CLIMATOLOGY OF SYNOPTIC SCALE ASCENT FOR WESTERN NORTH AMERICA: A perspective on storm tracks

by

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ABSTRACT

This study documents the climatological distribution and interannual variability in synoptic scale ascent for western North America. An alternative balance (AB) form of the omega equation is employed to diagnose ascent associated with synoptic disturbances in such a way as to exclude direct contributions from mesoscale motions such as orographic lift. Data from the European Center for Medium Range Forecasting (ECMWF) ERA-interim reanalysis is used with statistics presented for the 19 cool seasons (Oct-Apr) spanning 1989-2008.

Results indicate a sinusoidal primary ascent pathway across the eastern Pacific and Western North America. A secondary more zonal pathway is found across southern British Columbia and the high plains of Canada. The seasonal cycle of ascent is found to exhibit coherent temporal and regional patterns in maximized upward motions. Interannual variations in the spatial patterns of ascent are examined by means of Principal Component Analysis (PCA). The results show variations in the structure of the mean ascent pathway are closely related (r=-.82) with variations in the phase of the El Nino Southern Oscillation (ENSO). Secondary investigations explore the relationship between synoptic scale ascent and precipitation, a link which is found to be modulated by available moisture, latitude, and topography.

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CHAPTER 1

introduction

Background

The location, intensity, and frequency of transient synoptic disturbances have far reaching implications. On a global scale, such storms represent an important component of the climate system, transporting heat and moisture poleward, regulating the degree of cloudiness, and impacting the global hydrologic cycle. Regionally the role of these storms is no less important. For the mountainous regions of western North America, cool season storms provide significant contributions to mountain snowpack, which in turn provides critical water resources for the subsequent summer months (Serreze 1999). Significant shifts in the seasonal timing, location, or intensity of storms related to a warming global climate, as has been projected in recent studies (Bengtsson and Hodges 2006 ), may have significant hydrologic consequences throughout the region. Less obvious implications include variations in the duration and strength of persistent cold pool events throughout the intermountain west, which are common features in the quiescent interludes between storms and have significant impacts on regional air quality. Frequency and intensity of storms has further impacts on the burgeoning wind energy industry, which is vulnerable to climatic shifts not only in the mean intensity of wind but also in synoptic-scale (day to day) variability.

Despite the far reaching impacts of synoptic disturbances and numerous studies of storm tracks over the past half century, no clear picture has emerged for the structure and variation in synoptic-scale pathways across western North America (Figure 1.1) relative to the generally well understood storm tracks across the Pacific Ocean. This lack of clarity is due, in part, to the influence of terrain on the myriad techniques used to define storm tracks combined with the complexity of terrain-flow interactions on all scales of motion. For example, near surface storm metrics, such as cyclone tracking procedures, tend to show discontinuous storm tracks across interior western North America (Jeglum and Steenburgh 2010) while examination of mid and upper-tropospheric relative vorticity features exhibit more continuous tracks (Hoskins and Hodges 2002,). Even amongst studies of mid and upper-tropospheric fields, the latitude of the mean synoptic pathway varies over a range from 30N to 55N (Blackmon 1976, Hoskins and Hodges 2002)and the spatial patterns vary from zonal (Blackmon 1976) to amplified sinusoidal pathways (Hakim 2000). Furthermore, few studies have explicitly examined interannual variations in regional storm tracks nor the links between these variations and known patterns of planetary-scale variations such as the El Nino Southern Oscillation (ENSO) phenomenon.

Ambiguity in storm track location and variability notwithstanding, storm track studies rarely explore the synoptic implications of their particular metric. In the case of either a closed cyclonic circulation or an open wave trough, it is regularly observed that regions of precipitation and cloudiness are displaced from the kinematic center of the system, which is often the tracked quantity. Furthermore tracking studies tend to assign a storm to a single coordinate location, instead of allowing for the considerable areal coverage of storm impacts. The association of sensible weather with variance based methods is even less clear, where the requisite time scale and sample size for statistical averaging provide no insight into individual synoptic events. Yet, modern case studies of individual synoptic events tend to focus on the dynamics contributing to the impacts of a given system. Specifically, the analysis of mid-latitude weather systems frequently includes a diagnosis of synoptic scale ascent (Hoskins et al 1978, Hoskin and Pedder 1980, Keyser et al 1988, Martin 2007, West and Steenburgh 2010), which has profound impacts on precipitation, cloudiness, and the structure of the storm itself (Durran and Snellman , Martin). Despite its overt importance, few climatological studies have adopted ascent based metrics to address the location and magnitude of storms.

In light of the potential shifts in storm tracks due to anthropogenic climate change, the ambiguities in our present understanding of the distribution of storms, and a lack of consensus in storm tracking methodology, the purpose of the present study is to provide a clearer picture of the climatological mean storm tracks across western North America and the variability of those storm tracks seasonally and year-to-year. A novel ascent based approach, which is both dynamically constrained and filtered, is employed to diagnose the location and intensity of storms. A measure of storm track is obtained by accumulating, at 6-hour intervals, the vertically integrated mid-tropospheric contributions to synoptically-forced ascent during the cool seasons (Oct-Apr) from 1989-2008. Modes of variability in the position of the ascent based synoptic pathways are examined by means of principal component analysis (PCA), the results of which are examined for links with ENSO.

This paper is organized as follows: The remainder of Chapter 1 will review pertinent literature. Chapter 2 will provide details on the data sets employed and Chapter 3 will explore the methodology used for calculating synoptic scale vertical motions. The climatological results will be provided in Chapter 4, and the implications of these findings will be addressed in Chapter 5. The last section of this paper, Chapter 6, will provide a summary of the methodology and findings.

Prior Research

There is a rich history of research concerned with the distribution of synoptic-scale transient disturbances and their favored locations, or storm tracks, generally examined separately in terms of lower-, mid- or upper-tropospheric phenomena. The specific techniques used to examine these phenomena tend to fall into two major categories: (1) feature tracking, in which features are identified and their location and attributes through time are tracked, and (2) filtered variance, wherein tropospheric fields, such as geopotential height or relative vorticity, are filtered on synoptic-scales either temporally (order several hours to days) or spatially (order 0.5-5 thousand km) or both.

Lower-Troposphere

Interest in preferred storm tracks originated with early studies of Atlantic cyclones, which were seen to be responsible for the preponderance of sensible weather. Numerous studies have examined the Northern Hemisphere distribution of cyclone tracks, typically identifying and tracking closed contours in mean sea level pressure (MSLP). A small subset of these studies is relevant to western North America. Reitan et al. (1974) provide a cyclone frequency climatology for the North American continent and adjacent oceanic domain based on cyclone track data from the National Weather Service (NWS). A northern track roughly parallel to the United States/Canadian border was evident during January with cyclogenesis events strongly favoring the lee side of the Rocky Mountains in both Alberta and Colorado. During April, a track from the Gulf of Alaska east through British Columbia is found in addition to a region of high cyclone frequency anchored in Southeastern Colorado and extending northeast to the Great Lakes. Cyclogensis during April is maximized in a swath across the southwestern US and extending to an absolute maxima in Colorado.

Zishka and Smith (1980) developed a cyclone climatology, based National Oceanagraphic and Atmospheric Administration (NOAA) Environmental Data Service cyclone track maps spanning 1950-1977, consistent in many respects with Reitan et al. (1974). In addition to the total cyclone counts, a metric was introduced to delineate regions where the interannual variation in the number of cyclones is low yet the total storm count is high, suggesting preferred pathways for these mobile systems. For January, both the cyclone counts and regions of low ‘relative variability’ correspond well with the northern track demonstrated by Reitan et al. (1974). The intermountain west is characterized by infrequent cyclones and high relative variability.

While Zishka and Smith (1980) only addressed cyclone distributions for January and July, Whittaker and Horn (1981) provided a more comprehensive treatment of the seasonal cycle of cyclogenesis, though did not explicitly address cyclone tracks. Similar to Rietan (1974), the Great Basin (intermountain west) has a cyclogenesis peak in April. Colorado cyclones formation was found to be favored in March. For the domain as a whole, cyclogenesis is more frequent in the late winter before diminishing and migrating north during the boreal summer.

