SYNOPTIC AND MESOSCALE STRUCTURE OF A

WASATCH MOUNTAIN WINTER STORM

by

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ABSTRACT

An observational analysis of a frontal cyclone over the western United States was conducted using data collected during IOP-3 of the Intermountain Precipitation Experiment (IPEX). The analysis is based on conventional data, special radiosonde observations, high density surface observations collected by the MesoWest cooperative networks, NEXRAD radar data, wind profiler observations, and flight-level data from the NOAA P-3 aircraft. A complex frontal cyclone evolution was observed, including cold frontal distortion by the Sierra Nevada, and development of a lee trough and mesoscale cyclonic vortex over Nevada. As the system moved across the Intermountain region, a discontinuous progression of surface low pressure, and significant modification of the surface trough was observed. By the time the system reached northern Utah, mid and low-levels were decoupled, which was different from the coupled structure observed upstream in California.

The storm environment over northern Utah was examined in detail, focusing on factors important for cloud and precipitation development. Three storm stages were identified according to the transient synoptic and mesoscale features that were responsible for providing thermodynamic and kinematic conditions conducive for cloud and precipitation development. During each stage, mesoscale flow interactions with the unique topography of northern Utah greatly affected precipitation development and distribution. The importance of these mesoscale flow interactions for precipitation development, the
possible mechanisms for the complex vertical structure evolution, and the relative role of
topography is discussed.
CHAPTER 1

INTRODUCTION

The complex terrain of the western United States has a considerable effect on the structure and evolution of frontal cyclones and their attendant precipitation systems. Orographic effects resulting from the juxtaposition of synoptic flows over complex terrain affect the movement, formation and dissipation of cyclones and fronts at various scales. Documenting and understanding these effects is critical for accurate analysis and forecasting (e.g., Hill 1993), as well as validating numerical weather prediction models. Due to the limited availability of high density surface, upper-air, and radar observations, such interactions are poorly understood over the western United States, especially the Intermountain region. Observations collected by the MesoWest cooperative networks (Horel et al. 2002) and the Intermountain Precipitation Experiment (IPEX), however, provide a unique opportunity to examine frontal cyclone and precipitation system evolution over the western United States. IPEX was designed to improve the understanding of precipitation and precipitation processes over the Intermountain region, and was conducted during February 2000 when seven Intensive Observing Periods (IOPs) were held; additional information about IPEX is provided by Schultz et al. (2002). This study presents an observational analysis of IPEX IOP3, during which complex cyclone evolution and significant precipitation was observed over Utah and adjoining states.

The terrain of the western United States is quite complex. The Intermountain
region encompasses much of the western United States, spanning from the Sierra and Cascade ranges to the Continental Divide (Fig. 1.1). Much of the Intermountain region is characterized by basin and range topography where narrow, steeply sloped mountain ranges are separated by broad lowland valleys and basins. For example, northern Utah, where IPEX was focused, features the meridionally-oriented Stansbury, Oquirrh, and Wasatch Mountains, which rise 1500 – 2000 m above the surrounding lowlands (Fig. 1.2). Other significant topographic features include the Salt Lake and Tooele Valleys, the Great Salt Lake, and the Great Salt Lake Desert.

Cyclogenesis occurs frequently over the Intermountain region, especially downstream of the Sierra Nevada during late winter and early spring (e.g., Petterson 1956; Zishka and Smith 1980; Lee 1995). Nevertheless, only a few studies have examined Intermountain cyclone evolution. Based on a climatology of 237 events, Lee (1995) noted that most (75%) Intermountain cyclones develop during large-scale southwesterly flow over the Sierra Nevada, while the remaining events form under northwest large-scale flow. Based on these cases, Lee (1995) described Intermountain cyclogenesis as a two-stage process. First, the wind field adjusts due to the synoptic flow interacting with terrain, which forms a lee trough in southwest flow cases. Second, the cyclone deepens in situ as the large-scale forcing moves overhead.

One of the most significant effects of orography on mobile cyclones is to alter their vertical tilt (e.g., Buzzi and Tosi 1989). Interactions between transient upper-level disturbances and orographically-induced circulations may affect the vertical tilt. Little research has examined the vertical evolution across the Intermountain region. Lee (1995) noted decoupling between upper-level transient disturbances and Intermountain cyclones,
Figure 1.1. Topography and major geographic features of the western United States. Elevation (m) based on scale at lower left. Boxed area over northern Utah identifies location of Fig. 1.2.
Figure 1.2. Topography and major geographic features of northern Utah. Elevation (m) based on scale at upper left.
and comparison of time-height data from Reynolds and Kuciauskas (1988) and Long et al. (1990) revealed a significant change in vertical structure. As described by Reynolds and Kuciauskas (1988), a cold front was coupled with a mid and upper-level cold surge/humidity front over the Sierra Nevada, but by the time the storm reached the Tuschar Mountains of southern Utah, Long et al. (1990) observed a cold surge/humidity front approximately 7 h ahead of the surface cold front.

Several studies have investigated frontal evolution over the Western United States. Reynolds and Kuciauskas (1988) suggested that a southeastward moving cold front was distorted by the Sierra Nevada, but their subjective analysis was based on limited surface observations. Horel and Gibson (1994) examined a complex interaction between two cold fronts over the Intermountain region, but their analysis relied on limited surface observations and numerical model output. Hoffman (1995) analyzed the discontinuous progression of a cold front from California into Nevada, and proposed that frontal motion and development was controlled by a combination of upper-level forcing, topographic interactions, and diabatic boundary layer processes. Sanders (1999a) showed, for a cold front in the southwest United States, that frontogenesis and lysis were driven largely by diabatic processes related to the diurnal insolation cycle. Using more dense MesoWest observations, Blazek (2000) and Steenburgh and Blazek (2001) documented three major mesoscale topographic effects on a cold front as it traversed the western United States. First, the southeast moving cold front was distorted and split by the Sierra Nevada. Second, multiple mesoscale cyclonic circulations formed along the surface front across northern Utah. Third, topographic effects in the Snake River Plain of southern Idaho channelled and accelerated the cold front.
Much of the research on wintertime precipitation systems over the western United States has focused on the feasibility of cloud seeding to augment precipitation (e.g., Marwitz 1980; Marwitz 1987; Reynolds and Kuciauskas 1988; Long et al. 1990; Klimowski et al. 1998). Since the premise of these studies was to decipher the feasibility of cloud seeding, emphasis was placed on determining how the evolution of stability, mesoscale kinematics, and microphysics affected orographic precipitation development. Marwitz (1980) summarized this evolution based on 12 winter storms in the San Juan Mountains of Colorado. The storms generally evolved from stable to less stable regimes, with four storm stages identified as stable, neutral, unstable, and dissipation.

