AN OBSERVATIONAL AND NUMERICAL MODELING INVESTIGATION OF
GREAT SALT LAKE-EFFECT SNOW

by

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ABSTRACT

The structure and evolution of a lake-effect snow event associated with the Great Salt Lake (GSL) is described using observational and numerical modeling approaches. This event occurred in an environment characterized by low-level instability, large lake–land and lake–700 hPa temperature differences, and low-level flow nearly parallel to the major axis of the GSL. Localized heating over the relatively warm GSL is shown to have induced mesoscale pressure troughing, land-breeze circulations, and low-level convergence that led to the development of convective updrafts and a wind-parallel band of clouds and precipitation. The hyper-saline content of the GSL produced reduced moisture fluxes compared to fresh water. Resulting moisture fluxes were sufficient, however, to enhance precipitation rates. Orographically-induced circulations did not play a major role in the formation of the bands, but orographic uplift (subsidence) enhanced (reduced) precipitation rates. Model diagnostics and sensitivity studies are used to examine the predictability of this event given known uncertainties in the specification of lake/land properties and large-scale conditions.
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CHAPTER 1

INTRODUCTION

The prediction of lake-effect snowstorms that develop over and downwind of the Great Salt Lake (GSL) is one of the major forecast challenges facing meteorologists in northern Utah. Occurring several times each year, Great Salt Lake-effect (GSLE) snowstorms last from a few hours to more than a day, frequently produce snowfalls of 20–50 cm, and have contributed to the state record 129-cm lowland storm-total snowfall that was observed near Salt Lake City (SLC) from 24–28 February 1998 (Carpenter 1993; Slemmer 1998, Steenburgh et al. 2000). Despite significant improvement in observational technologies and numerical forecast systems, GSLE snowstorms remain difficult to predict with lead times of more than a few hours.

Previous studies have identified the climatological characteristics, large-scale conditions, and mesoscale precipitation structures associated with GSLE snowstorms. Based on lake-effect events identified by visual observations and spotter reports, Carpenter (1993) found that GSLE snowstorms were associated with post-cold-frontal northwesterly flow at 700 hPa\(^1\), a lake–700-hPa temperature difference of at least 17 °C (which approximately represents a dry-adiabatic lapse rate), and an absence of stable layers or inversions near or below 700 hPa. Steenburgh et al. (2000) used observations

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\(^1\) The GSL surface is near 850 hPa so the 700 hPa level represents an approximate steering level for lake-effect convection.
from a recently installed National Weather Service Weather Surveillance Radar-1988 Doppler (WSR-88D) near SLC to identify GSLE events between September 1994 and May 1998. During this period, 16 well-defined GSLE events were observed, with an additional 18 marginal events identified in which possible lake-enhancement of precipitation occurred coincident with other precipitation processes. Synoptic, mesoscale, and convective characteristics of these events were examined using National Centers for Environmental Prediction (NCEP) Rapid Update Cycle (RUC; Benjamin et al. 1991,1994) analyses, SLC radiosonde observations, and local WSR-88D radar observations. In addition to supporting the findings of Carpenter (1993) regarding the large-scale environment in which GSLE snowstorms occur, it was also found that such events are characterized by a positive lake–land temperature difference that usually exceeds 6 °C and are often initiated and most active during the overnight and early morning hours. It was hypothesized that large lake–land temperature differences contributed to the development of land-breezes and low-level convergence that focused the development of convection over the GSL. The greater frequency of lake-effect precipitation during the overnight and early morning hours may be related to the diurnal radiational surface heating and cooling cycle that modulates the magnitude of the lake–land temperature difference and associated over-lake land breeze convergence, similar to that described by Passarelli and Braham (1981) over Lake Michigan.

GSLE snowstorms share many similarities with lake-effect snowstorms over the Great Lakes region of the United States (Carpenter 1993; Steenburgh et al. 2000). Wiggin (1950) described the general characteristics of lake-effect snowstorms in the latter region, including their potential for large accumulations and significant variations in snowfall over
short spatial scales. Additionally, Wiggin (1950) noted that such storms were favored in polar continental air-masses during periods of large lake–air temperature differences, near-adiabatic lapse rates, and a long over-water fetch. Peace and Sykes (1966) studied a lake-effect snowband using a mesoscale surface observing network over the east end of Lake Ontario. It was found that a narrow convergence line accompanied the snowband and it was hypothesized that heating (presumably surface sensible and latent heating) caused the formation of the snowband with winds aloft controlling the location and movement of the band. Subsequent studies over the Great Lakes have identified a variety of lake-effect precipitation structures including:

(a) Broad area coverage, which may include multiple wind-parallel bands or open cells (Fig. 1.1a). The wind-parallel bands have been shown to be organized by horizontal roll convection in the boundary layer (Kelly 1982, 1984). Snowfall due to this type of lake-effect snowstorm is typically light compared to other types of lake-effect precipitation, but covers a large area (Niziol et al. 1995). This type of precipitation structure most often occurs when strong synoptic wind flow is perpendicular to the long axis of a lake (Hjelmfelt 1990, Niziol et al. 1995).

(b) Shoreline bands that form roughly parallel to the lee shore due to the convergence of a land breeze with the large-scale wind field (Ballentine 1982, Braham 1983, Hjelmfelt and Braham 1983; Hjelmfelt 1990; Fig. 1.1b). Significant accumulation can occur because these snowbands tend to remain stationary (Niziol et al. 1995).

(c) Midlake bands that form when large-scale flow is parallel to the long axis of a lake and a lake–land temperature contrast exists. Convergence produced by land breezes from the opposite shorelines produces an elongated band of precipitation
which may extend over the lee shore (Fig. 1.1c). These events produce the heaviest lake-effect snowfall events (Peace and Sykes 1966, Passarelli and Braham 1981, Braham 1983, Hjelmfelt 1990, Niziol et al. 1995).

(d) Mesoscale vortices that form in a polar air mass under conditions of a weak surface pressure gradient and large lake–air temperature differential (Fig. 1.1d). They have been found over lakes Michigan, Superior, and Huron (Forbes and Merritt 1984).

Steenburgh et al. (2000) found that precipitation during GSLE events was most frequently characterized by the irregular development of radar echoes over and downstream of the GSL (Fig. 1.2a,b). The most commonly observed organized precipitation structures were solitary wind-parallel bands resembling midlake bands found over the Great Lakes (Fig. 1.2c), and broad-area coverage precipitation shields that formed near the GSL lee shoreline (Fig. 1.2d). Structures resembling shoreline-parallel bands of Lake Michigan, however, have been observed subsequently as will be discussed in later chapters (e.g. Fig. 1.2e). In addition, lake-effect precipitation sometimes occurred in concert with orographic precipitation (e.g., Fig. 1.2f) or embedded within a broader-scale precipitation shield associated with synoptic-scale or local orographic lifting. Significant enhancement of predominantly lake-effect events is also possible when lake-induced precipitation features such as solitary wind-parallel bands interact with the downstream orography.

Several studies have used numerical models to examine lake-effect snowstorm dynamics (e.g., Lavoie 1972, Ballentine 1982, Hjelmfelt and Braham 1983, Hjelmfelt 1990). Using a three-layer primitive equation model, Lavoie (1972) found that frictional convergence due to land–water roughness contrasts, and surface sensible heating due to lake–air temperature differences, produce upward vertical motion and elevated inversion
Figure 1.1. Schematic diagram of four types of lake-effect snowstorms on Lake Michigan. Arrows denote the low-level wind flow and contours represent cloud elements. (a) Broad-area coverage with embedded wind-parallel bands. (b) Shoreline-parallel band. (c) Midlake band. (d) Mesoscale vortex. Adapted from Hjelmfelt (1990).
Figure 1.2. Lowest-elevation base-reflectivity analyses from the WSR-88D radar at (a) 1445 UTC 25 February 1998, (b) 1325 UTC 27 February 1998, (c) 1300 UTC 27 November 1995, (d) 1810 UTC 30 March 1998, (e) 0830 UTC 07 December 1998, (f) 2350 UTC 17 January 1996. Reflectivity (dBZ) according to scale at left.
heights near the lee shoreline of Lake Erie. The lake–air temperature difference was found to be dominant. Hjelmfelt (1990, 1992) examined the importance of low-level instability, lake–land temperature difference, sensible and latent heat fluxes, topography, capping inversions, and upstream moisture in producing lake-effect snowstorms over Lake Michigan. He found that both shoreline-parallel and midlake snowbands were favored by strong lake–land temperature difference, weak stability, and the absence of capping inversions at low elevations. Moderate cross-lake flow enhanced land-breeze induced convergence, thus strengthening shoreline-parallel bands. Midlake snowbands, however, were favored by strong wind flow parallel to the long axis of the lake. Weaker wind flows combined with strong lake–land temperature differences tended to produce mesoscale vortices instead of midlake bands. Upstream moisture was also found to be important in enhancing lake-effect precipitation and land-breeze strength due to latent heat release from condensation. Ballentine et al. (1998) described a successful simulation of a Lake Ontario snowband using the Pennsylvania State University–National Center for Atmospheric Research Mesoscale Model version 5 (MM5) with a maximum nested-grid resolution of 5 km. The simulation reproduced the observed precipitation distribution, although changes in snowband location in response to the evolving synoptic-scale flow had timing errors of a few hours.

The purpose of this dissertation is to advance current knowledge of GSLE snowbands using observational and numerical-modeling approaches. Specific questions that will be addressed include:

• Can present-day mesoscale models accurately simulate the mesoscale circulations and precipitation patterns observed during GSLE snowstorms? Assuming successful
simulations of GSLE snowbands in research-mode, what are the predictability limits when these models are applied in an operational environment? Does the “fixed” surface forcing of the lake and surrounding topography extend predictability, as hypothesized by Paegle et al. (1990), or do small errors in surface characteristics and the upstream and regional flow characteristics limit forecast skill?

- How do sensible and latent heat fluxes influence the development and intensity of lake-effect precipitation? For example, how does localized surface heating over the GSL, and diurnally-modulated radiational heating and cooling of the surrounding land-surface, influence thermally-driven lake- and terrain-induced circulations that may trigger some GSLE events? How significant are moisture fluxes from the GSL? Does the hyper-saline composition of the GSL significantly affect latent heat flux (compared to fresh water) and snowband evolution or intensity?

- Do shoreline roughness contrasts influence the development of GSLE snowbands?

- How important are topographical effects such as orographic uplift and low-level flow blocking and channeling? To what degree are GSLE events triggered or enhanced by the local topography?

The remainder of this dissertation is organized as follows. Chapter 2 describes the regional orography and unique characteristics of GSL hydrology, composition, and air-lake interactions. Chapter 3 presents a detailed observational analysis of a GSLE snowband using RUC analyses, radar observations, and surface observations from the Utah Mesonet (Stiff 1997) and is followed by Chapter 4 which examines the mesoscale structure and evolution of the event using a numerical simulation. Chapter 5 presents the
Several unique aspects of the land-surface properties and orography of northern Utah may influence lake-effect precipitation processes. These include the region’s intense and complex vertical relief, and the varying hydrologic structure, thermal characteristics, and hyper-saline composition of the GSL. The GSL surface is located just below 1300 m with several steeply-sloped mountain ranges extending to over 3000 m located south and east of the lake (Fig. 2.1). To the east and southeast are the Wasatch Mountains, which are oriented roughly meridionally and rise abruptly to elevations of 2500–3500 m. South of the GSL, a series of meridionally-oriented mountain ranges (i.e., the Cedar, Stansbury, and Oquirrh Mountains) interspaced by lowland valleys are found. The Oquirrh Mountains rise directly from the GSL shoreline to heights of 2500–3250 m; similar elevations are found in the Stansbury Mountains. These mountain ranges can produce considerable orographic uplift and terrain-driven convergence.