With the availability of reanalysis products as well as the prospect of shifts in storm tracks due to a warming climate, a renewed focus has been applied to deduce the distribution of cyclones during the past ten years. As a portion of a comprehensive survey of numerous measures of storm tracks for the Northern Hemisphere, Hoskins and Hodges (2002; hereafter HH), presented feature-tracking climatologies for MSLP minima and 850 hPa relative vorticity maxima (comparable to tracking surface cyclones) in the ECMWF ERA-15 during 1979-2000. HH uses two primary measures of storm tracks for these lower tropospheric fields: (1) feature density and (2) track density. Feature density is biased somewhat towards slowly moving systems making large contributions to total density, while track density restricts the contribution from a single system. The track and feature densities for MSLP are both qualitatively similar with the results of Reitan et al. (1974) and Zishka and Smith (1980) with a northern storm track consisting of tenuously-connected maxima across the Canadian Rockies, a local maximum just downstream of the Rocky Mountains in Colorado, and high densities in the Gulf of Alaska. The mean intensity for MSLP features shows a tongue of locally high intensity systems extending from the Pacific Northwest through the intermountain west and onto the high plains of Colorado, indicating that the relatively weak track densities for intermountain region may correspond to strong systems. Although MSLP and 850 hPa relative vorticity yield broadly similar results for the hemisphere as a whole, significant differences can be found across western North America. The 850 hPa relative vorticity statistics exhibit near zero densities for tracks and features over elevated topography and a coastal band of higher feature and track densities south to Baja California. This mountain minima and coastal maxima is not seen in the MSLP diagnostic.

Following a methodology developed by Blackmon (1976), HH also employs a 2-6.5 day filtered variance (standard deviation) metric for MSLP and 850 hPa relative vorticity, the results of which are similar the above tracking statistics. Variance in MSLP maximizes near 50⁰ N and is connected to upstream and downstream maxima over the Pacific and North Atlantic, respectively. There is a ‘ridge’ of zonally minimized variance centered on the western coast of North America, suggestive of the mean climatological ridge in geopotential heights. By comparison, variance in 850 hPa relative vorticity is everywhere minimized over the axis of the Rocky Mountains, showing no semblance of a west-to-east connected storm track, a result which is similar to the tracking statistics. However, the 850 hPa vorticity variance does not indicate the elongated coastal maxima present in the vorticity tracking version. In understanding these results, it should be remembered that the MSLP and 850 hPa relative vorticity fields require vertical interpolation due to the elevated landmass and complex topography of the western United States, which calls into question their representativeness for the region.

Noting the difficulty of using extrapolated fields in the presence of complex elevated topography, Wernli and Schwierz (2006) excluded much of the western US from a recent hemispheric climatology of MSLP cyclones using the ERA-40 reanalysis. However, the western US domain was subsequently examined by Jeglum and Steenburgh (2010, hereafter JS), who relied on multiple reanalysis products (NARR, ERA-40,ERA-interim) and both the MSLP and 850 hPa geopotential height fields to examine cyclone density. While the results of the study are found to be sensitive to the model resolution, with higher resolution products yielding higher cyclone counts, a consistent picture of multiple local maxima in cyclone density within the intermountain region is found. Furthermore, the total cyclone counts in the intermountain region is found to be comparable in magnitude to that found in the eastern Gulf of Alaska, which suggests previous cyclone studies not resolving an intermountain maximum may have suffered from data sparseness or coarse grid scales. Although HH did find a weak center of storm activity in the intermountain west, JS indicate this maximum is located further south, roughly in the lee of the high Sierra Mountains, which is in closer agreement to the distribution previously found by Klein et al.(1968). They examined twice daily National Meteorological Center analyses during winter from 1951-1964 and identified closed centers of circulation at four pressure levels (850 mb, 700 mb, 500 mb and 300 mb) for a domain centered on the intermountain west. Feature density, at 850 hPa, shows numerous local maxima, with the highest densities found just downstream of the Rocky Mountains with secondary maxima located from the Gulf of California extending north along the tri-state junction of California, Arizona, and Nevada (which is near the JS maxima). The distribution of upper level features was highest near the Gulf of California.

In one of the few cyclone studies to consider interannual variability, Myoung and Deng (2009) present a coastal cyclone activity function (CAF), which is a one dimensional measure for cyclones along the western coast of North America. Year to year changes in the CAF is presented in terms of the leading modes of variability that result from PCA. Their second mode exhibits a north-south dipole pattern, the time series of which is correlated with sea surface temperature in the Nino 3.4 region. Myoung and Deng (2009) also examine the fraction of winter season precipitation for the western US that can be attributed to coastal cyclones as well as how the modes of cyclone variability relate to variations in regional precipitation. Their findings demonstrate that coastal precipitation is strongly linked to coastal cyclones while precipitation in the interior western US is less directly associated with these oceanic storms.

Mid-Troposphere

An alternate perspective on storm tracks is obtained by analyzing mid tropospheric phenomena. To this end, Blackmon (1976) filtered the 500 hPa geopotential height field over the northern hemisphere so as to isolate synoptic-scale transients, which were generally assumed to reside in frequencies of 1-6 days and with zonal wave numbers of 7<n<18 (representing medium and short wave features). For a subset of medium waves (7<n<12) within the band of 2.5-6 days, the resulting ‘storm track’ is seen as to be more or less zonal and centered near 42⁰ N with strong connectivity across the western United States. This contrasts with any of the aforementioned cyclone studies which showed only tenuous continuity across this same region. Shorter waves (13<n<18) within the same timescale have centers of high variance further north, with a weak connection across the Canadian Rockies near 55⁰ N. Yet, short wave features within the 1-2 day temporal band show a stronger connection in this same region. Collectively these results show sensitivity to the wave and time scale of choice with medium waves generally found to have a maximum variance further south than shorter waves. As compared with cyclone studies, greater east-west connectivity is observed in the variance maxima and no Great Basin or Gulf of California maximum is found. Interestingly, Blackmon (1976) also presents a climatology of 500 hPa vorticity wherein high values of cyclonic vorticity tend to correspond to regions of high filtered variance for northern potions of the domain, but fail to coincide across the southwestern US where variance was generally low. The presence of the documented strong southern vorticity maxima is, however, roughly in agreement with cyclone density centers observed by JS as well as Klien (1968).

HH extended their storm track analysis to the filtered variance and tracking of mid-tropospheric (500 hPa) omega (vertical motions). Variance is maximized in a roughly zonal band, centered near 33⁰ N, which possesses a weak sinusoidal meander from the eastern Pacific across the western US. The position of this inferred omega storm track is considerably south that of lower tropospheric variances, as well as the variance documented by Blackmon (1976) which maximized between 45⁰ N and 55⁰ N. The tracking statistics for 500 hPa omega yield a less continuous pathway. Track density is minimized over the continental divide and exhibits more strongly meridional features than the variance metric. Specifically a northwest-southeast maximum in track density is found along the northwest coast of North America, which then extends inland towards Arizona. The HH results for the 500 hPa omega metrics bear similarity to those employed in the present study and will be further discussed in Chapter 5 of this paper.

Vertical motions and the variance of geopotential height are closely related to the presence of troughs in the 500 hPa surface, climatologies of which have been compiled by Sanders (1988), Lefvre and Nieslon-Gammon (1995; hereafter LNG), Dean and Bosart (1996), and Hakim (2000; hereafter HK). Sanders (1988) subjectively tracked troughs along the 552 dam contour of the 500 hPa surface for nine cold seasons. Results were presented in the context of tough-lysis and trough-genesis and generally showed an excess of lysis events upstream of the Rocky Mountains and genesis events downstream. Little information about the overall density of these features was provided in the study. In a comprehensive and objective climatology of 500 hPa troughs, LNG provide a more inclusive study based on a measure of Eularian Centripetal Acceleration (ECA), which is the product of the geostrophic wind and the geostrophic curvature vorticity. Maximum values of ECA are typically collocated with tough axis in the height field, and thus it serves as an objective measure of location and intensity that can be readily tracked. The resulting track statistics for western North America show a connected region of maximized feature frequency that meanders south from the Gulf of Alaska to northern Baja California, before turning zonally along the US/Mexico border then northeast towards the Great Lakes. Local maxima are found within the ‘storm track’ in the Gulf of Alaska, west of Baja California, and over the Great Lakes. In southern British Columbia a weak indication of a more northern track penetrating the otherwise minimized frequency values is found. .

Dean and Bosart (1996) provide a similar perspective on 500 hPa trough distribution as a portion of a hemispheric study on trough merger and fracture. Using a trough identification criteria based on closed circulation centers, defined as local maxima in geostrophic absolute vorticity relative to a zonal base state, the total number of trough occurrences is presented for the period of September 1957 through May 1989. Findings show a strong maximum in occurrence for the Gulf of Alaska, a local maximum over southern California and a weak track across the Canadian Rockies proximal to the US border. There is also a local maximum found near south east Colorado and the northern panhandle of Texas.