Other studies have examined aspects of wintertime storms over northern Utah. Astling (1984) found a high frequency of cool-season precipitation events at SLC during the early morning which were coincident with peak values in convergence of thermally-driven flows. This may be partly explained by Steenburgh et al. (2000) and Steenburgh and Onton (2001) who showed that Great Salt Lake-induced thermally-driven flows were more important than orographic processes in some northern Utah snowstorms. Williams and Heck (1972) illustrated that the scale of precipitation events over northern Utah is typically smaller than that found over the eastern United States. Dunn (1983) explained precipitation distribution over northern Utah by relating the orientation of topography to wind directions. Horel and Gibson (1994) showed, for one case, three factors important for precipitation development: synoptic ascent, orographic lift, and lifting associated with a shallow cold front. Slemmer (1998) compiled radar climatologies from KMTX during January – March 1997, and conducted detailed snow depth measurements of 13 snowstorms, finding significant spatial precipitation gradients within events. Cheng (2001)
compared observed precipitation to quantitative precipitation (QPF) model forecasts for all seven IPEX IOPs, showing substantial spatial gradients both within and between each IOP. Such precipitation gradients illustrate the importance of mesoscale terrain-induced processes for precipitation development and distribution across the complex terrain of northern Utah. Most recently, Cox (2002) showed, with Dual-Doppler radar analysis during IPEX IOP3, complex flow interactions near the northern Wasatch barrier on the meso-gamma scale (2 – 20km), including terrain-induced convergence and divergence, thereby illustrating the importance of local orographic interactions.

Mesoscale, terrain-induced effects have also been shown to enhance precipitation in lowland areas near the topography producing them. Mass (1981) showed how deflected flow around the Olympic Mountains converged in the lee to form the Puget Sound Convergence Zone, while Wesley et al. (1995) showed how anticyclonic deflection of the low-level flow in northern Colorado can generate postfrontal convergence. Andretta and Hazen (1998) illustrated how flow up the Snake River Plain converged with down canyon flows in the upper Snake River Plain, and Gheusi et al. (2002) showed how the concave shape of the Alps deflected low-level flow, producing convergence. In all these cases, the terrain-induced convergence acted to generate or augment precipitation.

IPEX provided a detailed data set to explore cyclone evolution and the mesoscale characteristics of a major winter storm across the Intermountain region. The observational analysis of IPEX IOP3 presented in this thesis examines several important questions. How do transient synoptic and mesoscale systems impact the spatial distribution of precipitation? How does the structure of mobile, frontal cyclones change over the Intermountain region? How important is the role of topography in modifying the structure
and scale of precipitation systems? With these questions in mind, the main objectives of this thesis are:

• To document the synoptic evolution of the IOP3 storm system across the western United States.
• To characterize the synoptic and mesoscale storm environment across northern Utah, including the evolution of stability, wind, and moisture associated with synoptic and mesoscale features.
• To understand the relationship between the storm environment and temporal and spatial variations in precipitation across northern Utah.
• To demonstrate the relative importance of topography on the precipitation distribution across northern Utah.

The remainder of this thesis is organized as follows. Chapter 2 describes the observational data sources used and methods employed for subjective analyses. Chapter 3 presents the synoptic evolution. Chapter 4 details the mesoscale evolution and relationship to observed precipitation, focusing on northern Utah. Summary and conclusions are offered in Chapter 5.
Manual surface analyses were based on high density observations that were collected by the MesoWest cooperative networks (Horel et al. 2002) from approximately 2500 stations in the western U.S. and 250 stations across northern Utah. MesoWest observations were quality controlled by comparing the observed station values to an estimate based on multivariate linear regression (Splitt and Horel 1998), and by spatial and temporal consistency checks performed while preparing the manual analyses. Observations deemed questionable by either approach were not considered in the analysis.

Manual surface analyses were prepared in the following manner. Following Steenburgh and Blazek (2001), 1500-m pressure was used rather than sea level pressure since 1500 m is near the mean elevation of Intermountain observing sites. Surface fronts were identified according to the traditional definition of a front as an elongated area of strong horizontal temperature gradient (e.g., Bluestein 1986; Keyser 1986), and inspection and integration of all available MesoWest observations with the constraints of producing an analysis that was temporally, horizontally, and vertically continuous. In this case, a sufficient temperature gradient for a cold front was a temperature drop of at least 3°C immediately after cold frontal passage. No warm fronts were analyzed for this case, and in general, the surface feature was analyzed as a trough and not a cold front. Although not presented, subjective analyses of potential temperature were also used to help determine
the existence of fronts in complex terrain, as suggested by Sanders (1999b).

Gridded data used for plan-view upper-air analyses were provided by the National Centers for Environmental Prediction Rapid Update Cycle (RUC2; Benjamin et al. 1994). Originally featuring a horizontal grid spacing of 40 km, the RUC2 was downloaded and archived at the University of Utah on a 60-km horizontal and 25-hPa vertical grid. MesoWest observations and RUC2 gridded data were plotted using GEneral Meteorological PAcKage (GEMPAK) software.

Upper-air observations used in the analysis included conventional 12-h and special 3- or 6-h observations collected by Intermountain National Weather Service (NWS) upper-air sites. Over northern Utah, 3-h upper-air soundings were available from SLC and two National Severe Storm Laboratory mobile labs, NSSL-4 and NSSL-5, which were located ~75 km and ~15 km upstream of the Wasatch Mountains, respectively (see Fig. 1.2 for locations). In addition, 915 MHz wind profiler data was available from Dugway Proving Grounds (DPG), about 150 km southeast of SLC, and at several sites in California that were operated by the NOAA Environmental Technology Laboratory (ETL). Flight-level data from the NOAA P-3 research aircraft and Velocity Azimuth Display (VAD) winds from several NEXRAD sites were also used, but not presented.

Composite Weather Surveillance Radar-1988 Doppler (WSR-88D) radar analyses were created from data that were received and archived in Next-Generation Radar (NEXRAD) Information Dissemination Service format (Baer 1991), which has a spatial resolution of 8 km and an approximate reflectivity resolution of 5 dBZ. Due to problems with radar use in complex terrain and poor spatial coverage, many areas of the western United States such as central Utah and Nevada are not well covered by the NEXRAD
network (e.g., Westrick et al. 1999). Over northern Utah, WSR-88D radar analyses from Promontory Point (KMTX) were based on NIDS-format data with a spatial resolution of 1 km and an approximate reflectivity resolution of 5 dBZ.
At 0000 UTC 12 February 2000, a well-developed midlatitude cyclone was centered just off the west coast of the United States (Fig. 3.1) with the concomitant cloud shield covering much of the western United States (Fig. 3.2). Ahead of the surface cyclone and attendant cold front, widespread precipitation was observed across much of northern and central California (Fig. 3.2). At upper levels, split 500-hPa flow was present off the United States west coast where two upper-level shortwave troughs, one off the California coast, the other off the Pacific Northwest coast, had phased to produce a deeper upper-level trough (Fig. 3.2a). With southwest, cross-barrier, crest-level flow at 700-hPa ahead of this trough (Fig. 3.2b), an elongated lee trough existed east of the Sierra Nevada (Fig. 3.1).