Lowland regions between these mountain ranges, including the Salt Lake and Tooele Valleys, are approximately 25 km in width. These valleys feature broadly sloped relief that may also contribute to orographic precipitation enhancement. For example, the city of Tooele, 17 km from the GSL shore, is located 215 m above the northern portion of the Tooele Valley and GSL, while broadly sloped benches rise typically 150-400 m above lake level on the West, South, and East sides of the Salt Lake Valley. This relief is
Figure 2.1. Geographic features of northern Utah. Surface elevation in meters shaded according to scale at bottom. Station locations discussed in text are: Salt Lake City (SLC), Ogden (OGD), Provo (PVU), Tooele (TOO), Hat Island (HAT), Gunnison Island (GNI), Interstate-80 (S17), and the Promontory Point NEXRAD radar site (KMTX). Railroad causeway identified by a dashed line.
comparable to that found east of Lake Michigan and the Tug Hill Plateau of northern New York where significant orographic enhancement of lake-effect precipitation from Lake Ontario occurs (Muller 1966, Hjelmfelt 1992, Niziol et al. 1995).

Other important orographic features include the Great Salt Lake Desert, a lowland area west of the lake, and the Raft River Mountains northwest of the lake. Thus, flow from the northwest, which is associated with lake-effect storms (Carpenter 1993; Steenburgh et al. 2000), must traverse substantial topography before moving over the GSL.

The GSL is the largest body of water in the United States west of the Great Lakes. It currently occupies an area of ~4400 km$^2$, is about 120 km long and 45 km wide, and has an average (maximum) depth of only 4.8 (10) m$^1$. Due to the lack of a drainage outlet, the lake’s size fluctuates due to interseasonal and interannual variations in precipitation and evaporation and has ranged between approximately 2500 and 6200 km$^2$ in area and 1278 and 1284 m in surface elevation since the mid 1850s (Arnow 1980, Wold et al. 1996).

Due to the GSL’s shallow depth, lake-surface temperature changes can occur rapidly, and little lag relative to climatological mean air temperatures at Salt Lake City (SLC; see Fig. 2.1 for location) is observed (Fig. 2.2). The average lake-surface temperature exhibits a maximum (minimum) near 1 August (1 February), similar to the timing of the maximum (minimum) mean air temperature at SLC on 24 July (5 January). From late winter through summer, the mean lake temperature is similar to the mean air temperature at SLC, but during the fall through early winter, the mean lake surface

1. Information on lake levels, salinity, surface area, and depth provided by the United States Geological Survey, Salt Lake City, Utah and Wold et al. 1996.
temperature exceeds the mean air temperature at SLC by 2–3 °C, apparently due in part to a small seasonal lag in lake temperature.

Carpenter (1993) suggested that lake-surface temperatures may correlate with the preceding week’s mean air temperature. In the past, estimates of lake-surface temperature using this method were necessary for operational forecasting due to the lack of real-time observations (lake-temperature observations from the Saltair Boat Harbor in Fig. 2.2 are from historical data and are available only bimonthly). However, starting in late summer 1998, lake-surface temperatures have been observed at the Hat Island Utah Mesonet site (HAT; see Fig. 2.1 for location). A comparison between the mean daily lake temperature and mean daily air temperature at SLC and HAT for the initial 5-month observation period is presented in Fig. 2.3. Fluctuations in lake temperature can be seen to coincide with air
temperature changes associated with transient synoptic-scale systems at both HAT and SLC. However, lake temperatures changed slowly enough that large lake–air temperature differences were observed at times due to occasional rapid air temperature changes. Additionally, lake surface temperatures slightly exceeded air temperatures through most of this fall and early winter period similar to the climatological mean curve (Fig. 2.2).

The GSL is a terminal lake (i.e., has no outlet) and can be up to 8 times as saline as ocean water. Currently, the lake is divided by an earthen railway causeway which limits mixing between the northern and southern sections, named Gunnison Bay and Gilbert Bay, respectively (Sturm 1980, Butts 1980, Newby 1980). Gunnison Bay has only limited fresh-water inflow and generally features salinity near saturation (27%). Salinity in Gilbert Bay, which has several fresh-water inlets, has ranged from 6–15% and is currently near 9% (as of December 1998). Due to the high salinity, the lake never freezes over except
near fresh-water inlets. Because the lake never freezes over in its entirety and can warm rapidly, lake-effect snow is possible from early fall through late spring (Steenburgh et al. 2000). The salinity also acts to reduce saturation vapor pressure and latent-heat fluxes compared to those found under similar conditions over fresh water. Due to the molecular attractions between solvent (water) and solute (dissolved salt ions), the saturation vapor pressure over the GSL must be measured experimentally. Dickson et al. (1965) have done this for very high salinities such as those found in Gunnison Bay. At lower salinities, saturation vapor pressure is closer to that of an ideal solution and may be approximated by Raoult’s law. Since the major salt dissolved in the GSL waters is sodium chloride (NaCl), a possibly more accurate approximation may be achieved by consideration of the vapor pressure reduction due to pure NaCl solution. The sodium chloride approximation uses data from Harned and Owen (1958) to solve eq. 7 of Low (1969). Fig. 2.4 presents the results of these three approximations. Given the current salinity of Gunnison (Gilbert) Bay, the ratio of saturation vapor pressure over saline water to saturation vapor pressure over fresh water is approximately 0.70 (0.94). Due to this reduction in saturation vapor pressure over Gunnison (Gilbert) Bay, all upward moisture fluxes based on the bulk aerodynamic calculation (Krishnamurti 1996) would be eliminated if the boundary layer dewpoint were as much as 5 °C (.9 °C) lower than the lake temperature.2 The implications of salinity on moisture fluxes and precipitation will be discussed further in Chapters 4 and 5.

2. This example was calculated using a lake temperature of 5 °C and results vary less than 13% over the range of lake temperatures which have been observed during GSLE events. A lake temperature near 5 °C is the most common temperature observed during GSLE events (Steenburgh et al. 2000).
Figure 2.4. Reduction of saturation vapor pressure ($e_s$) compared to that of a plane surface of pure water ($e_{s0}$) as estimated using Raoult’s Law with observed Great Salt Lake brine constituents (solid), determined experimentally for a pure NaCl solution at 0°C (long/short dashed) and 25°C (long dashed), and determined experimentally from Great Salt Lake brine by Dickson et al. (1965) at O°C (short dashed) and 25°C (medium dashed).
CHAPTER 3

OBSERVATIONAL ANALYSIS OF 7 DECEMBER 1998 SNOWBAND EVENT

Synoptic Setting

To examine the synoptic and mesoscale evolution of the 7 December snowband, regional-scale analyses from the National Centers for Environmental Prediction (NCEP) Rapid Update Cycle 2 (RUC2), upper-air soundings from the Salt Lake City International Airport (SLC), radar observations from the KMTX WSR-88D radar, and surface observations from the Utah Mesonet are presented in Figs. 3.1-3.14. At 1200 UTC 6 December 1998, roughly 12 h prior to the onset of the lake-effect event, a large-scale upper-level trough was resident over the western United States, with a 500 hPa trough axis extending poleward from southern California into western Washington (Fig. 3.1c). At 700 hPa the trough axis was located west of the Utah-Nevada border, with a region of significant moisture (i.e., RH > 70%) located coincident with and upstream of this feature (Fig. 3.1b). There was a weak contrast in temperature across the trough with 700 hPa temperatures over southern Utah near -12 °C compared to -16 °C over western Washington and Oregon. A sea level low pressure center was located northwest of Las Vegas beneath a region of cyclonic absolute vorticity advection (Fig. 3.1a,c). Weak sea level pressure troughing extended northeastward from the low center into northern Utah. The observed sounding at SLC shows veering winds with height from the surface to 700 hPa implying
Figure 3.1. Regional RUC2 analyses (a, b, c) and observed upper air sounding taken at SLC (d) at 1200 UTC 6 Dec 1998. (a) Sea level pressure (every 2 hPa) and near-surface winds (full barb denotes 5 m s\(^{-1}\)). (b) 700-hPa temperature (every 2 °C), wind [as in (a)], and relative humidity (%), shaded following scale at upper-right). Trough axes labelled by dashed lines. (c) 500-hPa geopotential height (every 60 m) and absolute vorticity (x10\(^{-5}\) s\(^{-1}\), shaded following scale at upper-right). Trough axes denoted by dashed lines. (d) Skew-T, log-P diagram with temperature (°C) and dewpoint (°C) in heavy solid lines. Short dashed line represents surface parcel ascent. Long dashed line represents modified surface parcel ascent. Filled circle represents lake temperature. Wind as in (a).
low-level warm advection ahead of the trough (Fig. 3.1d). Conditions were not favorable for lake-effect precipitation with southerly to southwesterly flow, a series of stable layers, and 5–20 °C dewpoint depressions evident at low levels.

Twelve hours later at 0000 UTC 7 December, shortly after the onset of lake-effect precipitation, the 500 (700) hPa trough axis had moved over (downstream of) SLC (Fig. 3.2b,c). Although the signature of this trough was relatively weak at the surface, low-level winds became northwesterly to northerly (Fig. 3.2a) and low-level cold advection developed over northern Utah as inferred from backing winds in the SLC sounding at and below 700 hPa (Fig. 3.2d). In fact, the lowest 700-hPa temperatures were now located just upstream of northern Utah (Fig. 3.2b). Visible satellite imagery showed the passage of a band of clouds across the GSL with passage of the 700 hPa trough between 1400 and 1900 UTC, but no precipitation was reported over SLC and OGD (not shown). At SLC, the lapse rate below 650 hPa was slightly more stable than moist adiabatic with a near-adiabatic layer near the surface. The sounding was relatively moist with dew-point depressions of 5 °C or less through most of the troposphere. The pattern described above is similar to that found at the onset time of lake-effect events by Steenburgh et al. (2000).

Other characteristics of the environment were also favorable for the development of lake-effect precipitation. With lake-surface and 700-hPa temperatures of 5 °C and -15.9 °C, respectively, the lake–700-hPa temperature difference of 20.9 °C (12.4 K km⁻¹) exceeded the threshold required for GSLE precipitation identified by Carpenter (1993) and Steenburgh et al. (2000).¹ In addition, although the surface parcel in the SLC sounding, had no convective available potential energy (CAPE), localized sensible and

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¹ Lake surface temperature measured at Hat Island and 700-hPa temperature observed from SLC sounding.
Figure 3.2. Regional RUC2 analyses (a, b, c) and observed upper air sounding taken at SLC (d) at 0000 UTC 7 Dec 1998. (a) Sea level pressure (every 2 hPa) and near-surface winds (full barb denotes 5 m s\(^{-1}\)). (b) 700-hPa temperature (every 2 °C), wind [as in (a)], and relative humidity (%), shaded following scale at upper-right). Trough axes labelled by dashed lines. (c) 500-hPa geopotential height (every 60 m) and absolute vorticity (x10\(^{-5}\) s\(^{-1}\)), shaded following scale at upper-right). Trough axes denoted by dashed lines. (d) Skew-T, log-P diagram with temperature (°C) and dewpoint (°C) in heavy solid lines. Short dashed line represents surface parcel ascent. Long dashed line represents modified surface parcel ascent. Filled circle represents lake temperature. Wind as in (a).
latent heat fluxes over the GSL offer significant potential for boundary-layer destabilization, as illustrated by the hypothetical lake-modified surface parcel\(^2\) path in Fig. 3.3d. This hypothetical modified surface parcel contained 589 J Kg\(^{-1}\) CAPE, a typical value when compared to the results of Steenburgh et al. (2000), which identified 211, 2198, and 852 J Kg\(^{-1}\) as the minimum, maximum, and mean values, respectively, observed during 4 years of GSLE events. In addition, the lake–land temperature difference was 8 °C, near the mean value for GSLE events (Steenburgh et al. 2000).\(^3\) This could result in the development of land-breeze circulations and convergence over the GSL.