Approaching 500 hPa trough distributions from a potential vorticity (PV) perspective, HK provides a climatology of 500 hPa vorticity maxima (VM) for 33 DJF winter seasons. The PV framework, explored at length in Hoskins et al (1985), is one in which downward deflections of the dynamic tropopause surface (1.5 Potential Vorticity Units (PVU) in this case) are seen as the source of strong cyclonic relative vorticity features in the mid troposphere. The climatology of VM presented by HK is given as the percentage of the total number of time steps for which a VM is present at a grid point as compared to the total number of time steps during the study period. As such, these results are not identical to tracking statistics, but nonetheless provide a sense of synoptic pathways. Results are divided into the distribution of vorticity features within four ascending classes of intensity. Moderate (4 × 10-5 s-1 ≤ ζg ≤ 4 × 10-5 s-1 ) 500 hPa relative vorticity frequency is shown to be maximized in a connected band from the Gulf of Alaska eastward into the Canadian plains, with an embedded maximum in the lee of the Canadian Rockies.. The distribution of strong and extreme vorticity (8 × 10-5 s-1 ≤ ζg ≤ 30 × 10-5 s-1 ) features is quite different, showing a connected maxima of high observed frequency from the Gulf of Alaska south along the Pacific seaboard, then curving east across the northern Gulf of California, and eventually northeast into the central US. Embedded maxima are found both in the Gulf of Alaska and the Gulf of California regions. The Gulf of California maximum is comparatively stronger when only extreme events are considered. This distribution of synoptic pathways is spatially similar to that documented in both LNG and Dean and Bosart (1996). Not surprisingly, this distribution is also similar to the 500 hPa vorticity climatology included in Blackmon (1977)

A related PV perspective on the connection between upper level tropopause features and mid-tropospheric response is provided by Elbern et al (1998). In a climatological study of tropopause folds, which are assumed to represent a subset of strong mid level troughs, the distribution of joint occurrences of the intersection of the 1.6 PVU surface and the 400 hPa level and strong values of divergent Q-vectors (representing downward motions) are documented. This dual criterion was designed to identify sub grid scale tropopause folds in coarse gridded analysis products and was tested against higher resolution mesoscale simulations. For the winter season, the tropopause fold distribution is spatially quite similar to the vorticity distribution documented by HK, with a local maximum over the northern Gulf of California. Results for both Elbern and HK will be further explored in the subsequent discussion section of this paper.

Upper-Troposphere

Moving now to exclusively upper tropospheric measures of storm track, HH present a number of fields that were both tracked and analyzed for temporal variance. These fields include geopotential height at 250 hPa, relative vorticity at 250 hPa, potential vorticity on the 330 K isentropic surface, and potential temperature on the 2 PVU surface. The two variance measures on the 250 hPa surface both show a continuous storm track across western North America but differences in the north-south location of the track are not trivial. Variance in the height field exhibits a broad connected maximum centered near 50⁰ N, while the narrower region of maximum vorticity variance is center further south at roughly 42⁰ N. Both fields are more or less zonal in the distribution of large variances and show lower maximum values proximal to the Rocky Mountains. In contrast, the variance of the potential vorticity on the 330K isentropic surface has a local maximum over the intermountain west which is embedded within a connected ribbon of maximized variance centered near 43⁰ N. The intermountain west variance maximum is larger than any values found upstream over the Pacific, though still lower than those found in the Atlantic storm track. Yet another perspective is provided by the variance in the potential temperature of the dynamic tropopause surface, which shows an elongated pacific maximum that extends zonally into the southwestern US near 30⁰ N.

HH tracking statistics for each of the above fields does little to solidify the dispersed representation of the high altitude storm track, and in fact may further complicate it. Track densities for negative 250 hPa height anomalies are located somewhat south of the variance maximum for the same field. The connectivity of the track across the region is modest, showing numerous local maxima within the western US, including a well defined maximum in the lee of the Rocky Mountains. The relative vorticity track density is considerably different than both the associated variance field and the related 250 hPa height fields. Specifically, the track density exhibits a sinusoidal pattern where in maxima are displaced north near the Pacific Northwest and then south towards the Gulf of California before turning east and northeast into the central US. There is a weaker connected track just north of the US/Canadian border. This perspective is strikingly similar to the HK 500 hPa relative vorticity frequency, yet unlike all of the other upper tropospheric fields. Tracking statistics for PV 330K and potential temperature at 2PVU show reasonable agreement with their associated variance fields.

In light of the divergent depictions of the meriodonal location of the upper-tropospheric storm track, it would seem reasonable to conclude that each of these fields are dominated by different processes, which while interpretable, does not provide a clean picture of the location of the storm track aloft.

The uncertainty of upper tropospheric synoptic pathways can be generalized at all lower levels with no singular consensus in storm track locations emerging. This is not to say that there is complete disagreement in the distribution of storminess, as we have seen that elements of numerous studies suggest similar geographic structures, such as maxima near the northern Gulf of California (LNG,HK), and hint at vertically connected processes such as lee cyclogenesis in southeast Colorado (HH). Nor should dissimilarities between different levels be taken as erroneous representations of a single process, but instead as providing representations of numerous processes related to synoptic transients. For example, it seems intuitive that we find more connected mid- and upper-tropospheric fields where open wave features readily propagate whereas surface fields are noisy in the presence of complex topography. This is especially true for cyclone fields, where there is a necessary spin up time and length scale required to generate a closed circulation. On the opposite end of the spectrum, fields that examine approximately conserved quantities, such as absolute and potential vorticity, require no time scale for development as their presence is more or less continuous in time. While such technique based differences are explicable, a more concise representation of the regional storm track would be useful.

The following sections of this paper will attempt to provide a clearer interpretation of the storm track across western North America.

chapter 2

DATA AND METHOD

*Reanalysis Data*

Data from the European Center for Medium Range Weather Forecasts (ECMWF) ERA-Interim global reanalysis provide the four dimensional representation of the state of the atmosphere used in this study. The ERA-Interim is a state of the art reanalysis system that covers the ‘data rich’ period from 1 Jan. 1989 through the present. Analyses are produced using the Cy31r1/2 version of the ECMWF Integrated Forecasting System (IFS) configured with a T255 triangular truncation horizontal grid and 60 vertical levels (ECMWF newsletter 110,111,115). The analysis system employs a four-dimensional variational assimilation (4D-Var) system that assimilates a broad range of observations within a 12-hour window. Data sources include cloud-track winds, satellite radiances, satellite scatterometer winds, radio-occultation measurements (since 1996), surface observations (wind, temperature, pressure, and humidity), operational radiosonde data, altimeter wave heights, and ozone profiles. Significant improvements over past ECMWF reanalysis products (ERA-40, ERA-15) are due in part to the improved 4D-Var system and improved model physics but also due to a variational bias correction system. As described by Dee and Uppala (2009), variational bias correction helps to maintain consistency in the analyses in the presence of ‘data events’ wherein the number and quality of observations changes in time, including the introduction of new data streams.

The quality of the reanalyses over the western United States depends on the model’s resolution (~80 km) of the complex topography of the region (Fig. 2.1a). A qualitative comparison of the ERA-interim terrain to that obtained from a 1/3° (~36 km) digital elevation model (Fig. 2.1b) suggests that the ERA-interim coarsely represents the regional orography, failing to resolve important features such as the “high” Sierra Nevada Mountains near their southern terminus in California. Terrain smoothing is evident as well elsewhere in the domain such as central Nevada, where the cross barrier scale of individual mountain ranges is small.

The gridded model data used in this study were obtained from the ECMWF data server (http://data-portal.ecmwf.int/data/d/interim\_daily/). The region bounded by 22-58° N and 70-160° W (Fig. 2.1a) was selected for this study and covers the majority of the contiguous United States and an adjacent portion of the eastern Pacific Ocean. Post-processed geopotential, temperature, relative humidity, zonal and meridional wind, vertical velocity, and relative vorticity are provided at 6 hourly time steps while accumulated precipitation is available at 12 hour intervals. Dynamic fields are available on 37 isobaric surfaces at a reduced horizontal resolution of 1.5°(~165 km). Although additional levels will be used in Chapter 3 to evaluate the analysis approach, this study focuses most heavily on five isobaric surfaces (700,600,500,400,300 hPa) with the lowest pressure surface, 700 hPa, above the model’s terrain except for a single gridpoint in central Colorado.