Baroclinity associated with the front was weak, with most stations observing a 2–3°C drop as the cold front moved inland. One the strongest frontal passages was at Santa Rosa (STS) around 0000 UTC, where the cold front was accompanied by a brief temperature spike followed by a rapid 4°C fall, temporary shift of surface winds from south-southeast to south, and a pressure minimum (Fig. 3.3). Observations from a NOAA/ETL wind profiler located near STS at Richmond (RMD) suggested that the wind shift accompanying the front sloped rearward with height (Fig. 3.4a). The vertical structure of the cold front at RMD resembled that of classic cold front with a density current-like nose.
Figure 3.1. Manually analyzed surface map for 0100 UTC 12 February 2000. 1500-m pressure isobars every 2 hPa. Surface station reports include 1500-m pressure (tenths of hPa with leading 8 truncated), and winds (full (half) barbs denote 5 (2.5) m s$^{-1}$). [Note: 0100 UTC data used because of data acquisition problems at 0000 UTC.]
Figure 3.2. Upper-level, satellite, and radar analyses at 0000 UTC 12 February 2000. (a) RUC2 analysis of 500-hPa geopotential height (every 60 m) and absolute vorticity (x10^{-5} s^{-1}, shaded following scale at bottom). (b) RUC2 analysis of 700-hPa temperature (every 2 °C), wind (full and half barb denote 5 and 2.5 m s^{-1}, respectively), and relative humidity (%), shaded following scale at bottom). (c) Infrared satellite image. (d) Composite NEXRAD imagery.
Figure 3.3. Meteograms of temperature (red), dewpoint (green), relative humidity (blue), wind speed (solid), gust (dashed), direction (circles), and altimeter setting (brown) at Santa Rosa (STS) from 0000 UTC 11 February–2345 UTC 12 February 2000; see Fig. 3.1 for location. Dashed line denotes cold frontal passage.
Figure 3.4. Wind profiler data for 12 February 2000 from (a) Richmond (RMD) and (b) Sacramento (SAC). Wind (full (half) barbs denote 5 (2.5) m s$^{-1}$). Y-axis is meters above ground level and hour UTC is given on x-axis. Solid lines denote wind shift accompanying frontal passage. See Fig. 3.1 for locations.
and rearward-sloping structure (e.g., Bond and Shapiro 1991). By the time the surface cold front reached Sacramento (SAC) around 0400 UTC, it was marked by a 3°C temperature fall, slight wind shift, and pressure minimum (not shown). NOAA/ETL wind profiler observations suggested that the wind shift accompanying the front at SAC sloped rearward, and the initial depth of the front was shallow (Fig. 3.4b). Winds shifted only from south to southwest at SAC with frontal passage possibly due to the orientation of the front approaching from the southwest and effects of strong, channeled prefrontal winds in the Sacramento Valley (not shown).

By 0600 UTC, the equatorward 500-hPa vorticity maximum was centered over central California (Fig. 3.5a), and strong cyclonic curvature associated with the amplifying upper-level trough was evident in satellite imagery (Fig. 3.5b). The 1500-m low center was located over southwest Oregon and the surface cold front was approaching the windward slopes of the Sierra Nevada as the lee pressure trough remained stationary (Fig. 3.6). Meanwhile, independent of the lee trough, and associated with mid-tropospheric ascent induced by differential vorticity advection (Fig. 3.5a), a mesoscale cyclonic vortex or incipient lee cyclone developed over central Nevada (Fig. 3.7). This circulation was distinguishable only in the low-level winds and remained stationary for approximately 10 h (0100 UTC–1100 UTC 12 February 2000) before dissipating as the upper-level trough and remnants of the Pacific cold front approached. This circulation did not develop along an existing baroclinic zone or develop new fronts, and may be partly an artifact of local terrain channeling effects. Nonetheless, the circulation represented an area of enhanced low-level vorticity. The circulation may have dissipated due lack of baroclinity, small circulation diameter, and other topographic effects, but sparse data
Figure 3.5. Upper-level and satellite analyses at 0600 UTC 12 February 2000. (a) RUC2 analysis of 500-hPa geopotential height (every 60 m) and absolute vorticity ($\times 10^{-5}$ s$^{-1}$, shaded following scale at bottom). (b) Infrared satellite image.
Figure 3.6. Manually analyzed surface map for 0600 UTC 12 February 2000. 1500-m pressure isobars every 2 hPa. Surface station reports include 1500-m pressure (tenths of hPa with leading 8 truncated), and winds [full (half) barbs denote 5 (2.5) m s$^{-1}$]. Boxed area identifies location of Fig. 3.7.
Figure 3.7.  Low-level manual streamline analysis for 0600 UTC 12 February 2000. Surface station reports include winds [full (half) barbs denote 5 (2.5) m s$^{-1}$].
prohibited conclusive results.

From 0600–0900 UTC, the cold front weakened and distorted as it was retarded by the Sierra Nevada, but accelerated northward through the Sacramento Valley and southward through the San Joaquin Valley (Fig. 3.8). As the cold front moved over and to the lee of the northern and central Sierra Nevada, its thermodynamic and kinematic structure became less coherent. At Reno (RNO), a minimum in pressure was observed around 0800 UTC, with a gradual temperature fall soon thereafter, but pressure did not rise and winds did not shift notably until 1200 UTC (Fig. 3.9). Further to the south, where the Sierra Nevada crest is highest, the front did not appear to surmount the terrain. Instead, the front split with the southern portion moving southward and eastward into southern California where it dissipated.

By 1200 UTC 12 February, the primary low center was located in central Oregon and the remnants of the low-level cold front had moved into the Sierra Nevada lee trough which, over northern Nevada, had begun to progress eastward (Fig. 3.10). The 500-hPa absolute vorticity maximum in the southern branch of the jet was located over central California, with 500-hPa cyclonic vorticity advection found over and to the lee of the Sierra Nevada (Fig. 3.11a). The leading edge of cold air and cold advection at 700-hPa was located over eastern Nevada, well in advance of the surface pressure trough (Figs. 3.10 and 3.11b). A large area of cloud cover was present over most of the Intermountain region (Fig. 3.11c). Widespread radar echoes and precipitation were observed over much of Utah and eastern Nevada (Fig. 3.11d); echo-free regions in eastern Nevada and Utah are not covered by the NEXRAD network.