By 1200 UTC 7 December 1998 when lake-effect precipitation was occurring in a well-organized band extending into Tooele County, the 500 hPa trough was located well downstream of Utah with ridging building over the western United States (Fig. 3.3c). Over northern Utah, moist (RH > 80%) north to northwesterly flow was evident at 700 hPa with the coldest temperatures at this level located just south of the GSL (Fig. 3.3b). High sea level pressure was found over eastern Nevada with light surface winds over northern Utah (Fig. 3.3a). The sounding was moist (dewpoint depressions < 5 °C) and conditionally unstable below 650 hPa, with a strong inversion near 500 hPa (Fig. 3.3d). As was the case 12 h earlier, the sounding was conditionally unstable, but significant potential for boundary-layer destabilization by localized heating over the GSL remained. In fact, the modified parcel CAPE at this time was 611 J Kg\(^{-1}\), the highest value observed during the event. In addition, the lake–700 hPa temperature difference was 22.5 °C (13.0 K km\(^{-1}\)) and the lake–land temperature difference was 10 °C.

\(^2\) A modified surface parcel here is defined by a temperature and dewpoint set equal to the average of the lake and land temperature and dewpoint (lake dewpoint set equal to lake temperature). The near surface air temperature observed from the SLC sounding is referred to as land temperature here.

\(^3\) Lake–land temperature difference defined as the difference between the lake temperature measured at Hat Island and the near-surface air temperature observed in the SLC sounding.
Figure 3.3. Regional RUC2 analyses (a, b, c) and observed upper air sounding taken at SLC (d) at 1200 UTC 7 Dec 1998. (a) Sea level pressure (every 2 hPa) and near-surface winds (full barb denotes 5 m s⁻¹). (b) 700-hPa temperature (every 2 °C), wind [as in (a)], and relative humidity (%), shaded following scale at upper-right). Trough axes labelled by dashed lines. (c) 500-hPa geopotential height (every 60 m) and absolute vorticity (x10⁻⁵ s⁻¹, shaded following scale at upper-right). Trough axes denoted by dashed lines. (d) Skew-T, log-P diagram with temperature (°C) and dewpoint (°C) in heavy solid lines. Short dashed line represents surface parcel ascent. Long dashed line represents modified surface parcel ascent. Filled circle represents lake temperature. Wind as in (a).
By 0000 UTC 8 December 1998, lake-effect precipitation had ended. At this time, the 500-hPa ridge axis was moving over northern Utah and the sea level high pressure system was located over eastern Utah (Fig. 3.4a,c). At 700 hPa, temperatures had climbed to -12 °C (Fig. 3.4b), presumably from large-scale subsidence beneath the building upper-level ridge and, as can be inferred from veering winds with height at SLC (Fig. 3.4d), lower to middle tropospheric warm advection. The SLC sounding also shows that the inversion base that was previously located near 500 hPa (Fig. 3.3d) had lowered to 700 hPa (Fig. 3.4d). Modified CAPE was now only 184 J Kg⁻¹, the lake-700 hPa temperature difference was 18.5 °C (10.5 K km⁻¹), and the lake-land temperature difference was under 5 °C. These values were near or below the minima observed during lake-effect events observed by Steenburgh et al. (2000). Correspondingly, only shallow, nonprecipitating cumulus were observed over the region.

**Mesoscale Structure**

Between 2200 UTC 6 December and 0400 UTC 7 December, after the passage of the surface and upper-level troughs, disorganized convective cells forming primarily over the lake and advecting downstream to the southeast were observed in radar analyses from the KMTX WSR-88D radar (not shown). At 0400 UTC 7 December, the last of these cells were drifting into the Tooele Valley and the first long-lived snowband (snowband A) began to form near the western shoreline of the GSL (Fig. 3.5)⁴. The upwind extension of this snowband is characterized by reflectivities of 5–20 dBZ which may represent only clouds or virga with little or no snow reaching the surface. The band was approximately parallel

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⁴ At this and subsequent times, a few erroneous high returns exceeding 50 dBZ were found over the high peaks of the Wasatch, Oquirrh, and Stansbury Mountains due to a failure of the ground-clutter suppression algorithm.
Regional RUC2 analyses (a, b, c) and observed upper air sounding taken at SLC (d) at 0000 UTC 8 Dec 1998. (a) Sea level pressure (every 2 hPa) and near-surface winds (full barb denotes 5 m s\(^{-1}\)). (b) 700-hPa temperature (every 2 °C), wind [as in (a)], and relative humidity (%, shaded following scale at upper-right). Trough axes labelled by dashed lines. (c) 500-hPa geopotential height (every 60 m) and absolute vorticity \(\times 10^{-5} \text{s}^{-1}\), shaded following scale at upper-right). Trough axes denoted by dashed lines. (d) Skew-T, log-P diagram with temperature (°C) and dewpoint (°C) in heavy solid lines. Short dashed line represents surface parcel ascent. Long dashed line represents modified surface parcel ascent. Filled circle represents lake temperature. Wind as in (a).
Figure 3.5. Lowest-elevation (.5° elevation angle) base-reflectivity analysis from KMTX WSR 88D doppler radar and Utah mesonet observations at 0400 UTC 7 December 1998. Radar reflectivity shaded according to scale. Station observations include wind barbs (station centered at tip of wind barb; full barb = 5 m s⁻¹), temperature (°C; upper left), and dewpoint (°C; lower left). Incipient snowband A denoted by heavy dashed line. Topographic contours shown every 500 m in solid lines (see Fig. 2.1 for elevations). Lake outline shown with thin dashed line.
to the wind flow on Promontory Point (PRP; see Fig. 3.5 for location), a mountain top observing site approximately 800 m above lake-level that represents a near-steering-level wind for boundary-layer convection. Weak low-level confluence into the northern end of this band, as observed during similar banding events over Lakes Michigan and Ontario (e.g. Peace and Sykes 1966, Passarelli and Braham 1981, Braham 1983) is suggested by a weak shift in surface winds from northerly to northwesterly as the band passed over Gunnison Island (GNI [see Fig. 3.5 for location]; Fig. 3.6a). Elsewhere, surface winds were generally light and northwesterly to northeasterly and surface temperatures ranged from -2 °C to -10 °C with the warmer temperatures found over and near the GSL.

Over the next 75 min, snowband A intensified and at 0515 UTC was located near the west shoreline (Fig. 3.7). Meanwhile, a second snowband (snowband B) developed over the southernmost arm of the GSL and northeast portion of Tooele county. The wind flow in the northern Tooele Valley was confluent toward the northern half of this band. Weak confluence into the northern portion of snowband A is also suggested by the northwest surface wind at GNI and north-northeast surface wind at Hat Island (HAT; see Fig. 3.5 for location).

By 0630 UTC snowbands A and B were beginning to merge into a single, solitary snowband (Fig. 3.8). At this time, snowband A had just passed over HAT where winds shifted from northerly to northwesterly, suggesting low-level confluence along the northern portion of the snowband axis (Fig. 3.6b). Significant changes in temperature or dewpoint were, however, not observed (cf. Figs. 3.7, 3.8). Fifteen min later snowband A moved westward back across HAT resulting in a wind shift back to northerly (Fig. 3.6b; radar analysis not shown). Farther downstream, surface winds beneath snowband A
Figure 3.6. Time-series of winds from 0300 UTC to 1600 UTC 7 December 1998 at (a) Gunnison Island (GNI), (b) Hat Island (HAT), and (c) Promontory Point (PRP). Full barb denotes 5 m s⁻¹. Arrows mark approximate time that snowband A crossed observation site.
Figure 3.7. Lowest-elevation (0.5° elevation angle) base-reflectivity analysis from KMTX WSR 88D doppler radar and Utah mesonet observations at 0515 UTC 7 December 1998. Radar reflectivity shaded according to scale. Station observations include wind barbs (station centered at tip of wind barb; full barb = 5 m s\(^{-1}\)), temperature (°C; upper left), and dewpoint (°C; lower left). Snowbands A and B denoted by heavy dashed lines. Topographic contours shown every 500 m in solid lines (see Fig. 2.1 for elevations). Lake outline shown with thin dashed line.
Figure 3.8. Lowest-elevation (.5° elevation angle) base-reflectivity analysis from KMTX WSR 88D doppler radar and Utah mesonet observations at 0630 UTC 7 December 1998. Radar reflectivity shaded according to scale. Station observations include wind barbs (station centered at tip of wind barb; full barb = 5 m s⁻¹), temperature (°C; upper left), and dewpoint (°C; lower left). Snowbands A and B denoted by heavy dashed lines. Topographic contours shown every 500 m in solid lines (see Fig. 2.1 for elevations). Lake outline shown with thin dashed line.
appeared to be divergent over the western Tooele Valley, perhaps due to convective outflow (Fig. 3.8). In the eastern Tooele Valley, surface winds remained convergent toward the axis of snowband B.

The radar reflectivity analysis at 0815 UTC shows the solitary snowband that developed from the merger of snowbands A and B at one of its most organized stages (Fig. 3.9). At this time the snowband extended from just west of HAT southeastward into Tooele County and was nearly parallel to the flow on PRP. Reflectivity values of 20–30 dBZ comprised much of the snowband and likely represent moderate to heavy snow. Isolated reflectivity values approaching 40 dBZ were observed within the band over the lake, the city of Tooele, and the western slopes of the Oquirrh Mountains. At this time, confluent flow that was previously observed over the Tooele Valley beneath snowband B was weakening as winds were becoming northerly or northwesterly.

Over the next 135 min the snowband became more meridionally oriented and by 1030 UTC extended from near the center of the GSL southward into Tooele County (Fig. 3.10). The meridional snowband orientation represented a slight departure from the near-steering-layer flow direction observed at PRP and developed because of a rapid eastward migration of the snowband formation point (not explicitly shown), perhaps due to a greater eastward penetration of the west shoreline land breeze. As a result, snowband orientation was controlled not just by the steering-layer wind, as would be the case if the formation point was stationary, but also by the formation point motion. On the mesoscale, surface wind observations continued to suggest that the northern portion of the snowband was associated with low-level confluence. Surface winds at HAT veered from northerly to westerly with snowband passage, as occurred between 0600 and 0700 UTC, although the
Figure 3.9. Lowest-elevation (.5° elevation angle) base-reflectivity analysis from KMTX WSR 88D doppler radar and Utah mesonet observations at 0815 UTC 7 December 1998. Radar reflectivity shaded according to scale. Station observations include wind barbs (station centered at tip of wind barb; full barb = 5 ms⁻¹), temperature (°C; upper left), and dewpoint (°C; lower left). Snowband A denoted by heavy dashed line. Topographic contours shown every 500 m in solid lines (see Fig. 2.1 for elevations). Lake outline shown with thin dashed line.
Figure 3.10. Lowest-elevation (.5° elevation angle) base-reflectivity analysis from KMTX WSR 88D doppler radar and Utah mesonet observations at 1030 UTC 7 December 1998. Radar reflectivity shaded according to scale. Station observations include wind barbs (station centered at tip of wind barb; full barb = 5 ms⁻¹), temperature (°C; upper left), and dewpoint (°C; lower left). Snowband A denoted by heavy dashed line. Topographic contours shown every 500 m in solid lines (see Fig. 2.1 for elevations). Lake outline shown with thin dashed line.
wind shift appeared to follow the passage of the reflectivity band by 15 to 30 min (Fig. 3.6b). In addition, over-lake convergence was also suggested by the westerly wind at HAT and northeasterly wind at Antelope Island (See Fig. 2.1 for location). This mesoscale wind pattern may have been related to the development of land-breeze circulations due to localized heating over the lake surface. Temperatures over and near the GSL were generally higher than at surrounding locations, but a lack of wind observations prevented diagnosis of wind flows along the western and eastern shorelines. Farther downstream over Tooele County, which was beneath the southern portion of the band, winds were strongly diffluent (Fig. 3.10). Note that at Lake Point, where the Oquirrh Mountains rise abruptly from the GSL, surface winds had shifted from northeasterly to westerly (cf., Figs. 3.9 and 3.10).

By 1315 UTC the snowband was once again aligned along the steering-layer flow and extended southeastward from the GSL over the western Salt Lake Valley (Fig. 3.11). With clearing skies, temperatures dropped rapidly to -10 °C or lower in the central and western Tooele Valley, substantially colder than temperatures over the GSL (Fig. 3.11). As a result, thermally-driven down-valley and off-shore winds developed over the Tooele Valley. Overall, the regional wind pattern suggests the presence of strong low-level convergence over the GSL near the axis of the snowband.