*Computing Omega*

The spatial and temporal distribution of synoptic-scale vertical motions in mid-latitude weather systems has long been recognized as an important factor in precipitation and storm intensification processes. Ascent leads to the supersaturation necessary for precipitation and clouds (Rose and Lin 2003) as well as alters the environmental lapse rate, which in turn may favor mesoscale vertical motions (Durran and Snellman 1987). Differential vertical motions may stretch columns of air vertically and generate cyclonic vorticity (Martin 2007). Using quasi-geostrophic reasoning, vertical motions are viewed as part of a secondary circulation necessary to maintain thermal wind and hydrostatic balance of the primary geostrophic flow (Hoskins et al 1978, Hosking and Pedder 1980, Durran and Snellman 1987). These vertical motions are commonly viewed as being synoptically “forced” vertical motions and on the order of 0.1 m/sec (~-.1 Pa/sec), which is 2 orders of magnitude smaller than quasi-geostrophic scaling assumptions for horizontal motions.

Numerous formulations of the quasi-geostrophic omega equation have been devised starting with Sutcliffe (1947), revised and simplified to eliminate term cancelation by Trenberth (1978), and addressed at length by Hoskins et 1978 and Hoskins and Pedder (1980) who introduced the concise Q-vector notation. Subsequently, Robert Davies-Jones (1991, 2009) presented a generalized version of Hoskin’s Q-vector omega equation. Specifically, Davies-Jones (1991, 2009) demonstrated that an alternative balance (AB) form of the omega equation that reflects a closer approximation to the full primitive equations can be obtained by substituting the non-divergent component of the observed wind for the geostrophic wind in the calculation of the Q vector. A similar approximation was previously employed by Keyser (1988) to generalize the Petterssen frontogensis function in a Q-vector framework. The present study follows the conceptual frame work of the Hoskin’s Q-vector approach but uses the Davies-Jones substitution of the non-divergent wind field, in place of the geostrophic wind, in order to estimate synoptic-scale vertical motions.

To obtain the non-divergent wind the numerical method detailed by Krishnamurti and Bounoua (1995) is followed, wherin the ERA-interim wind field is split into the areal average, nondivergent, and irrotational components:

(Equation 2.1)

where is the streamfunction and is the velocity potential. Ignoring spherical geometry:

(Equation 2.2)

(Equation 2.3)

Thus, by using the relative vorticity ( calculated from the ERA-interim wind field, we may solve for on each pressure level using a successive over-relaxation algorithm.

Having obtained the nondivergent component of the wind is then simply

(Equation 2.4)

(Equation 2.5)

An example of this decomposition is provided in Chapter 3**.**

Using the nondivergent component of the observed wind, the generalized version of the Q-vector form of the omega equation is:

(Equation 2.6)

where is the static stability parameter, is the Coriolis parameter and is the vertical wind in pressure coordinates (Pa/sec). The Q vector is defined as:

(Equation 2.7)

where all horizontal derivatives are taken on isobaric surfaces, R is the dry air gas constant, P is the pressure level (Pa), is the vector non-divergent observed wind (m/s), and T is the temperature (K). The units of are K/m/s. The generalized Q vector may be interpreted as the vector rate of change of the horizontal temperature gradient following the motion of a fluid particle moving along a trajectory defined by the non-divergent observed wind. Alternately, one may view the *Q* vector as describing the lower branch of the secondary ageostrophic circulation and points toward regions of rising motion (Hoskins and Pedder 1980). To the extent that the right hand side (RHS) of the omega equation can be assumed to be sinusoidal, the left hand side (LHS) is commonly approximated as

(Equation 2.8)

which implies that convergent *Q* vectors can be associated with rising motion. While this approximation helps to provide context, an accurate representation of synoptic-scale vertical motions requires solving for omega from the omega equation (Clough 1996). Before proceeding with this calculation, it is useful to consider the meaning of the individual terms on the LHS of the omega equation:

(Equation 2.9)

**A B**

When integrated with appropriate lateral and vertical boundary conditions, the first term (A) represents the horizontal coupling and the second term (B) provides vertical coupling of at different levels and locations. Consequently, it is important to consider the full three dimensional field of the RHS in order to properly resolve the magnitude and sign of the vertical motions for a specific location. Expanding on earlier work by Durran and Snellman (1987) and using the electrostatic concept of action at a distance, Clough (1996) examined the impact of point, dipole, and quadra-pole spherical distributions of as a function of their location in the vertical, latitude, and stability. He demonstrated that increasing latitude or static stability confines the vertical and horizontal influence of . The influence of multiple sources of, especially positive-negative dipoles, localizes omega and confines it to a spatial pattern similar to that of (which supports the sinusoidal approximation mentioned above). Durran and Snellman (1987) showed earlier that longer wavelength features within the distribution of dominate, which leads to smoother appearing omega fields compared to the RHS term. Hence, small-scale features in the reanalyses resulting from poorly resolved terrain-induced or convective processes will tend to be less apparent in the final omega field relative to synoptic-scale features. The case study in Chapter 3 will support these statements.

Since the post-processing of the reanalyses reduces the horizontal resolution to ~ 160 km, the horizontal gradients in wind and temperature, the convergence of the Q vector, and the magnitude of the derived vertical motions are reduced as well. Elbern (1998) showed that a mesoscale model provided considerably larger values of than those derived from coarse ECMWF analysis grids. About 70% of the difference in was attributed to reduced gradients of geostrophic wind with the remaining fraction due to reduced gradients of potential temperature. Hence, the magnitudes of synoptic-scale vertical motion derived from the post-processed ERA-Interim reanalyzis likely underestimate those that could be obtained from the original T255 grids.

The common method of successive over relaxation is used to solve the full three-dimensional omega equation (Stuart 1967). Homogeneous zero boundary conditions are used for both lateral and vertical boundaries except in the case study presented in Chapter 3 where the impacts of the vertical boundary conditions are documented. Vertical motions resulting from poorly-resolved mountain flows are eliminated by setting the bottom boundary to zero, which also helps to filter other mesoscale vertical motions.

*Statistical Methods*

The climatological distribution of synoptic scale ascent, as calculated using the omega equation, is established by an accumulation method. The maximum ascent (specifically, minimum omega less than a threshold of -.075 Pa/s) at 700, 600, 500, 400, or 300 hPa at each grid point is stored for each 6-h period; thus, a single field of maximum ascent is obtained for each time period. The threshold value of -0.075 Pa/s was established from prior studies and subjective evaluation of preliminary results using this specific data set. The maximum ascent at each grid point is subsequently accumulated over case studies or monthly or seasonal periods and then divided by the total number of time intervals within the samples to obtain estimates of mean synoptic-scale ascent that is suitable for assessing lift associated with time scales spanning from a single synoptic-scale system to the entire 19-year climatology. In addition, the mean ascent intensity (Pa/s) is obtained by restricting the relevant sample at each grid point to only those times when ascent greater than the threshold is observed. Finally, the mean ascent frequency is the number of times ascent greater than the threshold is observed at each grid point divided by the total number of 6-hour time steps in the sample. This metric represents the percentage of time a grid point experiences significant synoptic-scale ascent.

The climatological mean ascent, ascent intensity, and ascent frequency are derived using maximum ascent values over the 19 cool seasons (Oct-Apr). Collectively these three metrics are measures storm tracks in that indicate regions for which synoptic-scale ascent is locally prevalent and intense. Interannual variations for each parameter at each grid point are computed from the appropriate standard deviation of the individual seasonal means about the climatological mean. Cool-season standardized anomalies for each year are then calculated for each grid point by removing the climatological mean and dividing by the standard deviation. Month-to-month variations are examined similarly.

The correlation matrix derived from the 19 cool-season standardized anomaly fields are examined for coherent modes of interannual variability using principal component analysis. Empirical orthogonal functions (EOFs), which represent spatial patterns of anomalies in ascent, are obtained by projecting the initial data onto the vector space defined by the eigenvectors (principal components) of the correlation matrix. The correlation between the pattern of ascent in a given year and each respective EOF is provided by the set of principal component time series. The eigenvalues associated with each of the principal components, and thus each EOF, define the amount of interannual variance explained by that principal component. Correlations between the principal component time series and the time series of indices of known modes of interannual climate variability as well as seasonal precipitation anomaly time series help to assess to what extent these modes represent physical processes. Statistical tests are used to examine the robustness of the resulting correlations.

chapter 3

Calculating Omega

The purpose of this chapter is to illustrate calculating omega during a strong synoptic disturbance that traversed the western US on 15 April 2002. The structure and evolution of this system, dubbed the ‘taxday storm’, was examined in detail by West and Steenburgh (2010; hereafter WS10). For clarity, all the figures presented in this chapter are centered on the western United States, even though the calculations were completed for the entire grid introduced in the previous chapter.