The surface pressure trough moved rapidly across Nevada between 1200–1500
Figure 3.8. Synoptic evolution of surface features (conventional symbols used) from 0000 UTC–2100 UTC 12 February 2000. Labels represent hour UTC 12 February 2000.
Figure 3.9. Meteograms of temperature (red), dewpoint (green), relative humidity (blue), wind speed (solid), gust (dashed), direction (circles), and altimeter setting (brown) at Reno (RNO) from 0000 UTC 12 February–2345 UTC 12 February 2000; see Fig. 3.6 for location.
Figure 3.10. Manually analyzed surface map for 1200 UTC 12 February 2000. 1500-m pressure isobars every 2 hPa. Surface station reports include 1500-m pressure (tenths of hPa with leading 8 truncated), and winds [full (half) barbs denote 5 (2.5) m s\(^{-1}\)].
Figure 3.11. Upper-level, satellite, and radar analyses at 1200 UTC 12 February 2000. (a) RUC2 analysis of 500-hPa geopotential height (every 60 m) and absolute vorticity (x10^{-5} s^{-1}, shaded following scale at bottom). (b) RUC2 analysis of 700-hPa temperature (every 2 °C), wind (full and half barb denote 5 and 2.5 m s^{-1}, respectively), and relative humidity (%), shaded following scale at bottom). (c) Infrared satellite image. (d) Composite NEXRAD imagery.
UTC, reaching Lovelock (LOL) around 1300 UTC and Elko (EKO) around 1500 UTC (Fig. 3.12). During this period, the mean speed of the surface pressure trough varied from 20–25 m s\(^{-1}\), substantially faster than the near-surface flow. Therefore, the movement of the surface trough across Nevada did not appear to be controlled by near-surface advective processes. Instead, its movement was better correlated with upper-level vorticity advection immediately to the east of the upper-level trough axis, similar to that observed in other events (e.g., Hess and Wagner 1948; Schultz and Doswell 2000).

By 1800 UTC 12 February, the 500-hPa trough axis was strongly negatively tilted with the southern branch vorticity maximum centered over southern Nevada (Fig. 3.13a). The surface low was centered over the Idaho–Oregon border with the attendant surface pressure trough extending southeastward through northeast Nevada and southward along the Utah–Nevada border (Fig. 3.14), and a second 1500-m pressure minimum was present in eastern Idaho. Widespread clouds and precipitation were observed ahead of the surface pressure trough, with scattered clouds precipitation in its wake (Figs. 3.13c and 3.13d). Baroclinicity at mid-levels over most of the Great Basin was weak. The strongest 700-hPa cold advection was found over and just east of the continental divide of Colorado and New Mexico, whereas very weak cold advection was evident over Nevada (Fig. 3.13).

By 0000 UTC February 13, the 500-hPa absolute vorticity maximum had weakened and moved over the four corners region (Fig. 3.15a) and both the mid level (Fig. 3.15b) and surface troughs (Fig. 3.16) had moved across northern Utah. An elongated region of low 1500-m pressure, with two embedded pressure minima, extended from Idaho into Wyoming. Nonetheless, clouds and precipitation continued along the Wasatch Mountains and northern Utah (Figs. 3.15c and 3.15d), culminating in the passage of a
Figure 3.12. Altimeter pressure time series from 0000 UTC 12 February to 0000 UTC 13 February. (a) Lovelock (LOL). (b) Elko (EKO). See Fig. 3.10 for locations. Dashed line denotes minimum pressure or surface trough passage.
Figure 3.13. Upper-level, satellite, and radar analyses at 1800 UTC 12 February 2000. (a) RUC2 analysis of 500-hPa geopotential height (every 60 m) and absolute vorticity ($\times 10^{-5}$ s$^{-1}$, shaded following scale at bottom). (b) RUC2 analysis of 700-hPa temperature (every 2 °C), wind (full and half barb denote 5 and 2.5 m s$^{-1}$, respectively), and relative humidity (% , shaded following scale at bottom). (c) Infrared satellite image. (d) Composite NEXRAD imagery.
Figure 3.14. Manually analyzed surface map for 1800 UTC 12 February 2000. 1500-m pressure isobars every 2 hPa. Surface station reports include 1500-m pressure (tenthsofhPa with leading 8 truncated), and winds [full (half) barbs denote 5 (2.5) m s$^{-1}$]. Primary (secondary) low pressure noted with larger (smaller) font size.
Figure 3.15. Upper-level, satellite, and radar analyses at 0000 UTC 13 February 2000. (a) RUC2 analysis of 500-hPa geopotential height (every 60 m) and absolute vorticity (x10^{-5} s^{-1}, shaded following scale at bottom). (b) RUC2 analysis of 700-hPa temperature (every 2 °C), wind (full and half barb denote 5 and 2.5 m s^{-1}, respectively), and relative humidity (% , shaded following scale at bottom). (c) Infrared satellite image. (d) Composite NEXRAD imagery.
Figure 3.16. Manually analyzed surface map for 0000 UTC 13 February 2000. 1500-m pressure isobars every 2 hPa. Surface station reports include 1500-m pressure (tenths of hPa with leading 8 truncated), and winds [full (half) barbs denote 5 (2.5) m s$^{-1}$]. Primary (secondary) low pressure noted with larger (smaller) font size.
mesoscale convective line through Salt Lake City just before 0100 UTC. Soon after, precipitation abated across northern Utah as dry and more stable conditions infiltrated.

The evolution of the major synoptic features of this event are summarized in Fig. 3.8. A surface cold front made landfall onto the coast of California at 0000 UTC and featured a rearward sloping structure. As the front moved inland, it deformed, split, and weakened. The thermodynamic signature of the cold front weakened as it moved through northern California and merged with a lee trough in northwest Nevada. The remnants of the Pacific front and lee trough then moved as a surface pressure minimum rapidly across Nevada where limited surface and upper-level observations prevented a detailed analysis of its vertical structure. Eventually, over Utah, a complex kinematic structure was observed with a well-defined mid-level trough preceding the surface trough between 3 and 4 h. Factors contributing to the development of the complex kinematic features and their importance on precipitation development and distribution over northern Utah are examined in Chapter 4.
Precipitation development and distribution across northern Utah was affected considerably by the changing stability, moisture, and kinematic environments that accompanied the passages of the synoptic and mesoscale features described in Chapter 3. Therefore, this storm will be divided into three stages according to the temporal evolution of the synoptic and mesoscale features. The first stage, the post-upper-level jet stage, began around 0900 UTC with the passage of an upper-level jet and associated a 500-hPa vorticity maximum (Fig. 3.5a). As the upper-level jet approached, low and mid-level relative humidities increased across northern Utah (Fig. 4.1), and by 0900 UTC, mountain precipitation developed with precipitation falling at most lowland locations by 1200 UTC. At 1200 UTC, southerly flow was generally present at the surface over northern Utah ahead of the approaching low center (Fig. 3.10), while a 10–12 m s⁻¹ southerly low-level jet was centered ~1 km above ground level (AGL) at SLC (Fig. 4.1) and DPG (not shown). Above the jet core, winds veered with height and were southwesterly at the crest level of the Wasatch (~700-hPa).

