-significant increases in frequency. Having a consistently statistically significant increase in the phase of the NPJ may mean that there is enhanced predictability of persistent ridges in the GOA or CAN domains when compared to the PAC and WUS domains. However, each shift of the NPJ to different phases is associated with different levels of predictability as shown in Winters et al. (2019) and Winters et al. (2021). Overall, the NPJ offers the most potential predictability of persistent ridges due to common statistically-significant frequency anomalies up to 10 days prior to persistent ridge formation. However, much more work would be needed to translate these frequency anomalies into actionable predictability through an increased
understanding of the physical and dynamical mechanisms behind shifts between and persistence within certain phases of the NPJ.

6.1.3 Case Study: Longest-Lasting Persistent Ridge over Western North America, December 1980–January 1981

The final component of this thesis is a case study analysis of the longest-lasting persistent ridge in both the CAN and WUS domains, occurring in late December 1980 through late January 1981. This particular case had widespread temperature and precipitation impacts across the US throughout late December and all of January. In general, western CONUS and Alaska had record warmth, eastern CONUS had record cold, and all of CONUS had near-record dry conditions for January 1981. In addition to surface weather impacts across the US, the persistent ridge over western NA was linked to a wavenumber-3 pattern across the entire Northern Hemisphere. Most importantly, a zonally-elongated trough across the entire Pacific basin and a blocking high over eastern Siberia helped to create the flow conditions necessary for the persistent ridge formation and maintenance over western NA. Further context for this case was added from assessing the phases of the MJO and NPJ during this persistent ridge. The MJO was relatively inactive until around 15 January 1981 when it emerged between phases 6 and 7, meaning the MJO could not explain the formation and maintenance of this persistent ridge. The NPJ, however, was equatorward shifted or extended for most of the duration of the persistent ridge, yet again demonstrating the close relationship between the NPJ and persistent ridges in this region.

The case study was subdivided into four periods representing key periods in the persistent ridge lifecycle: formation over the western United States, building into western Canada, a
representative period of ridge maintenance, and dissipation of the persistent ridge. In the four days leading up to persistent ridge formation (16–20 December), there was a negatively-tilted ridge extending from western North America up toward Alaska on 16 December 1980. Along with the amplified ridge, three separate cyclones in the NPAC began to form and track along the northern edge of the jet stream. These cyclones contributed to the amplification of the ridge near Alaska that resulted in a fracturing of the northern extension of the ridge along the Bering Strait by 18 December. This resulted in a zonally-elongated blocking pattern across the Bering Strait, preventing any cyclones over the NPAC from entering the Arctic Ocean. This set up the beginning of the time-mean trough over the NPAC and shifted the storm track all the way to the eastern NPAC. By 20 December, the remnants of the previous ridge over western NA began to amplify once again and become persistent in the WUS domain.

After the initial persistence of the ridge in the WUS domain, it took an additional week for the ridge to further amplify in western Canada. Around 25 December, a strong AR was making landfall from northern California to southern British Columbia that was associated with a 972 hPa parent cyclone in the Gulf of Alaska. This parent cyclone was situated to the northwest of the persistent ridge and could not move much with the configuration of the large-scale flow pattern in the NPAC. By 27 December, the blocking high over the Bering Strait began to retrogress toward eastern Siberia and allowed for the cyclone in the Gulf of Alaska to move toward the northeast. The closer proximity of the cyclone to the persistent ridge, along with the prolonged period of warm air advection, latent heating around the AR, and fluxes of anticyclonic vorticity allowed for the persistent ridge to amplify and become persistent over western Canada on 28 December.
For much of the lifecycle of this persistent ridge case, similar dynamics were occurring that allowed for the maintenance of the general structure of the ridge. By early January 1981, the earlier blocking high was now squarely over Siberia, but the NA persistent ridge had amplified toward the northwest, still effectively cutting off the NPAC storm track from entering the Arctic Ocean. This allowed many additional cyclones to form in the western flank of the NPAC jet along the coast of Asia, develop and track toward the western edge of the NA persistent ridge. In addition, cold air advection along the eastern edge of the Siberian blocking high maintained the baroclinic instability across the NPAC, allowing for the recurrent cyclones on the north side of the jet stream.

Finally, there was a long sequence of events relating to the dissipation of the persistent ridge in late January 1981. On 17 January, a low-latitude ridge developed on the south side of the NPJ around 160°E around the same time the MJO became active in phases 6 and 7, likely contributing to that low-latitude ridge forming. The low-latitude ridge broke up the stable, anomalous trough that dominated for most of the ridge lifecycle, setting the stage for further perturbations to the overall flow pattern. By 19 January, the persistent ridge over NA had become very narrow and meridionally extended. Eventually, the NA persistent ridge folded, indicative of anticyclonic wave breaking. As this occurred, the low-latitude ridge amplified further as two cyclones developed concurrently in the eastern NPAC around 20 January. The westernmost cyclone amplified the low-latitude ridge, essentially cutting off the eastern cyclone and its associated trough. With the trough cutoff, the western cyclone could not reinforce the NA persistent ridge, resulting in the end of the persistent ridge lifecycle by 22 January 1981.

6.2 Suggestions for Future Work
While this thesis covers multiple aspects of understanding persistent ridges, there is still additional work that would further improve our understanding. Increasingly, there is a debate about the nature of extreme weather and flow patterns with ongoing anthropogenic climate change. To be able to verify climate model projections of changes in the coming decades, identifying trends, and if they are robust, in persistent ridge frequency, duration, or location in reanalysis datasets would be needed. Because 500-hPa geopotential height has not risen uniformly across the Northern Hemisphere, the data would first need to be detrended to get a more accurate trend in persistent ridges during the wintertime. Another limitation of this study is the coarseness of the reanalysis data used as a compromise for the longer data period. Additional calculations could be performed using a higher resolution reanalysis such as ERA5, a product that now extends back to 1959 in an experimental form.

One major caveat within this study is the directionality of relationships between persistent ridges and atmospheric teleconnections. In chapter 4, all relationships are relative to periods before, during, or after a persistent ridge is present within one of the four domains. However, this does not translate into predictive skill of determining when a persistent ridge is more or less likely to form. What is needed is a calculation of the likelihood of a persistent ridge forming 10 days from now given a certain phase of an atmospheric teleconnection. For example, it was found that 10 days prior to a GOA persistent ridge, poleward-shifted and retracted NPJ phases have a statistically significantly increase in frequency. Knowing this ridge-NPJ relationship, a calculation of the likelihood of a GOA persistent ridge could be calculated from all instances of a retracted or poleward-shifted NPJ to assess if the relationship between the NPJ and persistent ridges is predictive or just associative. If there was, hypothetically, a statistically-
significant increased likelihood of a GOA persistent ridge 10 days following a poleward-shifted
NPJ, then monitoring the phase of the NPJ could be another predictor for large-scale flow
patterns on the S2S timescale.
REFERENCES