During the next 90 min the snowband gradually deteriorated into an area of precipitation with embedded convective cores that was beginning to drift northeastward by 1445 UTC (Fig. 3.12). At this time surface winds remained convergent over the GSL, but the snowband structure and intensity were beginning to decay for two reasons. First, the near-steering-layers wind at PRP was weakening and beginning to veer to westerly (Fig.
Figure 3.11. Lowest-elevation (.5° elevation angle) base-reflectivity analysis from KMTX WSR 88D doppler radar and Utah mesonet observations at 1315 UTC 7 December 1998. Radar reflectivity shaded according to scale. Station observations include wind barbs (station centered at tip of wind barb; full barb = 5 ms$^{-1}$), temperature (°C; upper left), and dewpoint (°C; lower left). Snowband A denoted by heavy dashed line. Topographic contours shown every 500 m in solid lines (see Fig. 2.1 for elevations). Lake outline shown with thin dashed line.
Figure 3.12. Lowest-elevation (0.5° elevation angle) base-reflectivity analysis from KMTX WSR 88D doppler radar and Utah mesonet observations at 1445 UTC 7 December 1998. Radar reflectivity shaded according to scale. Station observations include wind barbs (station centered at tip of wind barb; full barb = 5 m s⁻¹), temperature (°C; upper left), and dewpoint (°C; lower left). Topographic contours shown every 500 m in solid lines (see Fig. 2.1 for elevations). Lake outline shown with thin dashed line.
3.6c), a direction with a much shorter over-water fetch that is generally not favorable for the development and maintenance of a solitary wind-parallel snowband. Second, warm-advection and subsidence were producing rapid stabilization at mid levels, limiting the depth of surface-based convection (Figs. 3.3 and 3.4). By 1900 UTC the near-steering-layer winds at PRP were west-southwesterly, the lake-effect precipitation area had drifted eastward, and new cells were no longer forming over the GSL (Fig. 3.13).

Radar Composite and Snowfall Distribution

To summarize the distribution and intensity of snowfall during this event, a composite radar-image was generated from the 155 NIDS-formatted radar scans taken from 0000 to 1455 UTC. This involved computing the percentage of time that reflectivity values exceeded 10 dBZ at each point within each radar scan (hereafter referred to as 10 dBZ FO; Fig. 3.14). This method was originally developed by Slemmer (1998) and was used by Steenburgh et al. (2000) to describe GSLE precipitation distribution as a function of various wind and thermodynamic variables. The composite reflectivity analysis shows that a band of frequent returns stretched from near HAT to the western slopes of the Oquirrh Mountains (see Fig. 2.1 for locations), with a secondary 10 dBZ FO maximum in the western Salt Lake Valley where the snowband was resident for a shorter period of time. The highest 10 dBZ FO region (60–80%) extended in a narrow band from near the southernmost tip of the GSL to the city of Tooele. Snowfall totals of 25 and 36 cm (18.8 mm liquid equivalent) were observed at the two reporting sites in this region. Outside this band of heavy snowfall, accumulations were much lower, as indicated by snowfall accumulations of 5 and 8 cm to the south and west, and reports of trace amounts in the eastern Salt Lake Valley.
Figure 3.13. Lowest-elevation (.5° elevation angle) base-reflectivity analysis from KMTX WSR 88D doppler radar and Utah mesonet observations at 1900 UTC 7 December 1998. Radar reflectivity shaded according to scale. Station observations include wind barbs (station centered at tip of wind barb; full barb = 5 m s⁻¹), temperature (°C; upper left), and dewpoint (°C; lower left). Topographic contours shown every 500 m in solid lines (see Fig. 2.1 for elevations). Lake outline shown with thin dashed line.
Figure 3.14. Frequency of occurrence (%) of WSR-88D radar reflectivity values greater than or equal to 10 dBZ (10 dBZ FO) from 0000 to 1455 UTC 7 December 1998 and observed snowfall totals (cm) from 0000 UTC December 7 to 0000 UTC December 8. 10 dBZ FO shaded according to scale at left. Topographic contours shown every 500 m in solid lines (see Fig. 2.1 for elevations). Lake outline shown with dashed line.
CHAPTER 4

SIMULATION OF 7 DECEMBER 1998 SNOWBAND

Mesoscale Model Description

Simulations by the Pennsylvania State University—National Center for Atmospheric Research Mesoscale Model version 5 (MM5; Grell et al. 1995) were used to further examine the evolution of the 7 December snowband event. The MM5 is a non-hydrostatic finite-difference atmospheric model employing a terrain-following sigma vertical coordinate.\(^1\) Simulations featured four one-way nested grids, named domains 1, 2, 3, and 4, with resolutions of 54, 18, 6, and 2 km, respectively (Fig. 4.1). Thirty-six variably spaced full-sigma levels were used in the vertical with resolution varying from approximately 10 hPa in the boundary layer to 30 hPa in the upper troposphere.\(^2\) Precipitation processes were parameterized in all four domains using a mixed-phase microphysical parameterization that includes predictive equations for cloud ice, cloud water, rain, and snow (Grell et al. 1995) and allows for supercooled water below 0 °C and unmelted snow above 0 °C. The Kain-Fritsch cumulus parameterization (Kain and Fritsch

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1. The sigma coordinate is defined as:

\[
\sigma = \frac{p_0 - p_t}{p_s - p_t}
\]

where \(p_0\) is pressure (of a reference state), \(p_t\) is the pressure at the model top (100 hPa in this simulation), and \(p_s\) is the reference surface pressure. This coordinate tends to parallel the topography at low levels and pressure surfaces at upper levels.

2. The sigma levels were located at \(\sigma = 1.0, 0.99, 0.98, 0.96, 0.93, 0.90, 0.87, 0.84, 0.81, 0.78, 0.75, 0.72, 0.69, 0.66, 0.63, 0.60, 0.57, 0.54, 0.51, 0.48, 0.45, 0.42, 0.39, 0.36, 0.33, 0.30, 0.27, 0.24, 0.21, 0.18, 0.15, 0.12, 0.09, 0.06, 0.03, 0.0\) with the model top at 100 hPa.
Figure 4.1. Four domains of MM5 simulation. Domains 1 through 4 (D1 through D4) at 54 km, 18 km, 6 km, and 2 km resolution, respectively. Grid sizes for domains 1 through 4 are 64X80, 112X112, 64X64, and 94X94, respectively.
1993) was used in domains 1, 2, and 3 to represent subgrid scale convective precipitation. Boundary-layer processes were parameterized using the so-called Blackadar scheme that accounts for the vertical mixing of horizontal wind, temperature, mixing ratio, cloud water, and cloud ice in the boundary layer (Blackadar 1976, 1979; Zhang and Anthes 1982). One significant modification was made to the boundary-layer parameterization to account for the impact of lake salinity on saturation vapor pressure and surface moisture fluxes (Fig. 2.4). North of the railroad causeway (Fig. 2.1), the saturation vapor pressure of lake water was set to 70% of that observed for fresh water. This reduction was based on recent salinity observations in the north arm of the GSL (27%) and the saturation vapor pressure measurements obtained for lake water by Dickson et al. (1965) and presented in Fig. 2.4. In the south arm, the saturation vapor pressure was set to 94% of that observed for fresh water based on the observed salinity (9%) and estimates of reduction obtained using Raoult’s law (Fig. 2.4). Other model parameterizations include a long- and short-wave atmospheric radiation scheme that accounts for interactions with the atmosphere, clouds, precipitation, and surface (Dudhia 1989), and a radiative upper boundary condition that allows gravity-wave energy to pass unreflected through the upper boundary (Klemp and Durran 1983).

Observed terrain data, bilinearly interpolated onto the MM5 grid and filtered with a two pass smoother/desmoother, provided the model terrain. For the high-resolution domains (3 and 4), a 30-second resolution data set was used, while 5-min and 10-min resolution data was used for domains 2 and 1, respectively. All landuse information was derived from a 10-min resolution data set, though the Great Salt Lake land-use and elevation was corrected to match the lake shoreline. The domain 4 model topography
represents most of the major terrain features of northern Utah, although mountain crest levels and slopes are somewhat lower and less steep than observed (cf. Figs. 2.1 and 4.2).

Analyses for model initialization, data assimilation, and boundary conditions were generated for domains 1 and 2 at 12-h intervals from 1200 UTC December 6 to 0000 UTC December 8 in the following manner. First, operational surface and upper-level analyses from the National Centers for Environmental Prediction (NCEP) Eta model (Black 1994; Rogers et al. 1995; Rogers et al. 1996), which were available at 80-km horizontal and 50-hPa vertical resolutions, were interpolated onto each domain’s horizontal grid. This provided a first guess for a modified Cressman-style analysis (Benjamin and Seaman 1985) that incorporated rawinsonde and surface data. After the removal of super-adiabatic lapse rates below 500 hPa, the analysis was interpolated to sigma coordinates and the integrated mean divergence was removed to avoid the production of spurious gravity waves. MM5 sea-surface temperatures were generated from operational NOAA analyses that were available on a 1° lat x 1° lon grid.

Four dimensional data assimilation was used to constrain large-scale error growth in domains 1 and 2. Following Stauffer and Seaman (1990), this involved using Newtonian nudging to relax the model simulation to the gridded analyses that were generated using the methods described above. Linear interpolation in time was used between the analyses which were at 12-h intervals. For domain 1, FDDA was used during the entire 36-h simulation, while FDDA was used for domain 2 for only the first 12-h.

Since the number of observations available were not sufficient to adequately resolve features on scales consistent with their grid resolution, initial analyses for domains 3 and 4 were generated simply by interpolation of analyses from their parent grids. FDDA
Figure 4.2. Topography in 2 km resolution domain 4 of MM5 simulation. Surface elevation shaded according to scale at bottom. Labeled cross-sections used later in chapter.
was also not used on these domains, although degradation of forecast skill from large-scale error growth should be reduced because of the superior lateral-boundary conditions provided by the use of FDDA on the outer domains (Vukicevic and Paegle 1989). Domain 3 was initialized at the same time as domains 1 and 2 (1200 UTC 6 December). Domain 4 was initialized 12 h later at 0000 UTC 7 December. Because of computational resource limitations, domain 4 was run after the integration of domains 1–3 was complete. As a result, boundary conditions for this domain were provided by hourly domain 3 output files. The GSL temperature was set to 278 K, the approximate mean lake temperature at HAT during the event period.

**Simulated Structure and Evolution**

**Synoptic-scale Evolution**

Analyses from the 18-km resolution domain are presented in Figs. 4.3–4.5 to examine the synoptic-scale evolution of the model simulation. At 0000 UTC 7 December, the simulated 500-hPa trough axis extended from Arizona to northern Idaho (Fig. 4.3c) and the coldest 700-hPa temperatures were located just upstream of the GSL (Fig. 4.3b). Over northern Utah, northwesterly flow was found at 700 hPa and the surface, with the relative humidity at the former level exceeding 70% (Figs. 4.3a,b). The most notable differences between the model and the RUC2 analyses were the lack of a well-defined 700-hPa trough extending northward through eastern Utah and the placement of the 500-hPa trough axis approximately 50-100 km too far west in the vicinity of the GSL (cf. Figs. 3.2b,c and 4.3b,c). The model-derived sounding at SLC (Fig. 4.3d) showed northwesterly winds extending from the surface to 500 hPa, where the winds abruptly backed to
Figure 4.3. Analysis of the 18-km domain 2 MM5 simulation valid at 0000 UTC 7 December 1998. (a) Sea level pressure (every 2 hPa), near-surface winds (full barb denotes 5 m s⁻¹), and 12-h accumulated precipitation (mm, shaded according to scale at upper right). (b) 700-hPa temperature (every 2 °C), wind [as in (a)], and relative humidity (%), shaded following scale at upper right). Trough axes labelled by dashed lines. (c) 500-hPa geopotential height (every 60 m) and absolute vorticity (x10⁻⁵ s⁻¹, shaded following scale at upper-right). (d) Skew-T, log-P diagram at the SLC grid point with temperature (°C) and dewpoint (°C) in heavy solid lines. Short dashed line represents surface parcel ascent. Filled circle represents model lake temperature. Wind as in (a).
southwesterly, and a conditionally unstable lapse rate below ~750 hPa. The simulated sounding agreed well with the observed, although some minor differences were evident (cf. Figs. 3.2d and 4.3d). In particular, the observed layer of backing winds near 700 hPa was not found in the model sounding and the simulated surface temperature appeared to be too cold (-7.0 °C) compared to the observed temperature of -3.3 °C. The latter was a reflection of the elevation of the model terrain, which for domain 2 (18-km horizontal resolution) was 422 m (45 hPa) higher than the actual surface at the nearest grid point to SLC. At a given pressure level, the model temperature closely resembled the observed. The sounding derived from the higher resolution domain 4 simulation (not shown) had a near surface temperature of -4.3 °C just 34 m (4 hPa) higher than the actual surface.