Figure 3.1 (needs the date in the caption and the cross section) demonstrates the process for computing the nondivergent component of the ERA-interim analyzed wind at a time of strong ascent associated with the taxday storm. Following the removal of the zonal and meridional mean wind from the entire domain, the relative vorticity is calculated (Figure 3.1a). As shown in Fig. 3.1b, the stream function, is then calculated from the relative vorticity field and the non-divergent components of the wind derived from the gradients in .

The approach used to determine ascent is best illustrated by cross sections along the transect WE in Fig. 3.1 as the taxday storm evolves (Figures 3.2-3.8) **(full standalone figure captions).**  The top panel in each figure shows the ERA-interim vertical motion field on the post-processed grid while the bottom panel shows the vertical motion obtained by solving the AB omega equation. Common to each panel for reference is the dynamic tropopause (2 PVU) surface and distribution of potential temperature while the dashed lines in the lower panels indicate the level of maximum ascent for gridpoints where the derived omega is less than -0.75 Pa/sec at one or more levels.

The tax day storm was associated with a significant upper-tropospheric disturbance progressing from west to east through the cross-section, although strong near -surface frontogenesis took place during the later stages (WS10). Overall, the vertical motion derived from the AB omega equation exhibits a smoother distribution than the post-processed omega and tends to associate vertical motion more closely with the tropopause disturbance and associated strong gradients in potential temperature that extend downward into the troposphere. This is particularly evident at many times during the evolution of the storm when the ERA-interim shows rising motions downstream and well removed from the tropopause depression that are likely related to downstream mountain-flow interactions. For reference, the high surface PV values often evident in the cross sections between 110-105 W reflect the high terrain of Colorado, which is associated frequently at these times with an orographically-forced omega couplet. Hence, the absence of such vertical motions in the AB omega field is considered appropriate since we are interested in the motions directly associated with synoptic transients and not orographic or other mesoscale-induced motions.

The magnitudes of the ERA-interim omega tend to be larger than those derived from the AB omega equation, especially where heavy precipitation and strong latent heat release is evident in both observations and in the ERA-interim reanalyses (Figs. 3.5 and 3.6). In addition to the fact that ascent arising from convective processes is intentionally attempted to be filtered, the ascent derived from the AB omega equation assumes dry static stability, which is an overestimate of the stability when the atmosphere is saturated. As discussed in the previous chapter, higher stability implies weaker derived vertical motions.

Further insight into the role of stability, the number of vertical layers included in the integration, and the impact of lower boundary conditions on the solution to the AB omega equation is provided in Figs. **3.9-3.10.** The top panel in each figure shows the RHS of the AB omega equation, while the subsequent panels show the derived omega computed with different assumptions. As discussed in Chapter 2, the tends to have finer-scale structures than the derived omega fields. The impact of static stability is quite evident, since the magnitudes of the derived omega fields are greatly reduced in the stratosphere as a result of high static stability and increased somewhat in tropospheric regions of weak static stability compared to the corresponding magnitudes of in those respective regions. **S**olving the omega equation using only 5 vertical layers as employed in the next chapter does not substantively change the distribution of derived vertical motion (compare the b and c panels in the two figures). Similarly, minor differences are apparent if the 5-layer solution uses lower boundary conditions prescribed by the ERA-interim 700 hPa omega compared to setting the lower boundary condition to be zero (compare the c and d panels). Hence, all subsequent results employ zero lower boundary conditions and 5 pressure levels for computational efficiency.

Figs. 3.11 (12z fig messed up)-3.16 show the evolution of maximum omega derived from the AB omega equation at six hour time steps from 06 UTC 15 April 2002 through 06 UTC 16 April 2002 relative to radar and incrared satellite imagery. The maximum ascent in excess of the specified threshold could be taking place on any of the five pressure surfaces from 700-300 mb while the descent values are displayed only if there was no ascent exceeding the ascent threshold at any level. Through the early phases of tax day storm, a quasi-symmetric dipole of rising and sinking motions is found centered about the 500 hPa trough axis. This dipole pattern breaks down later in the event, and vertical motions become less spatially coherent and predictable in their locations and intensity from simply the location of the mid-tropospheric trough axis. Beginning at 00 UTC 16 April (Fig. 3.14), upwards motions become elongated in an arc and partially encircle a region of downward motion. Over the next 12 hours, the areas of strongest vertical motion bifurcate into weaker and more localized centers. The symmetry in structure and magnitude across the trough axis is no longer evident as upstream of the trough a robust region of sinking motions can still be found despite the more dispersed downstream features.

The agreement between the distribution of clouds and precipitation and the location of rising motions tends to increase through this event. Initially, at 12 UTC, much of the strong synoptically driven rising motions are found in the cloud sparse air upstream of an antecedent region of clouds and precipitation in northern Utah and southern Idaho (analyzed by WS10 as an older frontal boundary). However, by 18 UTC 15 April, the distribution of ascent and the precipitation and cloud structures are in better agreement, as a new region of clouds are generated in central Nevada and in an arc north through Idaho and Montana as well as precipitation is greatly enhanced.

The increasing relationship with time between precipitation and rising motion is likely a consequence of the timescales required to drive air first to saturation and then supersaturation. To elucidate this scenario, a simple scaling argument can be applied to a hypothetical parcel originating near the base of the trough. A parcel would need to ascend to roughly 600 hPa to achive sufficient cooling to reach saturation if we assume a starting condition of 700 hPa, 0⁰ C, and 50 % relative humidity. If we further assume that the parcel is being advected at a rate of 20 m/s through a region with ascent of 1 Pa/sec (~ 0.1 m/s), the parcel would reach saturation in approximately 10,000 sec and have traveled downstream by about 200 km. Examining the location of the nascent cloud arc at 06 UTC, it is found to initiate approximately 200 km away from the trough axis along a streamline, which is roughly consistent with this simple scale analysis. Hence, while the cloud free air near the trough axis may seem inactive, it may actually be a vital preconditioning zone for air that will eventually be involved in precipitation processes. In the case of the tax day storm, rising motions intensify quickly as the approaching trough digs south and collapses in wavelength.

Figure 3.17 summarizes the mean ascent during the tax day storm. As described in Chapter 2, only maximum ascent with omega less than -0.075 Pa/s is accumulated from each time step, summed in this case over the 36 h period, and then divided by the 6 contributing time steps. The storm mean ascent suggests that the greatest vertical motion begins off the Oregon coast, intensifies across Nevada, and eventually extends across parts of Utah, Idaho, Wyoming and Montana. The splitting of maximum upward motion, that begins at 00Z 16 April is evident as well. Hence, this event would be described as having a storm track that begins along the west coast, reaches peak intensity over eastern Nevada, and then splits into two distinct tracks later on.

Animations of maximum ascent for every 6 h period during the 19 years were created and extensively examined to investigate other cases (not shown). The correspondence of the derived vertical motion with synoptic experience on the evolution of cool-season storms was quite apparent. Hence, it is reasonable to derive ascent statistics for any timescale of interest from individual cases to entire seasons, and this approach appears to represent a scalable approach to document storm tracks.

chapter 4

Results

*Climatological-Mean Synoptic-Scale Ascent*

Summary statistics for the geographic distribution and variability in synoptically-forced ascent are presented in this chapter. The methods required to compute synoptic-scale ascent during each 6 h period within the 19 cool seasons (Oct 1989-Apr 2008) have been described in the previous two chapters. The climatological-mean ascent is presented in Figure 4.1a**.** A continuous belt of high mean ascent (<-0.02Pa/sec) is found in a roughly sinusoidal pattern across the domain and will be described here as defining the climatological mean storm track across the United States. Ascent is maximized in a broad region from the western edge of the domain northeastward to the British Columbia coast, where the first of three local maxima (labeled A in Fig. 4.1a) is encountered. Maximum A, with a value near -0.03 Pa/s, is located in a region well known for its frequent storminess and copious precipitation. The band of synoptic-scale ascent then swings south in a narrow band along the western US coast before penetrating inland across Northern California. Mean ascent increases across southern Nevada and then reaches a second maximum, labeled B, over southern Arizona, which is one of the most arid regions of the nation. The apparent inconsistency between synoptic-scale ascent and precipitation in this region will be discussed in Chapter 5.

The storm track defined in terms of climatological-mean ascent curves to the east-northeast splitting somewhat across New Mexico into portions of Texas and Colorado before reaching a third weaker maximum, labeled C, over Oklahoma. As will be discussed further in Chapter 5, this third maximum is found slightly east of the eastern Colorado surface cyclone maximum. From central Oklahoma, the primary ascent pathway continues northeast to the Great Lakes region. The lowest amount of synoptic-scale ascent within the midlatitudes is found over the upper Missouri River basin.