The 1200 UTC SLC sounding revealed that, except for shallow layers near the surface and 775-hPa, potential temperature increased with height, indicating the atmosphere was stable to dry motion (Fig. 4.2). Equivalent potential temperature,
Figure 4.1. Salt Lake City time-height section from 0000 UTC 12 February – 0300 UTC 13 February 2000. Full (half) barbs denote 5 (2.5) m s$^{-1}$. Equivalent potential temperature contoured every 2˚C. Dark (light) shading denotes relative humidity greater than 90 (80)%. Major synoptic and mesoscale features labeled.
Figure 4.2. Vertical profiles of potential temperature (theta) and equivalent potential temperature (theta-e) for 12 February 2000 at SLC. (a) 1200 UTC. (b) 1500 UTC. Full (half) barbs denote 5 (2.5) m s\(^{-1}\). Critical cross barrier speed (\(U_c\)) given on upper left.
however, remained constant or decreased slightly with height from the surface to 600-hPa, indicating neutral to slightly convectively unstable conditions.

To assess the potential for topographic blocking in this case, we define the Froude number as

\[ Fr = \frac{U}{NH} \]  

where \( U \) is the mean cross-barrier wind speed, \( N \) is the square root of the Brunt-Vaisala frequency, and \( H \) is the mountain height. Blocking may be expected for flows where \( Fr < 1 \), with the degree of blocking increasing with decreasing \( Fr \) (e.g., Manins and Sawford 1982). Since horizontal and vertical variations in \( U, N, \) and \( H \) prevent an unambiguous calculation of \( Fr \) using observed data, we examine the blocking potential using the critical cross-barrier wind speed,

\[ U_c = NH \]

which represents the speed at which \( Fr = 1 \). This eliminates one source of ambiguity (\( U \)) and allows us to use a range of values for \( H \) to more carefully evaluate the potential for blocking. Critical cross-barrier wind speed ranges were calculated using the mean surface – 700-hPa \( N \) and mountain heights of 1400 +/- 400 m. Based on equation 4.2, blocking would be expected if \( U < U_c \).

As denoted on Fig. 4.2a, the critical cross-barrier wind speed range at 1200 UTC was 13.5 m s\(^{-1}\) +/- 3.9 m s\(^{-1}\). At this time, maximum below crest level winds of 10–12 m s\(^{-1}\) were present, generally less than the critical wind speed range, and low-level flow
blocking and channeling was observed across northern Utah (Fig. 4.3a). Difluence was observed upwind (south) of the Oquirrh Mountains and terrain channeled southerly flow was found over the Tooele and Salt Lake Valleys. This southerly flow merged with southwesterly flow over the eastern Great Salt Lake, forming a confluence line ~30–40 km upstream of the northern Wasatch Mountains.

Between 1200 UTC and 1500 UTC, low-level stability increased associated with subcloud evaporative cooling since light rain was reported with cloud bases varying from 1000m and 1300m AGL at SLC. By 1500 UTC convective stability was present from the surface to 800-hPa (Fig. 4.2), but a shallow (~25hPa) convectively and statically neutral stable layer was present immediately above 800-hPa. Stability decreased, however, between 600 and 500-hPa, but remained convectively stable (Fig. 4.2b).

Radar echoes were generally widespread (e.g., Fig. 4.4a) during the first storm stage, and KMTX radar images indicated embedded areas of higher reflectivity propagating through northern Utah at times (not shown). There were, however, significant spatial and temporal variations in precipitation. Elevations and locations for stations where precipitation and/or temperature data are given within this thesis are illustrated in Fig. 4.5. The Salt Lake and Tooele Valleys received little precipitation during the first storm stage (e.g., SLC trace, Fig. 4.6), whereas to the north, considerable precipitation was observed at OGD and SNX (Fig. 4.6) where the maximum precipitation rate was ~1.5 mm h⁻¹ (Fig. 4.6). One of the factors that may have contributed to low precipitation amounts across the Salt Lake and Tooele Valleys was subcloud evaporation given that cloud base varied between 800 and 1100m AGL at SLC, and unsaturated conditions below cloud base (Fig. 4.1). This unsaturated low-level environment may have been aided by downslope low-
Figure 4.3. Northern Utah manual streamline analysis at (a) 1200 UTC 12 Feb, (b) 1800 UTC 12 Feb, (c) 2200 UTC 12 Feb, (d) 0000 UTC 13 Feb 2000. Station data represents elevations between 1280m – 2200m above sea level. Full (half) barbs denote 5 (2.5) m s\(^{-1}\).
Figure 4.4. Base reflectivity radar imagery from Promontory Point (KMTX) at (a) 1100 UTC 12 Feb, (b) 2000 UTC 12 Feb, (c) 2200 UTC 12 Feb, and (d) 0100 UTC 13 February 2000. Reflectivity scale (dbz) given in (a). KMTX location denoted in (d).
Figure 4.5. Location and elevation for selected stations. See Fig. 1.2 for major geographic features.
Figure 4.6. Precipitation data for selected lowland stations from 0600 UTC 12 February – 0600 UTC 13 February 2000. (a) Hourly precipitation. (b) Accumulated precipitation. See Fig. 4.5 for station locations and elevations.
level flow (Fig. 4.3a) to the lee of the mountain ranges that form the southern termini of the Salt Lake and Tooele Valleys, as well as precipitation shadowing due to cross-barrier, southwesterly 700-hPa flow over the Stansbury and Oquirrh mountains. Factors that may have contributed to greater precipitation at OGD and SNX included more saturated low-levels than SLC (Fig. 4.7a), and mesoscale ascent associated with the confluence zone that was evident upstream of the northern Wasatch (Fig. 4.3a and 4.3b).