At 1200 UTC 7 December, the simulated 500-hPa trough axis was located downstream of Utah and an upper-level ridge was building over the western United States (Fig. 4.4c). At 700 hPa, the coldest temperatures were located near northern Utah where northwesterly flow was found (Fig. 4.4b). In this region, the simulated relative humidity was slightly lower than analyzed by the RUC2 (Fig. 3.3b). At the surface, sea level pressure in the higher resolution MM5 (Fig. 4.4a), showed more mesoscale structure than the RUC2 (Fig. 3.3a), but there were no substantial differences in the placement of synoptic-scale features including the position of sea level pressure high that was centered over the Great Basin. Comparison of the simulated and observed soundings (cf. Figs. 3.3d and 4.4d) revealed a model warm bias between 500 and 700 hPa and cold bias near the surface, resulting in more stable low-level lapse rates than observed. It should be noted, however, that modification of the low-level temperature and dewpoint due to heat and
Figure 4.4. Analysis of the 18-km domain 2 MM5 simulation valid at 1200 UTC 7 December 1998. (a) Sea level pressure (every 2 hPa), near-surface winds (full barb denotes 5 ms\(^{-1}\)), and 12-h accumulated precipitation (mm, shaded according to scale at upper right). (b) 700–hPa temperature (every 2 °C), wind [as in (a)], and relative humidity (%), shaded following scale at upper right). Trough axes labelled by dashed lines. (c) 500–hPa geopotential height (every 60 m) and absolute vorticity \((x10^{-5} \text{ s}^{-1})\), shaded following scale at upper-right). (d) Skew-T, log-P diagram at the SLC grid point with temperature (°C) and dewpoint (°C) in heavy solid lines. Short dashed line represents surface parcel ascent. Filled circle represents model lake temperature. Wind as in (a).
moisture fluxes from the GSL was likely under-represented due to the relatively coarse 18 km resolution of this grid, since only 12 grid points represent the lake at this resolution.

By 0000 UTC 8 December the simulated 500 hPa ridge axis extended from southern California northeastward to central Montana and was just upstream of northern Utah (Fig. 4.5). This position was well forecast, although the amplitude of the simulated ridge was slightly less amplified than analyzed by the RUC2 (cf. Figs. 3.4c and 4.5c). At 700 hPa, the simulated flow remained northwesterly over northern Utah with temperatures rising to near -14˚C over SLC in response to low-level warm advection and middle-tropospheric subsidence (Fig. 4.5b). The MM5 sea level pressure forecast still featured more mesoscale structure than the coarser resolution RUC2 analysis, but correctly placed the synoptic-scale high pressure center over Utah (cf. Figs. 3.4a and 4.5a). The simulated SLC sounding showed veering winds with height at low-levels, implying warm advection, and an isothermal layer between 700 and 600 hPa (Fig. 4.5d). A weaker stable-layer was located between a shallow surface-based mixed layer and the base of the isothermal layer. These features captured the general character of the SLC sounding, although the static stability of the simulated isothermal layer was much weaker than the observed 5°C inversion (cf. Figs. 3.4d and 4.5d). Low-level temperatures were also ~3 °C colder than observed.

**Mesoscale Evolution**

This section uses output from domain 4 (2-km horizontal resolution) to examine the simulated mesoscale structure and evolution of this GSLE event. Fig. 4.6 presents the simulated low-level wind, lowest sigma-level temperature (~40 m AGL), and vertically-integrated precipitation (VIP)³ at 0400 UTC December 7. For model validation, surface
wind, temperature, and dewpoint observations from the Utah Mesonet are overlaid. At this time a region of low-level confluence was oriented along the western shoreline of the GSL. Model diagnostics at this and subsequent times showed this region of confluence was convergent and will hereafter be referred to as a convergence zone or line. VIP was located near the southern portion of this convergence zone and extended downstream along the eastern slopes of the Stansbury Mountains. Comparison with the corresponding WSR-88D radar reflectivity and mesonet analysis shows that this feature represented snowband A, which in the simulation appeared to be forming correctly near the western GSL shore but with the VIP region located south of the radar reflectivity band (cf. Figs. 3.5 and 4.6). This discrepancy could be due to model error, although it should be noted that VIP and radar reflectivity are not entirely consistent. Significant precipitation was also indicated in the radar analysis over the Tooele valley. Three weak VIP features were found in this region. Simulated low-level temperatures over the Great Salt Lake were above -4 °C, approximately 2 °C warmer than over surrounding regions of a similar elevation (Fig. 4.6). This model low-level temperature analysis agreed well with observed temperatures at most sites, with differences generally less than 2 °C (cf. Figs. 3.5 and 4.6). The most notable difference was over the Great Salt Lake Desert where the observed

3. The vertically-integrated precipitation is the total mass of parameterized rain and snow in a model column. It is used to illustrate the instantaneous position of the snowband at each analysis time.

4. Reflectivity calculated from model output was not used for comparison because of uncertainty in the elevation of the observed WSR-88D radar beam, which generally slopes upward at 0.5° and may be affected by refraction and beam spreading. Further uncertainties arise from the calculation of model-derived reflectivity. Thus, a qualitative comparison of the observed and modeled precipitation parameters in their most basic form is employed here.
Figure 4.5. Analysis of the 18-km domain 2 MM5 simulation valid at 0000 UTC 8 December 1998. (a) Sea level pressure (every 2 hPa), near-surface winds (full barb denotes 5 ms\(^{-1}\)), and 12-h accumulated precipitation (mm, shaded according to scale at upper right). (b) 700–hPa temperature (every 2 °C), wind [as in (a)], and relative humidity (%), shaded following scale at upper right). Trough axes labelled by dashed lines. (c) 500–hPa geopotential height (every 60 m) and absolute vorticity (x10\(^{-5}\) s\(^{-1}\), shaded following scale at upper-right). (d) Skew-T, log-P diagram at the SLC grid point with temperature (°C) and dewpoint (°C) in heavy solid lines. Short dashed line represents surface parcel ascent. Filled circle represents model lake temperature. Wind as in (a).
Figure 4.6. MM5 simulation on 2 km nest at 0400 UTC 7 December 1998. 0.995 sigma temperature (every 2 °C), vertically-integrated precipitation mixing ratio (kg m⁻², shaded following scale at upper right), and 10 m wind¹ (small wind barbs, full barb denotes 5 ms⁻¹). Observed wind, temperature (°C) and dewpoint (°C) denoted by station plots. Full barb denotes 5 ms⁻¹, temperature at upper left, and dewpoint at lower left. GSL shoreline denoted by dashed line. Snowband(s) denoted by large capital letter(s). Heavy dashed line represents axis of convergence.

¹. 10 m wind derived from lowest sigma (0.995) level wind. Extrapolation based on an assumed logarithmic profile with height.
(simulated) temperature was -3 °C (-6 to -8 °C). Wind directions and magnitudes near the convergence zone and over other regions were also in good agreement.

Over the next 90 min, simulated snowband A remained quasi-stationary and intensified. By 0530 UTC the VIP band extended from about the mid-point of the western GSL shoreline to the southeastern slopes of the Stansbury Mountains (Fig. 4.7). Meanwhile, the second snowband (snowband B) began to organize over eastern Tooele County and the western slopes of the Oquirrh Mountains. The forecast position of both snowbands was excellent, although they did not extend as far poleward as the corresponding radar reflectivity band (cf. Figs. 3.7 and 4.75). Wind and temperature observations at this time indicate that the model is in good general agreement with observations, although simulated temperatures were still too cold near the west boundary.

At 0630 UTC (Fig. 4.8), snowband A was becoming less organized and diminishing in precipitation intensity, although the convergence zone along the west shoreline was in nearly the same position and possessed a similar magnitude as at 0530 UTC. Although the VIP analysis did not suggest that snowband B was banded at this time, its associated cloud mass extended poleward toward Antelope Island in a well-organized band (not shown). The initial eastward movement of the simulated shoreline convergence zone and merger of snowbands A and B appeared to be slower than observed. (cf. Figs. 3.8 and 4.8). Temperatures in most locations, including the Tooele and Salt Lake valleys, were in good agreement, although the simulated temperatures in the western deserts had dropped to well below observed. The modeled wind field verified well against most land-based stations; however, the wind directions at Gunnison and Antelope Islands winds were

5. Since model output was not available at 0515 UTC, there is a 15-min difference between these two figures.
Figure 4.7. MM5 simulation on 2 km nest at 0530 UTC 7 December 1998. 0.995 sigma temperature (every 2 °C), vertically-integrated precipitation mixing ratio (kg m\(^{-2}\), shaded following scale at upper right), and 10 m wind (small wind barbs, full barb denotes 5 ms\(^{-1}\)). Observed wind, temperature (°C) and dewpoint (°C) denoted by station plots. Full barb denotes 5 ms\(^{-1}\), temperature at upper left, and dewpoint at lower left. GSL shoreline denoted by dashed line. Snowband(s) denoted by large capital letter(s). Heavy dashed line represents axis of convergence.
Figure 4.8. MM5 simulation on 2 km nest at 0630 UTC 7 December 1998. 0.995 sigma temperature (every 2 °C), vertically-integrated precipitation mixing ratio (kg m⁻², shaded following scale at upper right), and 10 m wind (small wind barbs, full barb denotes 5 ms⁻¹). Observed wind, temperature (°C) and dewpoint (°C) denoted by station plots. Full barb denotes 5 ms⁻¹, temperature at upper left, and dewpoint at lower left. GSL shoreline denoted by dashed line. Snowband(s) denoted by large capital letter(s). Heavy dashed line represents axis of convergence.
off by roughly 60 degrees due to the model placing the convergence zone too close to the west shoreline.

The simulated precipitation field at 0830 UTC (Fig. 4.9) was significantly different than observed (Fig. 3.9; 0815 UTC). At this time, observed snowbands A and B had merged into a solitary snowband that extended from Hat Island to the Oquirrh Mountains. The simulated snowbands, however, were in one of their least organized stages and were just beginning to merge (Fig. 4.9). Nevertheless, the simulated convergence zone was still evident, had moved offshore, and appeared to be well-positioned based on the observation from Hat Island.

Re-intensification and merger of simulated snowbands A and B occurred over the next few hours in a manner that was similar to observed but delayed. This is illustrated by the evolution of the VIP between 1000 and 1300 UTC (Fig. 4.10), which can be compared with radar reflectivity analyses presented in Figs. 3.7-3.10. This sequence illustrates some of the difficulties of mesoscale quantitative precipitation forecasting with existing modeling systems. Although surface winds and temperatures were generally well simulated, and the model forecast was reasonably accurate earlier in the period, errors related to the timing of the merger and propagation of the bands were still apparent.