The climatological frequency of ascent, Fig. 4.1b, provides a nearly identical spatial perspective regarding synoptic ascent as shown in Fig. 4.1a. Regions along the climatological-mean storm track experience synoptic-scale ascent greater than the threshold roughly 13% of the time (or once every 7.5 days) with local maxima experiencing ascent approaching 20 % of the time.

Also apparent in both the mean ascent and frequency metrics is a much weaker secondary storm track that stretches zonally across the Canadian Rockies in southern British Columbia, east across the high plains to Lake Winnipeg, and then rejoins the primary storm track in the Great Lakes Region. This feature, which is likely associated with Alberta clipper storms, will be discussed further in the next chapter.

A different perspective on synoptic-scale ascent is provided by Fig. 4.2, which shows the climatological mean ascent intensity, defined in Chapter 2 to be the total accumulated ascent at each gridpoint divided by the number of 6 h periods when the ascent is greater than the specified threshold at each gridpoint. This metric simply defines the average maximum ascent independent of the number of times it actually takes place. For example, storms that reach 35oN, 135oW tend to have strong rising motions when they occur (Fig. 4.2) but, since they don’t happen as often (Fig. 4.1b), the climatological mean ascent in this area is relatively low (Fig. 4.1a). Of particular interest is that ascent over Arizona (northeastern Montana) occurs fairly frequently (infrequently) and the intensity of that ascent is high (low).

*Seasonality of Synoptic-Scale Ascent*

The climatological seasonal cycle of synoptic-scale ascent is presented in Figs. 4**.**3-4.5. The long term monthly mean ascent is shown in Fig. 4.3 while each month’s contribution to the entire cool-season’s accumulated ascent is shown in Fig. 4.4. For convenience, the panels for October are omitted from those figures. The month-to-month change in mean ascent is shown in Fig. 4.5, which highlights intraseasonal shifts in the location and magnitude of the storm tracks. Collectively, these three metrics show that regions of strong ascent migrate southward through February before advancing north again during March and April, which is consistent with the seasonal progression of mean upper-tropospheric ridging over the western North America. For the domain as a whole, December storms contribute the largest amount of ascent of any particular while February exhibits a more zonal pattern in the distribution of mean ascent compared with other months.

Embedded within these broad trends are important regional and intraseasonal variations that are recognizable from synoptic experience. Relative to October, mean ascent increases in most regions from October to November and shifts southward along the west coast (Fig. 4.5a). This southward shift continues from November to December with more ascent in December particularly evident in northern California and along a broad swath from Northern Mexico across the eastern US. Ascent generally is weaker during January throughout much of the domain except notably over the eastern Pacific Ocean and Gulf of Mexico, locations that exhibit climatologically low mean ascent (Figure 4.1a). Evident in all of these seasonal metrics, February storms brings a strong increase in ascent over the southwestern US and adjacent oceanic waters, and contribute strongly to the seasonal total ascent, especially for coastal California. The strong increase (decrease) in storminess over California (Pacific Northwest) from January to February reverses over the next month.

Ascent is strongly diminished nearly everywhere in the domain during April except for an exceptionally strong and spatially coherent increase in upward motions centered over the Great Basin and extending northeast into the high plains of Wyoming, western South Dakota and Nebraska (Fig. 4.5f). Locally, this strong increase also provides a large contribution to the seasonal total ascent of nearly 20% (Fig. 4.4f). This April peak in synoptic-scale ascent corresponds well with the seasonal intensification in precipitation and cyclone activity in the intermountain west.

A more concise summary of the seasonal cycle in synoptic-scale ascent in terms of the time of year with peak primary ascent at each grid point is provided in Fig. 4.6. The spatial distribution of these seasonal peaks shows remarkably coherent regional structures that agree well with synoptic experience. For example, peak ascent during November-December is evident over British Columbia and the North Pacific as well as over much of the eastern US and Canada. In contrast, California and adjacent regions of the Pacific experience a seasonal peak during February, Alberta experiences a peak during March, and the interior mountain west has the greatest ascent during April. However, it should be noted that the amplitude of the seasonal cycle is particularly weak over Alberta (not shown).

*Interannual Variability of Synoptic-Scale Ascent*

As briefly mentioned in Chapter 2, standardized anomalies for each of the 19 cool seasons were calculated for each grid point by removing the climatological mean and dividing by the year-to-year standard deviation. From the correlation matrix derived from the 19 cool-season standardized anomaly fields, two coherent modes of interannual variability were determined using principal component analysis (Fig. 4.8). The first principal component explains 23% of the interannual variance in ascent. Its spatial pattern reflects strong out-of-phase year-to-year variations between regions equatorward and poleward of the band of climatological mean ascent (Fig. 4.1a). This result is somewhat unusual compared to many other geophysical signals, as it is common for the first principal component to simply be the same sign over much of the domain with its peak amplitude in the center of the domain. The second principal component explains 12% of the interannual variance and exhibits multiple centers of action, the strongest of which is centered near 40oN 130oW. The small number of years in this data set preclude evaluating higher modes.

Interpretation of these two principal components is facilitated by superimposing the Multivariate El Nino Index (MEI) that represents the phase of ENSO (Wolter and Timlin 1998) on the first principal component time series (Fig 4.8a) and superimposing a measure of the Pacific/North American index (Gutzler and Wallace 1981) on the second principal component time series (Fig. 4.8b). The first principal component had its largest positive amplitude during the 1998 El Nino, which reflects a spatial anomaly pattern of enhanced ascent along 30oN over the Pacific Ocean and diminished ascent over the Pacific Northwest, British Columbia and much of the interior of the United States. Alternatively, increased ascent in the latter regions would be expected to have taken place during the 2008 La Nina cool season. Hence, synoptic ascent is shifted southward into a more zonal pattern during positive phases of the first principal component while the sinusoidal nature of the mean synoptic ascent is accentuated during negative phases. Similarly, enhanced synoptic-scale ascent would be expected to have taken place during the 1999 cool season over northern California and adjacent areas of the Pacific Ocean.

It is clear from Fig. 4.8a that there is a strong association between the first principal component of interannual synoptic-scale ascent and ENSO, with a linear correlation of r=0.76 between them. This correlation is quite large (even taking into consideration the limited number, ~7, of degrees of freedom in this sample), and is significant at the 95% level. Hence, this relationship reinforces the known relationships between positive ENSO phases forcing a southward displacement of the storm track along the west coast of the United States combined with diminished synoptic-scale ascent over the Pacific Northwest. For negative phases of ENSO, the opposite holds true, with enhanced synoptic lift in an amplified wave pattern across the North American continent. Composites of the synoptic-scale ascent during the three strongest positive and negative ENSO years within this 19 year sample are presented as well in Figure 4.9. Similarly, positive ENSO phases favor a strengthened and eastwardly elongated Pacific jet stream, while negative ENSO phases are associated with an amplified climatological wave pattern and a more variable eastern Pacific jet stream. The relationship between the second principal component and the PNA index should be considered tenuous, given the small sample size. The two indices share ~36% of their variance in common, which is not significant at the 99% confidence level assuming ~7 degrees of freedom.

DISCUSSION

Storm tracks across western North America were defined in the previous chapter in terms of the climatological mean occurrence and intensity of synoptic-scale ascent. Additional context for these results are presented here based on prior research. In addition, year-to-year variability of the storm tracks is related to variability in precipitation.

While details of the ascent-based methodology used in this work are novel, HH explored 500 hPa omega as well using tracking and variance approaches (see Chapter 1). It is not surprising that a number of their features differ from those presented in the previous chapter, since the approach adopted here focuses on synoptic-scale omega. For example, omega variance exhibits a weak wave in the otherwise zonal storm track whereas the present study suggests an amplified sinusoidal track. The HH negative omega (ascent) tracking statistics, on the other hand, do show a more meridional pattern across western North America with a local maximum in track density along the British Columbia coastline that corresponds well with that seen in Fig. 4.1a. Although omega variance is low across the southwestern United States, in contrast to the strong Arizona maximum evident in Fig. 4.1a, they did find a region of increased track density extending southeast from Nevada. However, no robust connection extended eastward in their study from the southwestern US towards the enhanced track densities found across the eastern half of the continent.

The ascent-based storm tracks exhibit greater correspondence with mid- and upper-tropospheric trough and vorticity climatologies. Despite using different metrics, HK, LNG, and the present study all depict a sinusoidal primary storm track and a weak secondary storm track across the Canadian Rockies. The location of local maxima and minima within the storm track also corresponds well, although both the HK and LNG tracks and maxima appear displaced westward near California.