The greatest precipitation amounts during the first storm stage were observed in the higher mountain areas (Fig. 4.8) with maximum precipitation rates ~4 mm h$^{-1}$ (Fig. 4.8). Note that hourly precipitation data from Ben Lomond Peak (BLP) are believed to be inaccurate due to sensor type. Orographic enhancement over mountainous areas likely dominated cloud and precipitation development with a mean, below crest-level cross-barrier component varying between 3–9 m s$^{-1}$ at 1800 UTC (not shown). Cross barrier flow at NSSL-4 of 7–9 m s$^{-1}$ was much greater than the 3–5 m s$^{-1}$ observed at NSSL-5. This can be attributed to the proximity of NSSL-5 to the northern Wasatch crest, where the low-level environment was within the upstream blocked flow regime. Such subtle variations reflect the importance of local orographic interactions when attempting to categorize the proper upstream storm environment. By 1800 UTC, the confluence line had moved closer to the northern Wasatch barrier (Fig. 4.3b), and precipitation rates remained steady at most locations.

The second storm stage, the post-mid-level trough stage, commenced with the passage of the mid-level trough at 1800 UTC (Fig. 4.1), and was marked by veering of the winds between 500-hPa and 700-hPa, an increase in wind speeds with a mid-level jet centered around 700hPa, and subtle changes in moisture and temperature. The mid-level
Figure 4.7. Vertical profiles of dewpoint ($T_d$) and temperature ($T$) in top panel, and potential temperature (theta), equivalent potential temperature (theta-e) in bottom panel for 1800 UTC 12 February 2000. Full (half) barbs denote 5 (2.5) m s$^{-1}$. Critical cross barrier speed ($U_c$) denoted. (a) and (b) NSSL-4. (c) and (d) NSSL-5. (e) and (f) SLC. See Fig. 1.2 for locations.
Figure 4.8. Precipitation data for selected mountain stations from 0600 UTC 12 February – 0600 UTC 13 February 2000. (a) Hourly precipitation. (b) Accumulated precipitation. See Fig. 4.5 for station locations and elevations.
trough arrived before the primary 500-hPa trough (Fig. 3.13) and surface trough (Fig. 3.14). Satellite imagery indicated the mid-level trough was embedded within a dense cloud shield ahead of areas of clearing to the west (Fig. 3.13c), and no distinct, organized radar reflectivity patterns were coincident with mid-level trough passage (not shown). The low-level flow remained similar to that at 1800 UTC (Fig. 4.3b) throughout the second storm stage, although the confluence line became increasingly confined to near the barrier, as synoptic southerly flow ahead of the approaching surface trough persisted (Fig. 3.14) until the arrival of the surface trough. As denoted on Figs. 4.7b, 4.7d, and 4.7f, the critical cross-barrier wind speeds at 1800 UTC ranged ~15 m s\(^{-1}\) to 16 m s\(^{-1}\) +/- 4 m s\(^{-1}\). Low-level flow remained blocked or partly blocked since maximum cross-barrier winds of 12 m s\(^{-1}\) at NSSL-4 and 7 m s\(^{-1}\) at NSSL-5 were less than the critical cross-barrier wind speeds. By 1800 UTC convectively neutral to slightly stable conditions were present in the low and mid levels at NSSL-4 (Fig. 4.7b), NSSL-5 (Fig. 4.7d), and SLC (Fig. 4.7f).

Between 1800 UTC and 2100 UTC at SLC, conditions generally grew more unstable in both the mid and low levels as temperatures dropped ~2°C between 300-hPa and 450hPa, 1°C or less from 500-hPa to 700-hPa, and little or no warming at most surface stations (not shown). By 2100 UTC neutral stability was present at NSSL-4 from the surface to 700hPa (Fig. 4.9), slight convective instability at SLC extended from the surface to 700hPa (Fig. 4.9), but stable conditions existed from the surface to 775-hPa at NSSL-5 (Fig. 4.9d) where there is evidence of a shallow cold pool. Also, slightly drier air infiltrated in the mid and upper levels after mid-level trough passage (not shown). By 2100 UTC, the critical cross-barrier wind speeds decreased slightly, ranging ~ 13.5 m s\(^{-1}\) to 15.5 m s\(^{-1}\) +/- 4 m s\(^{-1}\). Low-level flow remained mostly blocked at NSSL-5 since maximum
Figure 4.9. Vertical profiles of dewpoint ($T_d$) and temperature ($T$) in top panel, and potential temperature (theta), equivalent potential temperature (theta-e) in bottom panel for 2100 UTC 12 February 2000. Full (half) barbs denote 5 (2.5) m s$^{-1}$. Critical cross barrier speed ($U_c$) denoted. (a) and (b) NSSL-4. (c) and (d) NSSL-5. (e) and (f) SLC. See Fig. 1.2 for locations.
cross-barrier wind speeds of 10 m s\(^{-1}\) was much less than the critical cross barrier wind speed of 15.4 m s\(^{-1}\) (Fig. 4.9d), but the environment at NSSL-4 was conducive for less blocking since the maximum cross-barrier wind of 15 m s\(^{-1}\) was approximately equal to the critical cross barrier speed of 15.2 m s\(^{-1}\) at NSSL-4.

Significant spatial gradients in precipitation distribution occurred during the second storm stage. Little precipitation was again observed across the Salt Lake and Tooele Valleys (e.g., SLC Fig. 4.6b). Subcloud evaporation may partially explain low precipitation amounts, but the low-level environment became increasingly saturated (Fig. 4.9e), with cloud base lowering to 500m AGL by 2100 UTC at SLC. In contrast, significant low elevation precipitation was observed in the lowlands upstream of the northern Wasatch at SNX and OGD (Fig. 4.6b). The most intense precipitation enhancement in the lowlands upstream of the northern Wasatch occurred around 2000 UTC (e.g., Fig. 4.4b), when precipitation rates reached up to 2.5 mm h\(^{-1}\) at OGD (Fig. 4.6a). Significant mountain precipitation persisted (Fig. 4.8b) where moist low-level air (Figs. 4.9a and 4.9c) ascended the terrain, but precipitation rates remained steady (Fig. 4.8a) in spite of increased mean cross-barrier flow to 8–12 m s\(^{-1}\) at NSSL-4, and 4–6 m s\(^{-1}\) at NSSL-5 by 2100 UTC (not shown). This may have been due to effects with slightly drier air advected in the mid-levels behind the mid-level trough. The precipitation rates during the second stage at the mountain site SBE (Fig. 4.8a) were less than that at the lowland site OGD (Fig. 4.6a), suggesting the significance of terrain-driven confluence enhancing precipitation in the vicinity of OGD. The second stage ended with the arrival of a surface trough between 2100 and 2200 UTC.