At 1500 UTC the simulated convergence zone and snowband were aligned along the major axis of the GSL (Fig. 4.11). The overall flow pattern resembled that associated with mid-lake bands over Lake Michigan (e.g., Peace and Sykes 1966, Braham and Kelly 1982, Hjelmfelt 1990), with land breezes from the opposing lake shorelines converging near the lake axis. The largest simulated lake-land temperature differences were found at
Figure 4.9. MM5 simulation on 2 km nest at 0830 UTC 7 December 1998. 0.995 sigma temperature (every 2 °C), vertically-integrated precipitation mixing ratio (kg m⁻², shaded following scale at upper right), and 10 m wind (small wind barbs, full barb denotes 5 ms⁻¹). Observed wind, temperature (°C) and dewpoint (°C) denoted by station plots. Full barb denotes 5 ms⁻¹, temperature at upper left, and dewpoint at lower left. GSL shoreline denoted by dashed line. Snowband(s) denoted by large capital letter(s). Heavy dashed line represents axis of convergence.
Figure 4.10. Vertically-integrated precipitation mixing ratio (kg m\(^{-2}\)) from 2 km resolution MM5 simulation shaded following scale at upper right. (a) 1000 UTC 7 December 1998, (b) 1100 UTC 7 December 1998, (c) 1200 UTC 7 December 1998, and (d) 1300 UTC 7 December 1998.
Figure 4.11. MM5 simulation on 2 km nest at 1500 UTC 7 December 1998. 0.995 sigma temperature (every 2 °C), vertically-integrated precipitation mixing ratio (kg m⁻², shaded following scale at upper right), and 10 m wind (small wind barbs, full barb denotes 5 m s⁻¹). Observed wind, temperature (°C) and dewpoint (°C) denoted by station plots. Full barb denotes 5 m s⁻¹, temperature at upper left, and dewpoint at lower left. GSL shoreline denoted by dashed line. Snowband(s) denoted by large capital letter(s). Heavy dashed line represents axis of convergence.
this time with a narrow tongue of warm air (> -4 °C) oriented along the convergence zone axis.

Over land, a shallow nocturnal inversion had formed, most notably over the west desert, and surface temperatures were -10 °C or colder (Figs. 4.11 and 4.12). Temperatures were similar to observed except at the west desert observing point (S17; See Fig. 2.1 for location) where the simulated temperature was several degrees too low. The persistent model cold bias at this location may be related to errors in the specification of land-surface properties. The west desert land surface is composed primarily of salt flats, which at this time of year can be wet enough to form a salt slurry. Such a salt slurry would likely have a thermal inertia6 closer to water (.06 cal cm⁻² K⁻¹ s¹/²) than the desert land-surface that was specified in the model simulation (.02 cal cm⁻² K⁻¹ s¹/²). Since the areal coverage of the salt slurry is poorly known, it could not be accurately specified in the simulation. During the remainder of the simulation the area of precipitation drifted northeastward and weakened as low-level winds became southerly and southwesterly and conditions stabilized (Fig. 4.13; see also Figs. 4.4 and 4.5).

The total precipitation (liquid water equivalent) produced by the model simulation from 0000 to 1500 UTC 7 December 1998 is presented in Fig. 4.14. In comparison with Fig. 3.14, the model precipitation band stretching from just east of Stansbury Island into the eastern Tooele valley was very close to the observed position. Maximum simulated precipitation in this band was 19.3 mm, comparable but slightly higher than the observed

6. Thermal inertia, χ, is defined as:

\[ \chi = (\lambda C_s)^{1/2} \]

where \( \lambda \) is the thermal conductivity of the soil layer and \( C_s \) is the heat capacity per unit volume (Grell et al. 1995).
Figure 4.12. Skew-T, log-P diagram from 2km resolution MM5 simulation at 1500 UTC 7 December 1998. Temperature (°C) and dewpoint (°C) in heavy solid lines. Heavy dashed line represents surface parcel ascent. Data for model grid point nearest to Utah Mesonet site S17 in the Great Salt Lake Desert (see Fig. 2.1 for location).
Figure 4.13. MM5 simulation on 2 km nest at 2100 UTC 7 December 1998. 0.995 sigma temperature (every 2 °C), vertically-integrated precipitation mixing ratio (kg m⁻², shaded following scale at upper right), and 10 m wind (small wind barbs, full barb denotes 5 ms⁻¹). Observed wind, temperature (°C) and dewpoint (°C) denoted by station plots. Full barb denotes 5 ms⁻¹, temperature at upper left, and dewpoint at lower left. GSL shoreline denoted by dashed line.
Figure 4.14. Total precipitation (mm) produced by MM5 model simulation from 0000 to 1500 UTC 7 December 1998. Precipitation shaded according to scale at upper-right. Topographic contours shown every 500 m in solid lines (see Fig. 2.1 for elevations). Lake outline shown with dashed line.
maximum of 18.8 mm in Tooele. The simulation also captured the distribution of the precipitation just east of the Oquirrh Mountains. Based on radar analyses, the model appears to have over predicted precipitation over the western Tooele valley and Stansbury Mountains, although no surface snowfall measurements were available for direct validation. Simulated precipitation in this region occurred early in the model forecast, was shifted westward, and extended farther downstream than observed (cf. Figs. 3.7 and 4.6).

Model-Aided Diagnosis of Snowband Development and Evolution

Detailed analyses of the formative (0300 UTC) and mature (1300 UTC) stages of the snowband evolution are presented in Figs. 4.15-4.30. At 0300 UTC, the simulated snowband was located along the west shoreline of the GSL where there was low-level convergence between northerly flow over the GSL and northwesterly flow over land (Fig. 4.15a). The simulated low-level airflow and snowband position were similar to observed, although the modeled snowband was more developed and shifted slightly to the west of observed (not shown), and most closely resembled the observed structure approximately 2 h later at 0515 UTC (cf. Figs. 4.15a and b). Model surface analyses show a tongue of high potential temperature air and pressure troughing located just off of the west shore of the GSL (Fig. 4.16). Immediately west of these features, convergence was maximized along the strong gradient in potential temperature and pressure that was associated with a developing land-breeze front. A similar but weaker feature was located offshore of the east shoreline. Fig. 4.17 shows a cross-section of circulation vectors, virtual potential temperature, and cloud and precipitation mixing ratio taken perpendicular to the

7. Altimeter setting was used to reduce pressures to sea level values to avoid aliasing surface temperature gradients into pressure gradients.
Figure 4.15. Comparison of simulated snowband with observed. (a) Simulation at 0300 UTC 7 December 1998. 0.995 sigma temperature (every 2 °C), vertically-integrated precipitation (kg m$^2$, shaded according to scale), and 10 m wind (full barb = 5 ms$^{-1}$). (b) Lowest-elevation base-reflectivity analysis from KMTX WSR 88D (according to scale) and surface observations at 0515 UTC 7 December 1998. Observed wind, temperature (°C) and dewpoint (°C) denoted by station plots. Full barb = 5 ms$^{-1}$, temperature at upper-left, and dewpoint at lower-left.
Figure 4.16. MM5 simulation at 0300 UTC 7 December 1998. (a) 0.995 sigma potential temperature (every 0.2 °C). (b) Altimeter setting (every 0.025 hPa) and 0.995 sigma convergence (x 10³ s⁻¹) shaded according to scale at upper right.
Figure 4.17. Cross-section of 2km model simulation at 0300 UTC 7 December along line AB of Fig. 4.2. Winds in plane of cross-section according to scale at upper right. Solid lines denote virtual potential temperature every 2 °C. Total cloud (water + ice) mixing ratio (every 0.2 g kg⁻¹) denoted by dashed contours. Total precipitation (snow + rain) mixing ratio (g kg⁻¹) according to scale at upper left.
snowband at its formation point (Line AB in Fig. 4.2). Below 775 hPa the virtual potential temperature was nearly constant with height over the GSL, indicating weak static stability. Lower virtual potential temperature air was found over land to the west and east of the GSL. Approximately 6 km from the western shoreline low-level winds converged beneath the 5 Pa s\(^{-1}\) (50 cm s\(^{-1}\)) updraft that fed a shallow cloud band. Farther downstream along cross-section CD, a stronger updraft was evident and the cloud band was deeper (Fig. 4.18; see Fig. 4.2 for location). Precipitation fell in a narrow shaft that was approximately 10 km in width.

The surface pressure pattern evident in Fig. 4.16, which featured a mid-lake ridge separated by two near-shoreline troughs, appeared to be caused by the lake shoreline geometry. With northerly flow observed over most of the lake at this time (see Fig. 4.15a), high virtual potential temperature air and pressure troughing were located over the northern bays of the GSL, Gunnison and Bear River Bays. To the lee of Promontory Point, equatorward of an inflection in the lake shoreline, lower virtual potential temperatures and pressure ridging were found. Northerly flow moving over the Promontory Peninsula experienced a shorter over-water fetch and less flow modification than the flow to the west and east, resulting in the observed thermal and pressure structure (Fig. 4.16). Similarly, Phillips (1972) found that over-lake isotherms tended to parallel the up-wind coastline on Lake Ontario.

The strongest pressure gradient was located along the west shoreline and resulted in a substantial wind shift due to the down-gradient offshore acceleration of the low-level flow. The resulting convergence provided the necessary trigger for convection along the land-breeze front. Interestingly, the convergence line and precipitation band were
Figure 4.18. Cross-section of 2km model simulation at 0300 UTC 7 December along line CD of Fig. 4.2. Winds in plane of cross-section according to scale at upper right. Solid lines denote virtual potential temperature every 2 °C. Total cloud (water + ice) mixing ratio (every 0.2 g kg⁻¹) denoted by dashed contours. Total precipitation (snow + rain) mixing ratio (g kg⁻¹) according to scale at upper left.
coincident with the pressure gradient, rather than the pressure trough. This might be expected, however, since the air-flow over the lake would tend to keep the wind speed and direction nearly constant until significant forcing such as this gradient could produce some deflection. In a simulation of a shoreline snowband over Lake Michigan, Hjelmfelt and Braham (1983) also obtained a wind field that had the maximum convergence between the pressure minimum and the shoreline.

Three-dimensional trajectories, beginning at 1800 UTC 6 December and terminating on the lowest half-sigma level (approximately 40 m AGL) at 0300 UTC 7 December further elucidate the processes associated with snowband initiation (Fig. 4.19). Trajectories terminating in a line that was roughly perpendicular to the snowband demonstrate that the warmest low-level temperatures were associated with trajectories that experienced the greatest over-water fetch (Fig. 4.19a). Low-level convergence into the snowband was evident with a strong deflection of trajectories ending immediately west of the snowband (Trajectories in Fig. 4.19a remained near the model surface while approaching the convergence zone). This is more clearly illustrated by Fig. 4.19b which shows two lines of trajectories terminating to the west and east of the snowband, respectively. Near the west shore of the GSL, trajectories were deflected eastward by the strong offshore pressure gradient. East of the snowband, the flow was primarily meridional.

Boundary-layer modification by the GSL is illustrated by a series of soundings taken along trajectory 9 of Fig. 4.19a (Fig. 4.20). At point A (0000 UTC), where trajectory

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8. Because of the large amount of computer hard disk space required to store model output at higher temporal resolution, most trajectories presented in this dissertation were calculated using 30-min model output. This time difference was selected since trajectories calculated from 5- and 30-min output were not significantly different.
Figure 4.19. 9-h three-dimensional trajectories ending at 0300 UTC 7 December on the model’s lowest sigma level (sigma=.995). 0.995 Sigma level potential temperature (every 0.2 °C) shown only near lake-level. Terrain heights above 1500 m shaded in grey. (Note: some trajectories exit domain during 9 h.) (a) Cross-section perpendicular to snowband. Parcel locations at 30-min intervals along trajectory 9 denoted with large dots. Trajectory 9 parcel locations at 0000, 0100, 0200, and 0300 UTC denoted by letters a, b, c, and d respectively. Soundings from points a, b, c, and d shown in Fig. 4.20. (b) Two cross-sections parallel to snowband.
Figure 4.20. Soundings taken along trajectory 9 in Fig. 4.19a. (a) 0000 UTC at point a, (b) 0100 UTC at point b, (c) 0200 UTC at point c, and (d) 0300 UTC at end of trajectory (point d).
first moves near the GSL, the sounding featured a shallow surface-based mixed layer with a near-surface temperature of -4.8 °C and dewpoint of -9.0 °C (Fig. 4.20a). After limited interaction with the northern GSL and passing temporarily over land, the sounding at point B (0100 UTC) showed little change (Fig. 4.20b). The remainder of the trajectory path was over the GSL. One hour later, when the trajectory reached point C, the near-surface temperature had increased to -3.2 °C while the near-surface dewpoint has decreased slightly to -9.1 °C (Fig. 4.20c). The reduction in dewpoint may be surprising since the trajectory was located over the lake surface, but occurred for the following reasons. First, lower specific-humidity air from aloft was entrained into the boundary-layer as it grew 25-50 hPa in depth due to surface heating (Fig. 4.21b). At the same time, surface moisture fluxes were much lower than would have occurred over a body of fresh water because of the high salinity of the north arm of the GSL. This effect is illustrated by the surface latent heat flux analysis for 0200 UTC (Fig. 4.21a). North of the railroad causeway, where a 30% reduction in the fresh water saturation vapor pressure was specified in the simulation due to the observed 27% salinity, latent heat fluxes were reduced significantly, with the peak flux reaching 110 W m⁻². Over the next hour, however, trajectory 9 was almost entirely over the lower salinity (9%) south arm where the reduction in fresh water saturation vapor pressure was only 6% and latent heat fluxes reached over 170 W m⁻². By 0300 UTC (point D), the near-surface temperature along trajectory D had increased to -2.7 °C and the dewpoint decreased slightly to -9.3 °C (Fig. 4.20c).