The similarity of mid-tropospheric vorticity based measures and synoptic-scale ascent is likely a consequence of the role of vorticity in the omega equation:

1 2

In place of the divergence of the Q-vector, the two terms on the RHS represent (1) the vertical differential advection of absolute vorticity and (2) the horizontal Laplacian of temperature advection (both vorticity and temperature are expressed in terms of geopotential, . The Q-vector form shown in Chapter 2 is generally favored because the two RHS terms often tend to cancel one another. However, the RHS can be dominated in some synoptic situations by the vorticity advection term that itself is dominated by vorticity advection aloft. For example, consistent with differences between the location of vorticity centers and vorticity advection, the Arizona ascent maximum found in this study is east of the vorticity maxima of HK, which, in turn, is east of the Eulerian Centripetal Acceleration (ECA) maximum of LNG. Specifically, equivalent barotropic cutoff low pressure systems that are frequently found in the south eastern Pacific (Bell and Bosart 1976(?)) may contribute strongly to the LNG and HK maxima due to strong curvature and slowly moving vorticity centers, yet have less effect on ascent statistics.

Since the frequency of occurrence of events depends to a degree on the thresholds used to define the events, it is not surprising that the HK vorticity features are estimated to occur roughly 20-30% of the time within the storm track in contrast to 12-20% of the time for synoptic-scale ascent. However, the greater occurrence of vorticity features relative to synoptic-scale ascent may also arise from considering the complete RHS of the omega equation. If horizontal vorticity gradients are weak, the vertical derivative term is not strong, or temperature advection is relevant, then the presence of a vorticity center may have reduced impact on rising motions. The present study likely provides a more stringent restriction for assessing the relevance of synoptic-scale storms.

In light of the similarities detailed above and computational ease, it may be tempting to assume that adequate storm tracks could be produced using the advection of relative vorticity at 500 hPa or convergence of the Q-vector. To illustrate limitations for that approach, Fig. 5.1 shows the climatological mean distribution of the convergence of the Q-vector as opposed to the climatology obtained using the fully-integrated omega solution (Fig. 4.1a). Overall, higher mean values of convergence of the Q-vector are located further north than the mean values of omega, as a result of the latitudinal and stability dependence of omega. As discussed in Chapter 2, ascent will be greater in lower latitude regions with weaker stability for a given local source of -𝛻∙Q. Although Elbern et al. (1998) employed a strong positive 𝛻∙Q constraint to identify tropopause folds, his resulting climatology is quite similar to the ascent-based storm track and does not suffer from a northern bias. The requirement of a collocated depression of the dynamic tropopause may help to overcome the sensitivity to latitude.

*Cyclones*

This climatology of mid- and upper-tropospheric synoptic-scale ascent exhibits considerable similarity with surface cyclone climatologies. Three subsets of cyclones, Alberta Clippers, Rocky Mountain cyclones, and Great Basin cyclones all seem to be well represented in the ascent based storm track.

Thomas and Martin (20XX) provided a comprehensive climatology of the Alberta Clipper, demonstrating both the archetypal storm evolution (including inferred vertical motion) and the preferred tracks of these systems. Clippers tend to be weak systems originating from upstream Pacific vorticity maxima that subsequently traverse the mean ridge and Canadian Rockies. The common feature, of course, is the spin up of low level vorticity in the lee trough region and then the disturbance amplifies vertically and moves with westward tilt and stronger vorticity observed at 500 hPa later in its life cycle. Accordingly there is also an increasing Div-Q signal aloft as the systems progress east. This evolution corresponds very well with the details of the mean ascent in Figure 4.1a. Subjective analysis of seasonal animations of ascent suggests many of the features that contribute to the northern track propagate from the Gulf of Alaska, east across the Canadian Rockies, and then across the high plains of Canada consistent with the Alberta Clipper morphology. Additional contributions to this portion of the storm track from more northern sources, as well as retrogressing features related to deep circulations near the Great Lakes, are also observed.

Rocky Mountain cyclones, which have a strong genesis maximum in the lee of the Rocky Mountains near the southeastern corner of Colorado (Klein et al. 1968, Rietan et al. 1972, Zishka and Smith 1980, HH, JS) also appear to be present in the ascent-based storm track. Local maximum C, in Fig. 4.1a, is found just downstream of the cyclone density maximum which is suggestive of similar dynamics to those observed with the Alberta Clipper. In general, ascent maxima are likely to be found downstream of cyclogenesis maxima due to the timescales required for development as well as the tendency for ascent to be displaced to the east of the trough axis or cyclone center.

Great Basin cyclones have been found to occur most frequently in a region anchored on the lee side of the high Sierra Nevada mountains and extending northeast across Nevada towards the Great Salt Lake Basin in Utah (JS). Examining the fetch of high mean ascent and ascent frequency upstream of maximum B in Fig. 4.1a, it is apparent that these regions of cyclone occurrence are found beneath regions of high climatological ascent. Furthermore, JS finds a dramatic increase in cyclone occurrence for the Great Basin province during the month of April, the same month for which ascent maximizes across the region. These results are further supported by Whittaker and Horn (1981), who showed a maximum in great basin cyclogenesis in April.

*Arizona Ascent Maximum*

The Arizona maximum in mean synoptic ascent, maximum B in Fig 4.1a, may strike some readers as a strange or even erroneous feature of this study, since it coincides with a region of limited precipitation and clouds. However, the occurrence of precipitation requires moisture as well as lift. Rose and Lin (2003) documented a very low temporal correlation (on what time scale; interannual?) between precipitation rate and 700 hPa vertical motions for the southwestern US. Compositing all 6-h periods with ascent centered on maximum B, Fig. 5.2a shows that the core of rising motion coincides with relative humidity in the range of 40%. This low relative humidity coincident with the rising motion differs markedly from systems elsewhere in the domain, such as the Pacific Northwest (Fig. 5.2b), where relative humidity near 60% is typical throughout regions of ascent. For reference, the mean relative humidity at 600 hPa during ascent events is shown in Fig. 5.3b relative to the climatological mean relative humidity at that level during all 19 cool seasons (Fig. 5.3a). (I don’t understand what Fig. 5.3c is without the caption). More northern and elevated locations are much more likely to possess air near saturation in concert with ascent, thus making precipitation a more likely outcome (Fig. 5.3c). This result is hardly surprising as ascent is arguably the primary mode by which air reaches saturation. While the relationship between ascent and increased relative humidity is also observed for the southwestern US, the climatological base state is sufficiently dry that saturation is harder to attain.

While relative humidity is an important consideration for the initiation of precipitation processes, column integrated total water (the sum of water vapor and cloud condensate) may modulate the amount of precipitation that is possible. The climatology and storm related anomalies in TW are documented in Fig. 5.4. Interestingly, some regions within the primary storm track experience reductions, or neutral anomalies, in the presence of ascent. To the contrary, southern portions of the domain tend to see significant increases in total water.

*Precipitation Variability*

Appreciating that ascent by itself does not lead to precipitation, it is still intriguing to explore the links between interannual variations in synoptic-scale ascent and the variability in regional precipitation. Regressing the PC1 time series of mean ascent (red line in Fig 4.8a) with the time series of PRISM precipitation standardized anomalies at each grid point yields a now familiar north-south dipole pattern of ENSO-related precipitation variability for the western US (Fig. 5.5a). It should be remembered that estimating as few as 7 independent events, the correlations between seasonal anomalies of ascent and precipitation should be viewed with some caution. Positive (negative) correlations are found for the arid southwest (Pacific Northwest). Interesting local exceptions to this pattern exist, dominated by the tendency for low elevation arid (high elevation moist) regions to be positively (negatively) correlated with the PC1 time series. For the eastern portions of the US, positive correlations are found across the southeast, with negative correlations across the upper Mississippi and Ohio River basins. A similar pattern of precipitation anomalies arises as the third/first (I don’t understand what this means) depending EOF (EOFprecip1) of precipitation variations (Fig. 5.5b). Taking into consideration the limitations imposed from the small sample size the associated principal component time series for EOFprecip1 has a weaker correlation (r~=-.62) with the MEI than does the related EOF1 of synoptic ascent (r~= -.76), which may suggest that there is a more direct ENSO control on storm track than on the amount of precipitation. This disparity may again have to do with available moisture modulating the productivity of synoptic-scale ascent.