The third storm stage, the postsurface trough stage, began with the passage of a
surface trough between 2100 UTC and 2200 UTC. Drier air and increased wind speeds infiltrated after surface trough passage. As the surface trough moved across northern Utah, the accompanying change in winds varied greatly from station to station. At DPG, the wind shift associated with the surface trough was coincident with veering of the surface winds from southerly to west northwest (Figs. 4.3c and 4.10). Across the Salt Lake and Tooele Valleys, however, the surface wind shift lagged behind the minimum in surface pressure or surface trough passage (Fig. 4.3c). This may have been due to the weak surface trough interacting with southerly channeled flow.

Temperature changes associated with the surface trough varied from station to station. At DPG, there was very little temperature change (Fig. 4.11a), and at HAT (Fig. 4.11b) temperatures warmed. Temperature at the William Browning Building (WBB), however, dropped 3°C rapidly (Fig. 4.11c), and also decreased slightly atop Francis Peak (FPK) (Fig. 4.11d). At DPG and HAT, the surface feature is best labeled a trough, but at WBB, the surface feature qualifies as a cold front with a 3°C temperature drop. Diabatic processes and increased turbulent mixing associated with the surface trough may explain station to station temperature differences with surface trough passage. The vertical structure and evolution after the surface trough passed, however, is reminiscent of a cold front. The deepening structure of the postsurface trough layer, where convective instability and west-northwesterly winds were present, is seen at SLC from the surface to 725-hPa at 0000 UTC (Fig. 4.12), for NSSL-4 from the surface to 675-hPa at 0100 UTC (Fig. 4.12b), and for NSSL-5 from the surface to 625-hPa at 0230 UTC 13 Feb. (Fig. 4.12d). An inversion was present as low-level temperatures decreased in the posttrough layer, with a maximum temperature change of ~4°C around 650-hPa, but temperatures remained
Figure 4.10. Dugway (DPG) time-height section from 1200 UTC 12 February – 0000 UTC 13 February 2000. Winds [full (half) barbs denote 5 (2.5) m s\(^{-1}\)]. Y-axis represents meters above sea level, and x-axis hour UTC 12 February 2000. Dashed lines indicate mid-level and surface troughs. See Fig. 1.2 for location.
Figure 4.11. Temperature time series from 0000 UTC 12 February – 0000 UTC 13 February. Dashed line denotes surface trough passage. (a) DPG. (b) HAT. (c) WBB. (d) FPK. See Figure 4.5 for locations.
Figure 4.12. Vertical profiles of dewpoint ($T_d$) and temperature ($T$) in top panel, and potential temperature ($\theta$), equivalent potential temperature ($\theta_e$) in bottom panel for 2100 UTC 12 February 2000. Full (half) barbs denote 5 (2.5) m s$^{-1}$. Critical cross barrier speed ($U_c$) denoted. (a) and (b) NSSL-4. (c) and (d) NSSL-5. (e) and (f) SLC. See Fig. 1.2 for locations.
similar above the posttrough layer. A lack of synoptic continuity, and inconsistent surface temperature changes warrant the surface feature better labeled a trough; between 800-hPa and 600-hPa, however, wind and temperature changes are better classified as a weak cold front.

By the middle of the third storm stage, there was little change in the critical cross barrier wind speeds, ranging 13 m s$^{-1}$ - 15 m s$^{-1}$ +/- 4 m s$^{-1}$. Low-level flow, however, became more cross barrier and strengthened in the posttrough environment, with a maximum cross-barrier component of 19 m s$^{-1}$ for NSSL-4 at 0100 UTC and 15 m s$^{-1}$ for NSSL-5 at 0230 UTC (not shown). This suggests minimal flow blocking during the third storm stage.

Significant spatial gradients in precipitation distribution occurred during the third storm stage. An elongated east-west area of enhanced radar reflectivities near the southern shore of the Great Salt Lake was present immediately after surface trough passage (Fig. 4.4c). This was coincident with precipitation rate maxima at SNX and HAT (Fig. 4.6a), suggesting the surface trough acted as a focused lifting mechanism. This precipitation band moved southward quickly, however, and soon dissipated. Across the Salt Lake and Tooele Valleys, the majority of total storm precipitation came during the third storm stage with SLC receiving over 80%. In contrast, the mountain sites SBE and CLN received a little over a third of their total storm precipitation during the third storm stage. Significant precipitation occurred at the mountain locations (Fig. 4.8b) associated with increased mean below-crest level cross-barrier flow around 15 m s$^{-1}$ at both NSSL-4 and NSSL-5 (not shown), where maximum precipitation rates were ~4 mm h$^{-1}$ (Fig. 4.8a). By 0000 UTC 13 February, winds had shifted west-northwesterly from 700-hPa (Fig. 3.15b) to the
surface (Fig. 4.3d), and confluent flow previously observed in the lowlands upstream of the northern Wasatch was not present along with enhanced precipitation as previously noted over this region (e.g., OGD (Fig. 4.6b). At 0000 UTC 13 February, moist and unstable low-level conditions were present at SLC (Figs. 4.12e and 4.12f), and by 0100 UTC, a southward moving mesoscale convective line (Fig. 4.4d) brought cloud to ground lightning, and caused an aborted P-3 landing at SLC. The maximum precipitation rate at SLC of 2 mm h\(^{-1}\) was coincident with the convective line (Fig. 4.6a). Soon after convective line passage, precipitation abated due to drier (Figs. 4.12a and 4.12c) and more stable conditions.

The storm evolved through a notable cycle of stability across northern Utah. As precipitation onset began during the first storm stage, the low-levels (surface – 700-hPa) were convectively neutral to unstable, but the mid-levels (700-hPa – 500-hPa) were convectively stable (Fig. 4.13). Low-level stability increased until ~1500 UTC when stable conditions existed; however, mid-level stability decreased with near neutral convective instability at 1500 UTC. Between 1500 UTC and 1800 UTC, low and mid-level stability decreased with near neutral to neutral convective stability at 1800 UTC. After the mid-level trough passed and second storm stage began, low-level stability decreased with neutral to slightly convectively unstable conditions at 2100 UTC, and mid-level stability remained unchanged. The third storm stage, however, ushered in convectively unstable conditions in the low-levels, but more stable conditions developed in the mid-levels. During much of this case, low and mid-level stability were out of phase.