9. The moisture flux is related to latent heat flux by the following relation:

\[ M = \frac{L}{L_v} \]

where \( M \) is the surface moisture flux, \( L \) is the surface latent heat flux, and \( L_v \) is the latent heat of vaporization of water (\( L_v = 2.501 \times 10^6 \) J kg⁻¹ at 0 °C).
Figure 4.21. MM5 simulation at 0200 UTC 7 December 1998. (a) Surface latent heat flux (every 10 W m$^{-2}$). (b) Surface sensible heat flux (every 10 W m$^{-2}$).
Concurrently, snowband A began to develop, as inferred from the moist and conditionally unstable temperature profile between 800 and 750 hPa. Though the boundary layer moistening appears to be negligible for the last hour of this trajectory, it is possible that moisture was transported upward and out aloft, following a circulation similar to that seen in Fig. 4.17 and 4.18. This series of soundings shows that localized heating by the lake warmed, deepened, and further destabilized the boundary layer. Though near-surface dewpoints remained nearly constant over the lake, it is apparent that moistening of the boundary layer was occurring despite reduced moisture fluxes due to high lake salinity, though this moistening occurred through the lowest ~100 hPa of the well-mixed boundary layer.

By 1300 UTC the convergence zone and snowband were located near the mid-lake axis and extended downstream into the Tooele valley (Fig. 4.22). As noted previously, the simulated eastward movement of the snowband was delayed, so that the snowband structure at this time was similar the observed structure at 0815 UTC (cf. Fig. 4.22a,b). The simulated low-level potential temperature pattern over the lake was dominated by a tongue of warm air that extended down the mid-lake axis and was flanked by intense temperature gradients near the eastern and western shorelines (Fig. 4.23a). Intensification of the lake-land temperature gradient since 0300 UTC (cf. Figs. 4.16a and 4.23a) was due primarily to nocturnal cooling over the surrounding land mass (e.g., Fig. 4.12), particularly over the west desert. The offshore pressure gradient has also increased with general pressure troughing located along the warm tongue near the mid-lake axis (Fig. 4.23b). As a result, the west-shore land breeze and the magnitude of the offshore flow near
Figure 4.22. (a) Comparison of simulated snowband with observed. (a) Simulation at 1300 UTC 7 December 1998. 0.995 sigma temperature (every 2 °C), vertically-integrated precipitation (kg m⁻², shaded according to scale), and 10 m wind (full barb = 5 ms⁻¹). (b) Lowest-elevation base-reflectivity analysis from KMTX WSR 88D (according to scale) and surface observations at 0815 UTC 7 December 1998. Observed wind, temperature (°C) and dewpoint (°C) denoted by station plots. Full barb = 5 ms⁻¹, temperature at upper-left, and dewpoint at lower-left.
Figure 4.23. MM5 simulation at 1300 UTC 7 December 1998. (a) 0.995 sigma potential temperature (every 0.2 °C). (b) Altimeter setting (every 0.025 hPa) and 0.995 sigma convergence (x 10^3 s^{-1}) shaded according to scale at upper right.
the east shoreline had intensified, resulting in a well-defined convergence line near the mid-lake axis (Fig. 4.22a).10

The mesoscale circulations responsible for the snowband are further illustrated by the cross-section of virtual potential temperature, circulation vectors, and cloud and precipitation mixing ratio presented in Fig. 4.24 (see Fig. 4.2 for location). Compared to 10 h previously (Fig. 4.17), near-surface virtual potential temperatures to the west (east) of the GSL have decreased by 6-8 °C (2-3 °C) due to nocturnal cooling, resulting in the development of a more dense and stable boundary layer surrounding the GSL. Meanwhile, virtual potential temperatures over the GSL have remained relatively constant. The resulting contrast in boundary-layer temperature and surface pressure (see also Fig. 4.23) had intensified the offshore flow, particularly from the western shoreline. Where the two opposing flows met near the mid-lake axis, a narrow 6 Pa s⁻¹ (~60 cm s⁻¹) updraft was found beneath the developing cloud and precipitation band. Interestingly, the convergence zone and updraft are again not located directly over the pressure trough at this time, similar to 10 h previously, suggesting that the location of the updraft and convergence zone may not be directly related to the position of the pressure trough (Figs. 4.23, 4.24). The overall three-dimensional wind and thermal structure closely resembles that associated with mid-lake bands over Lakes Michigan and Ontario (e.g., Peace and Sykes 1966, Passarelli and Braham 1981, Braham and Kelly 1982, Hjelmfelt 1990).

A cross section along the center of the snowband at 1300 UTC is presented in Fig 4.25. Equivalent potential temperature at low levels increased from northwest to southeast as the flow crossed over the GSL, indicating reduced stability over the downwind region of

10 Latent heating in the snowband formation region may also enhance the strength of the mid-lake pressure trough and convergence zone.
Figure 4.24. Cross-section of 2km model simulation at 1300 UTC 7 December along line AB of Fig. 4.2. Winds in plane of cross-section according to scale at upper right. Solid lines denote virtual potential temperature every 2 °C. Total cloud (water + ice) mixing ratio (every 0.2 g kg⁻¹) denoted by dashed contours. Total precipitation (snow + rain) mixing ratio (g kg⁻¹) according to scale at upper left. Lake shorelines denoted with thick vertical lines.
Figure 4.25. Cross-section of 2km model simulation at 1300 UTC 7 December along line EF of Fig. 4.2. Winds in plane of cross-section according to scale at upper right. Solid lines denote virtual potential temperature every 2 °C. Total cloud (water + ice) mixing ratio (every 0.2 g kg⁻¹) denoted by dashed contours. Total precipitation (snow + rain) mixing ratio (g kg⁻¹) according to scale at upper left. Lake shorelines denoted with thick vertical lines.
the GSL. Over the downwind half of the GSL, updrafts formed which produced progressively larger cloud and precipitation mixing ratios as the flow neared the downwind shoreline. The updraft reached a maximum elevation of ~650 hPa near the downwind shoreline. Little orographic uplift was evident in this cross-section, though it was seen at other times (not shown). Vertical velocities were maximized near the downwind shoreline. Similarly, Lin and Smith (1986), using a two-dimensional linearized analytical model, found that a heat island produced maximum upward displacements on its downwind side due to advection by the mean wind. Additionally, a dome of more dense air was found over the Tooele valley at this time. It is possible that this dome of dense air also played a role in lifting the more buoyant lake-modified air was forced over the dense air trapped in the Tooele valley. Carpenter (1993) suggested that this mechanism may be partially responsible for GSL lake-effect snowstorms.

Three-dimensional trajectories, beginning at 0400 UTC 7 December and terminating on the lowest half-sigma level (approximately 40 m AGL) at 1300 UTC 7 December (Fig. 4.26) show stronger lake-induced circulations than at 0300 UTC (see trajectories 7-10 in Fig. 4.26a). Though the snowband is aligned with the longest axis of the lake, trajectories in Fig. 4.26b actually had a shorter over-water fetch than those terminating around the snowband 10 h earlier (Fig. 4.19) because of the stronger offshore flow. Boundary layer modification along a trajectory that had a relatively long over-water fetch and terminated near the warm tongue (Fig. 4.26b) is illustrated by the soundings in Fig. 4.27. At 1000 UTC, this trajectory was approaching the northern shoreline of the Bear River Bay (point a, Fig. 4.26b). At this location, a shallow surface inversion, presumably produced by nocturnal cooling, was located near the surface and another shallow stable
Figure 4.26. 9-h three-dimensional trajectories ending at 1300 UTC 7December on the model’s lowest sigma level (sigma=.995). 0.995 Sigma level potential temperature (every 0.2 °C) shown only near lake-level. Terrain heights above 1500 m shaded in grey. Parcel locations at 30-min intervals along sample trajectory (b) denoted with large dots. Trajectory parcel locations at 1000, 1100, 1200, and 1300 UTC denoted by letters a, b, c, and d respectively. Soundings from points a, b, c, and d shown in Fig. 4.27. (Note: some trajectories exit domain during 9 h.)
Figure 4.27. Soundings taken along trajectory ending at point d in Fig. 4.26b. (a) 1000 UTC at point a, (b) 1100 UTC at point b, (c) 1200 UTC at point c, and (d) 1300 UTC at end of trajectory (point d).
layer was located near 800 hPa (Fig. 4.27b). The lowest sigma-level temperature and
dewpoint were -8.9 °C and -10.4 °C, respectively. One hour later (1100 UTC), the
trajectory was located over the Bear River Bay, and had been over water for ~30 min
(point b, Fig. 4.26b). At this point, sensible heating over the lake surface had raised the
near-surface temperature to -6.3 °C and a shallow surface-based mixed layer had
developed. Near surface moisture increased slightly with the dewpoint reaching -9.8 °C, as
would be expected due to the reduced moisture fluxes associated with high salinity and
due to the limited time that the parcel was resident over this portion of the lake (Figs. 4.26b, 4.28a). The stable-layer near 800 hPa was still present and had lowered slightly
(Fig. 4.27b).

By 1200 UTC, the parcel had reached point c, which was located over the less
saline southern region of the GSL (Fig. 4.26a). The near surface temperature and dewpoint
had risen to -4.7 °C and -9.5 °C, respectively, and the stable layer near 875 hPa was
weakened (Fig. 4.27c). During the next hour, the parcel temperature and dewpoint
increased further to -3.8 °C and -8.4 °C, respectively, all stable layers and inversions below
~550 hPa had eroded away, and the sounding was conditionally unstable up to ~675 hPa
(Fig. 4.27d). Similar to the analysis of the boundary layer modification along a trajectory
at 0300 UTC, the impact of the lake on boundary-layer temperature is much greater than
on moisture, suggesting that the primary impact of the GSL is to destabilize the boundary
layer and organize convection through surface sensible heating (Fig.4.28b) and the
generation of thermally-driven circulations and low-level convergence. Moisture fluxes,
which were significantly larger over the southern half of the GSL, may not be necessary
for the development of the snowband, but may act to increase the total precipitation. The
Figure 4.28. MM5 simulation at 1200 UTC 7 December 1998. (a) Surface latent heat flux (every 10 W m$^{-2}$). (b) Surface sensible heat flux (every 10 W m$^{-2}$).
amount of low-level moisture in the upstream sounding thus becomes an important variable for lake-effect events. This topic will be explored further in Chapter 5.