CONCLUSIONS

This study examines the position and variability of storm tracks across western North America. A dynamically filtered and physically relevant method was introduced wherein the climatological distribution of ascent was shown to be representative of storm tracks. An alternative balance version of the omega equation was used to diagnose the component of ascent associated with synoptic-scale motions while dynamically filtering poorly-resolved mesoscale contributions, such as terrain forced flows. The maximum values of ascent among 5 mid-tropospheric levels were retained at each 6-hour interval over 19 cool seasons. Animations of the maps of ascent were examined subjectively during all 19 seasons and a case study was used to evaluate the sensitivity of the results to the technique. Summary ascent statistics over months, seasons, and the entire 19-year period yield a measure of the propensity of a location to experience synoptic-scale disturbances. Locations exhibiting high mean ascent and ascent frequency are thus said to be storm tracks.

The climatological mean ascent based storm track was found to occupy a sinusoidal belt across western North America. Variations in the position of the storm track were shown to be related to the phase of ENSO, with El Nino (La Nina) winters favoring a more zonal (amplified) and southern (northern) storm track. The relationship between variations in ascent and those in precipitation were observed to be modulated by available moisture and the saturated state of the atmosphere. The ascent based storm tracks were also shown to be consistent with elements of previous lower-, mid-, and upper-tropospheric storm track studies.

Remaining research questions pertain to the large scale dynamics that contribute to the observed structure of the storm track. In particular, the strong southward displacement of high mean ascent along the US west coast is a peculiar feature as is the ascent minima proximal to Montana. Additionally, the tendency for storms that are displaced south from the mean pathway to exhibit stronger rising motion poses questions about the structure and dynamics of these systems. Lastly, the ‘tax day storm’ case examined herein, suggests that idealized modeling studies examining the phase locking between upper level PV anomalies and surface thermal waves may yield insights into the evolution of storms in regions of complex topography.

REFERENCES

Bell, G. D., L. F. Bosart, 1989: A 15-year climatology of Northern Hemisphere 500 mb closed cyclone and anticyclone centers. *Mon. Wea. Rev.*, **117,** 2142-2164.

Blackmon, M. L., 1976: A climatological spectral study of the 500 mb geopotential height of the Northern Hemisphere. *J .Atmos. Sci.,***33**,1607-1623.

Clough, S. A., C. S. A. Davitt, and A. J. Thorpe, 1996: Attribution concepts applied to the omega equation. *Quart. J. Roy. Meteor. Soc.,***122,**1943-1962.

Davies-Jones, R. P., 1991: The frontogenetical forcing of secondary circulations. Part I: The duality and generalization of the Q vector. *J. Atmos. Sci.,* **48,** 497-509.

——, 2009: The frontogenetical forcing of secondary circulations. Part II: Properties of Q vectors in exact linear solutions. *J. Atmos. Sci.,* **66,** 244-260.

Dean, D. B.,and L. F. Bosart, 1996: Northern Hemisphere 500-hPa trough merger and fracture: A Climatology and Case Study. *Mon. Wea. Rev.,* **124,** 2644-2670.

Dee, D. P., and S. Uppala, 2009: Variational bias correction in ERA-Interim. *ECMWF Newsletter* *No. 119*.

Dee, D. P., P. Berrisford, P. Poli, and M. Fuentes, 2009: ERA-interim for climate monitoring. *ECMWF Newsletter* *No. 119*.

——, 2009:Variational bias correction of satellite radiance data in the ERA-Interim reanalysis. *Quart. J. Roy. Meteor. Soc.,***135,** 1830-1841.

Durran, D. R., and L. W. Snellman, 1987: The diagnosis of synoptic-scale vertical motion in an operational environment. *Wea. Forecasting,* **2,** 17-31.

Elbern, H., J. Hendricks, and A. Ebel, 1998: A climatology of tropopause folds by global analyses. *Theor. Appl. Climatol.,* **59,** 181-200.

Hakim, J. G., 2000: Climatology of coherent structures on the extratropical tropopause. *Mon. Wea. Rev.,* **128,** 385-406.

Hoskins, B. J., I. Draghici, and H. C. Davies, 1978: A new look at the 𝜔-equation. *Quart. J. Roy. Meteor. Soc.,***104,** 31-38.

Hoskins, B. J., and M. A. Pedder, 1980: The diagnosis of middle latitude synoptic development. *Quart. J. Roy. Meteor. Soc.,***106,** 707-719.

Hoskins, B. J., M. E. McIntyre, and A. W. Robertson, 1985: On the use and significance of isentropic potential vorticity maps. *Quart. J. Roy. Meteor. Soc.,* **111,** 877-946.

Hoskins, B. J., and K. I. Hodges, 2002: New perspectives on Northern Hemisphere winter storm tracks. *J. Atmos. Sci.,* **59,** 1041-1061.

Jeglum, M. E., and W. J. Steenburgh, 2010: Multi-reanalysis climatology of intermountain cyclones. *Mon. Wea. Rev.* ***??????????????????***

Klein, W. H., D. L. Jorgensen, and A. F. Korte, 1968: Relation between upper air lows and winter precipitation in the western plateau states. *Mon. Wea. Rev.,* **96,** 162-168.

Keyser, D., M. J. Reeder, and R. J. Reed, 1988: A Generalization of Petterson’s frontogenesis function and its relation to the forcing of vertical motion. *Mon. Wea. Rev.,* **116,**762-780.

Keyser, D., B. D. Schmidt, D. G. Duffy, 1992: Quasigeostrophic vertical motions diagnosed from along- and cross-isentropic components of the Q vector. *Mon. Wea. Rev.,* **120,** 731-741.

Lefevre, R. J., and J. W. Nielsen-Gammon, 1995: An objective climatology of mobile troughs in the Northern Hemisphere. *Tellus,* **47A,** 638-655***.***

Martin, J. E., 2007: Lower-tropospheric height tendencies associated with the shearwise and transverse components of the quasigeostrophic vertical motion. *Mon. Wea. Rev.,* **135,** 2803-2809.

Myoung, B., and Y. Deng, 2009: Interannual variability of the cyclonic activity along the U.S. Pacific coast: Influences on the characteristics of winter precipitation in the western United States. *J. Climate,* **22,**5732-5747.

Reitan, C. H., 1974: Frequencies of cyclones and cyclogenesis for North America, 1951-1970. *Mon. Wea. Rev.,* **102,** 861-868.

Rose, B. E. J., and C. A. Lin, 2003: Precipitation from vertical motion: A statistical diagnostic scheme. *Int. J. Climatol.,* **23,** 903-919.

Sanders, F., 1988: Life history of mobile troughs in the upper westerlies.  *Mon. Wea. Rev.,*  **116,** 2629-2648.

Serreze, M. C., M. P. Clark, R. L. Armstrong, D. A. McGinnis, and R. S. Puwarty, 1996: Characteristics of the western U.S.snowpack from SNOTEL data. *Water Resour. Res.,* **35,** 2145-2160.

Simmons, A., S. Uppala, D. Dee, and S. Kobayashi, 2007: ERA-interim: New ECMWF reanalysis products from 1989 onwards. *ECMWF Newsletter No. 110.*

Simmons, A., S. Uppala, and D. Dee, 2007: ERA-interim: New ECMWF reanalysis products from 1989 onwards. *ECMWF Newsletter No. 111.*

Stuart, D. W., 1967: The over-relaxation factor in the numerical solution to the omega equation. *Mon. Wea. Rev.,* **95,** 303-307.

Sutcliffe, R. C., 1947: A contribution to the problem of development. *Quart. J. Roy. Meteor. Soc.,* **73,** 370-383.

Thomas, B. C., and J. E. Martin, 2007: A synoptic climatology and composite analysis of the Alberta Clipper. *Wea. Forecasting,* **22,** 315-333.

Treberth, K. E., 1978: On the interpretation of the diagnostic quasi-geostrophci omega equation. *Mon. Wea. Rev.,* **106,** 131-137

Uppala, S., D. Dick, S. Kobayashi, P. Berrisford, and A. Simmons, 2008: Towards a climate data assimilation system: status update of ERA-Interim. *ECMWF Newsletter No. 115.*

Wernli, H., and C. Schwierz, 2006: Surface cyclones in the ERA-40 dataset (1958-2001). Part 1: Novel identification method and global climatology. *J. Atmos. Sci.,* **63,**2486-2507.

West, G. L., and W. J. Steenburgh, 2010: Life cycle and mesoscale frontal structure of an intermountain cyclone. *Mon. Wea. Rev.* ***??????????????????????***

Wolter, K., and M. S. Timlin, 1998: Measuring the strength of ENSO events - how does 1997/98 rank? *Weather,* **53,** 315-324.

Zishka, K. M., and P. J. Smith, 1980: The climatology of cyclones and anticyclones over North America and surrounding ocean environs for January and July, 1950-77. *Mon. Wea. Rev.,* **108,** 387-401.