Three distinct patterns of accumulated precipitation during IOP3 (Fig. 4.14) are apparent. First, substantial precipitation was observed in the higher terrain of the Wasatch,
Figure 4.13. Cycle of low-level (dotted) and mid-level (solid) convective stability from 0900 UTC 12 February to 0000 UTC 13 February 2000. Based on subjective interpretation of data collected at SLC, NSSL-4, and NSSL-5.
Figure 4.14. Total observed precipitation in mm during IOP3. Contours every 10 mm with observations annotated. Figure reprinted with permission from Cheng (2001).
Stansbury, and Oquirrh Mountains. Second, substantial precipitation amounts were observed in the lowlands upstream of the northern Wasatch over the eastern portion of the Great Salt Lake, in the vicinity of a well-developed, windward confluence zone. Third, there was significant suppression of precipitation both to the lee of the Wasatch Mountains and at many lowland locations including the Salt Lake and Tooele Valleys.
An observational analysis of a frontal cyclone from landfall through the Intermountain region was conducted. As the frontal cyclone moved rapidly inland, the cold front featured a classic, rearward sloping structure, but was weakened and blocked by the Sierra Nevada, causing the front to split into two sections. Concurrently, a lee trough and mesoscale cyclonic vortex (MCV) were present over Nevada. The circulation associated with the MCV remained stationary and independent of the lee trough. Eventually, the remnants (surface pressure trough and vague wind shift line) of the northern cold front section merged with the lee trough, which thereafter became mobile. A surface trough then moved rapidly (~20 m s$^{-1}$) across much of Nevada; its movement appeared governed by upper-level forcing immediately ahead of an upper-level trough axis. As the cyclone moved across the Intermountain region, multiple low pressure areas developed resulting in a discontinuous progression. The low-level structure of the surface trough was influenced by boundary layer effects. For example, surface wind shear associated with the surface trough became better organized during the afternoon hours over northeast Nevada. By the time the surface trough reached northern Utah it became more organized but still had a weak signature. The kinematic structure and temperature changes between 800-hPa and 600-hPa in the posttrough environment, however, resembled that of a weak cold front.
Over northern Utah, a mid-level trough arrived 3 to 4 h ahead of the surface trough, suggesting significant modification of the vertical structure as it moved across the Intermountain region. The vertical structure evolution of this case was similar to that observed by Reynolds and Kuciauskas (1988) and Long et al. (1990). Over the Sierra Nevada, Reynolds and Kuciauskas (1988) observed a cold front that was vertically coupled with what they noted a cold surge/humidity front. By the time the system reached the Tuschar Mountains of southern Utah, however, Long et al. (1990) showed passage of the cold surge/humidity front approximately 7 h ahead of the surface cold front. If one considers the movement of the cold surge/humidity front linear and continuous for the above cases, a propagation speed of ~16 m s$^{-1}$ is found. If the mid-level trough in this case is considered analogous to the aforementioned cold surge/humidity fronts, a propagation speed of ~15 m s$^{-1}$ is derived. These values are similar and reasonable advective speeds for the mid and upper-levels. Therefore, it is suggested that the mid-level trough propagated more or less continuously across the western United States.

The evolution of stability within this case was similar to the stable, neutral, unstable and dissipation stages proposed by Marwitz (1980) and Long et al. (1990) where stability evolves from stable to less stable regimes. This evolution across northern Utah can be explained by the changing thermodynamic and kinematic environments accompanying the transient synoptic and mesoscale features. The first stage began as an upper-level jet and associated vorticity maximum brought colder temperatures aloft, clouds and precipitation. This was similar to that observed by Marwitz (1980) who for most cases showed an upper-level baroclinic zone beginning the first or stable storm stage, and (Long et al. 1990) since the arrival of an upper-level cold surge/humidity front initiated their first or stable
stage. The second stage commenced with the arrival of a mid-level trough that provided slightly colder temperatures, increased wind speeds and drier air. This stage was similar to that observed by Long et al. (1990) since their second (neutral) storm stage began with the arrival of a second upper-level cold surge/humidity front, akin to the mid-level trough observed in this case. The third, postsurface trough stage began with the passage of a surface trough. This was similar to Marwitz (1980), and Long et al. (1990) where their third (unstable) stage began after the passage of the surface cold front/trough.

During the three storm stages, factors important for cloud and precipitation development and distribution across northern Utah varied. During the first and second stages, precipitation development was driven by a combination of synoptic ascent associated with vorticity advection increasing with height and topographic effects, including direct orographic ascent/descent and terrain-driven confluence. Much of the lowland precipitation, especially in lowlands distant from higher terrain, during the first two stages was generated by synoptic ascent, but for the most part, little fell due to significant subcloud evaporation and poor organization. In contrast, greater precipitation was observed in the lowlands upstream of the northern Wasatch, where a windward, terrain-driven confluence zone was present and likely enhanced precipitation development. Significant precipitation also occurred over the higher mountains locations where prolonged, moist, cross-barrier flow ascended the higher terrain. Terrain-driven confluence and diffluence was not as important for precipitation development during the third stage, but two organized mesoscale precipitation bands contributed greatly to precipitation distribution.

This case demonstrates the overwhelming importance of mesoscale terrain-driven processes, specifically, direct orographic ascent/descent and low-level terrain-driven circu-
lations, affecting precipitation development and distribution across northern Utah. For the most part, precipitation accumulation in this case exhibited a direct relationship with elevation, but enhanced precipitation in the lowlands upstream of the northern Wasatch demonstrated how climatological precipitation versus elevation relationship did not apply. Cloud and precipitation development and distribution across northern Utah was driven largely by local terrain-driven processes, interwoven within the accompanying synoptic-scale kinematic and thermodynamic changes.

Since similar synoptic patterns of southerly to southwesterly flow ahead of approaching winter storm systems across northern Utah are frequent, this case may benefit the understanding and forecasting of precipitation. Different storm stages can be identified according to the synoptic pattern, and these stages can be recognized as being conducive or detrimental to cloud and precipitation development at various locations. For example, as seen in this case, much of the precipitation that fell in the Salt Lake and Tooele Valleys occurred post surface trough, analogous to “postfrontal,” where low-level instability and moisture is maximized. Similar postfrontal trends have also been noted by Long et al. (1990) and Horel and Gibson (1994).

This research demonstrates the complexities of frontal cyclone evolution over the western United States, and the mesoscale effects terrain has on their accompanying precipitation systems. This case, and the recent work of others over the Intermountain region, further emphasizes the mesoscale nature of fronts, cyclones and their accompanying precipitation systems due mainly to interactions with topography.

Surface data collected by the MesoWest networks were invaluable for this analyses, but in order to document the complete vertical evolution of frontal cyclones over the
Intermountain region, more upper-level and surface data is needed. For example, the mid-level trough would not have been observed with greater than 3-h resolution data; therefore, the upstream evolution was unclear. Future work is needed to better understand the synoptic evolution over the Intermountain region, and the relative role underlying topography has in shaping the vertical structure and mesoscale kinematics of winter storms over northern Utah. Further observational case studies coupled with high-resolution mesoscale modeling should be conducted in order to build a better understanding.
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