As an example of the vertical circulations associated with the snowband at this time, three-dimensional trajectories that were ingested into the snowband are presented in (Fig. 4.29). These trajectories began at 1000 UTC 7 December and terminated at 1300 UTC 7 December on the 0.855 sigma level (761 hPa), which was near the top of the snowband. Trajectories 1–3 originated at low levels to the west of the GSL, converged toward the snowband axis, then rose rapidly and became diffluent aloft. Trajectory 1, which was the outer-most trajectory west of the snowband at 1300 UTC, was the innermost at the beginning of the trajectory. As it approached the low-level convergence zone, it was advected upward through the snowband. Trajectories 2 and 3 originated further to the south and remained near the surface until they rose rapidly when they were ingested in the narrow updraft that supported the snowband (e.g., Fig. 4.24). Trajectories ending in the eastern portion of the snowband followed a slightly different evolution. Trajectories 4–5 originated north of the GSL ~80 hPa above lake level, eventually descended to low levels as they converged toward the snowband, then rose rapidly as they were ingested into the snowband. Trajectory 6 was apparently ingested into the snowband near its point of formation, rose to 757 hPa, remained aloft and moved slightly away from the snowband, descended slightly to 803 hPa and converged again toward the snowband and rose to 761 hPa near the snowband. The evolution of pressure, temperature, water-vapor mixing ratio, and cloud water and ice mixing ratio along trajectory 1 are displayed in Fig. 4.30. At 1000 UTC, trajectory 1 was located west of the shoreline of the GSL at 849 hPa (Fig. 4.30). It descended to 875 hPa (~5 hPa above lake level) as it followed
Figure 4.29. 3-h three-dimensional trajectories ending at 1300 UTC 7 December on sigma level 0.855 (761 hPa). Pressure at initial parcel location denoted on figure. Trajectory line-width inversely proportional to parcel pressure. Terrain heights above 1500 m shaded in grey. Trajectory 1 parcel positions at 30 min intervals labeled.
Figure 4.30. Air parcel properties along trajectory 1 in Fig. 4.29. (a) pressure (hPa). (b) Potential Temperature (K). (c) Mixing ratio (g kg\(^{-1}\)). (d) Cloud water+ice mixing ratio (g kg\(^{-1}\)).
downward sloping terrain toward the lake shoreline after 1030 UTC. Potential temperature and water vapor mixing ratio increased 3.4 K and 0.4 g kg\(^{-1}\) from 1100 to 1130 UTC as the trajectory moved over the GSL (Fig. 4.30b,c). As the trajectory approached the low-level convergence zone and associated updraft prior to 1230 UTC, it began to rise upward with saturation and cloud water development beginning at around 1220 UTC. At approximately 1240 UTC, the trajectory reached its highest elevation and the cloud-water mixing ratio was near its peak value. The trajectory then sank gradually as it exited the snowband (Fig. 4.30a).

The physical picture obtained from the analysis above is consistent with the findings of studies over the Great Lakes that illustrate the role of thermally-driven land-breeze circulations in generating solitary snowbands (e.g., Passarelli and Braham 1981, Hjelmfelt and Braham 1983, Hjelmfelt 1990). In the present case, pressure troughing over the GSL induced by surface sensible and latent heat fluxes from the warm lake led to the formation of over-lake low-level wind convergence. A narrow linear convergence zone developed along the boundary between a developing land breeze near the west shoreline and the synoptic northerly flow near the surface. Forced by this convergence zone, a band of convective clouds and precipitation formed roughly parallel to the northwesterly steering-layer flow, aided by boundary layer heating, moistening, and destabilization over the lake. The cloud band, fed by heat and moisture fluxes over the lake, was sustained in relative equilibrium while cloud elements within the band were advected downstream over land into the Tooele valley. As nocturnal cooling increased the temperature contrasts across the GSL shoreline, land-breezes from both the west and east shorelines intensified, increasing the strength of the over-lake convergence and producing the most organized,
mature stage of the snowband. Once the snowband was established, latent heat release due to condensation in clouds may have been important for strengthening and sustaining the established circulations (e.g., Ballentine 1982, Hjelmfelt and Braham 1983, Hjelmfelt 1990).

It was also seen that shoreline geometry influenced the pattern of temperature and pressure over the lake, similar to observations over Lake Ontario (Phillips 1972). The warmest temperatures and lowest pressures were associated with the longest over-water trajectories. Moisture fluxes over the lake were affected by the influence of the high salinity in the lake. In fact, near-surface dewpoints remained nearly constant or even decreased slightly over the northern GSL due to reduced moisture fluxes in that region, suggesting that the primary role of the lake was to warm and destabilize the boundary layer, while moisture fluxes were less significant than they would be over a freshwater lake. As a result, the significance of upstream moisture is highlighted. The effects of surface fluxes of heat and moisture, salinity, upstream moisture, latent heat release due to condensation, as well as the roles of topography, surface friction and lake temperature will be examined further in Chapter 5.
CHAPTER 5

MODEL SENSITIVITY EXPERIMENTS

To more fully understand the effects of various physical processes and environmental characteristics, such as surface heat and moisture fluxes, orographic effects, surface friction, lake temperature, and upstream moisture, a series of model sensitivity experiments were conducted. The sensitivity experiments were run using the 2 km resolution domain 4 with one-way boundary conditions provided by domain 3 of the control simulation (hereafter CTL). The model parameterizations, topography, and initial and boundary conditions were the same as in CTL except for the changes described in each sensitivity experiment. These experiments are summarized in Table 5.1.

Effects of Surface Fluxes and Salinity

Surface fluxes of heat and moisture over relatively warm water have been shown to be important physical mechanisms leading to lake-effect snow in the Great Lakes region (e.g., Peace and Sykes 1966, Lavoie 1972, Passarelli and Braham 1981, Hjelmfelt 1990). Surface heat fluxes warm and destabilize the boundary layer while creating low pressure over the lake which drives land breezes and over-lake convergence. Moisture fluxes provide latent heat which may be released in convective precipitation processes driven by the destabilization and forced convection resulting from these sensible heat fluxes. This latent heat release in turn strengthens convection (e.g., Lavoie 1972, Hjelmfelt 1990). As
Table 5.1. Summary of sensitivity experiments. Precipitation amounts refer to total accumulation from 0000 to 1500 UTC 7 December 1998.

<table>
<thead>
<tr>
<th>Experiment Category</th>
<th>Experiment</th>
<th>Description</th>
<th>Domain Max. Precip. (mm)</th>
<th>Domain Avg. Precip. (mm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Control run</td>
<td>CTL</td>
<td>Full physics simulation</td>
<td>19.3</td>
<td>0.59</td>
</tr>
<tr>
<td>Surface fluxes and salinity</td>
<td></td>
<td>FRESH</td>
<td>Freshwater GSL</td>
<td>23.1</td>
</tr>
<tr>
<td>&quot;</td>
<td>NOLHFLX</td>
<td>No surface latent heat flux</td>
<td>9.0</td>
<td>0.30</td>
</tr>
<tr>
<td>&quot;</td>
<td>NOSHFLX</td>
<td>No surface sensible heat flux</td>
<td>7.2</td>
<td>0.42</td>
</tr>
<tr>
<td>Latent heat release</td>
<td>NOLHR</td>
<td>No latent heat release in condensation</td>
<td>10.7</td>
<td>0.49</td>
</tr>
<tr>
<td>Topography</td>
<td>FLAT</td>
<td>Flat lake-level topography</td>
<td>11.1</td>
<td>0.55</td>
</tr>
<tr>
<td>&quot;</td>
<td>NODST</td>
<td>Flat topography downstream of GSL only</td>
<td>10.8</td>
<td>0.51</td>
</tr>
<tr>
<td>Shoreline roughness contrasts</td>
<td>ROUGH1</td>
<td>Roughness length 10 cm over entire domain</td>
<td>65.1</td>
<td>1.39</td>
</tr>
<tr>
<td>&quot;</td>
<td>ROUGH2</td>
<td>Roughness length 10 cm over entire domain for surface momentum flux calculation only</td>
<td>10.1</td>
<td>0.53</td>
</tr>
<tr>
<td>Lake temperature</td>
<td>T+2</td>
<td>Lake temperature 280 K</td>
<td>34.7</td>
<td>0.78</td>
</tr>
<tr>
<td>&quot;</td>
<td>T-2</td>
<td>Lake temperature 276 K</td>
<td>12.0</td>
<td>0.45</td>
</tr>
<tr>
<td>Upstream Moisture</td>
<td>RH+10</td>
<td>Relative humidity increased by 10% in initial and boundary conditions</td>
<td>27.0</td>
<td>1.13</td>
</tr>
<tr>
<td>&quot;</td>
<td>RH-10</td>
<td>Relative humidity decreased by 10% in initial and boundary conditions</td>
<td>13.8</td>
<td>0.27</td>
</tr>
</tbody>
</table>
discussed in Chapter 2, the Great Salt Lake (GSL), unlike the Great Lakes, is a saline body of water which has a lower saturation vapor pressure than for fresh water, thus moisture fluxes are reduced or eliminated from what they would be for fresh water. Furthermore, the relatively small size of the GSL likely decreases the GSL’s ability to provide moisture in lake-effect events.

The effects of localized surface fluxes over the Great Salt Lake (GSL), as well as the role of the saline composition of the GSL, were examined using three sensitivity experiments in addition to the control simulation described in chapter 4. The first experiment assumed the GSL consisted of fresh water (FRESH), neglecting the effects of salinity on saturation vapor pressure that were accounted for in CTL. In the second experiment surface latent heat fluxes over the lake were neglected (NOLHFLX). The third simulation eliminated sensible heat fluxes over the GSL (NOSHFLX) and was conducted to determine whether moisture fluxes alone are sufficient to produce GSLE snow, and to what degree localized heating influences mesoscale circulations around the GSL.

Fig. 5.1 presents the total precipitation and average wind fields for 0000–1500 UTC from CTL, FRESH, NOLHFLX, and NOSHFLX\(^1\). FRESH produced 20% more domain-averaged precipitation than CTL (cf. Figs 5.1a,b). This difference was due to enhanced precipitation in FRESH during the stage of the simulation where the snowband was resident near the west shoreline of the GSL, as well as a tendency for the snowband to extend farther upstream during most of the simulation. The FRESH experiment shows that lake salinity has an effect on precipitation in this case and implies that a fresh water GSL might provide enough moisture to affect GSLE precipitation significantly. The averaged

\(^1\) Unless otherwise noted, meteorological fields described in this chapter are averaged from 0000 to 1500 UTC 7 December 1999.
Figure 5.1. Two-km resolution MM5 simulations from 0000 to 1500 UTC 7 December 1998. Total accumulated precipitation (every 2 mm) in solid contours and vector averaged 10-m wind (full barb = 5 m s$^{-1}$). Model terrain shaded according to scale at upper right. (a) CTL, (b) FRESH, (c) NOLHFLX, (d) NOSHFLX.
wind field was generally not changed much by the fresh water treatment of the GSL surface, except near the west shoreline where the most pronounced additional precipitation occurred in the fresh water simulation (cf. Fig. 5.1a, 5.1b). Fig. 5.2a shows the difference of low-level winds and altimeter setting between these two simulations. The vector difference field shows stronger convergence near the west shoreline of the GSL in FRESH. Stronger pressure troughing also occurred in this region, though it is not evident in Fig. 5.2a. This was likely produced by stronger latent heat release in FRESH due to the additional precipitation (Lavoie 1972, Hjelmfelt and Braham 1983, Hjelmfelt 1990).

NOLHFLX produced 49% less domain-average precipitation than the control run. Precipitation patterns remained similar to the control run throughout the simulation, though precipitation amounts were consistently below the values produced by the control run and precipitation bands did not extend as far upwind at times. This suggests that the GSL, though highly saline, still provided a significant source of moisture in this event. Though the wind field is apparently unchanged in this experiment (c.f. Figs. 5.1a, 5.1c), a difference plot reveals that the absence of latent-heat fluxes from the GSL has produced a slightly more divergent wind flow over the southern half of the GSL and the Salt Lake and Tooele valleys as well as higher pressure in these regions. Again the likely cause of these changes to the pressure and wind fields is the reduced amount of latent heat release from precipitation processes in NOLHFLX (Fig. 5.2b).

NOSHFLX produced 28% less precipitation averaged over the entire grid than CTL, which was notably more than NOLHFLX. Though the average precipitation amount in this simulation was similar to CTL, the precipitation patterns differed considerably. Precipitation fell in organized banded structures in CTL but in NOSHFLX only had a
tendency to fall over the Oquirrh and Stansbury Mountains and other orographically favored locations, as well as over the southern GSL. Banded precipitation structures occurred infrequently and for short time periods. The wind flows were significantly altered in this experiment, most notably over and near the GSL. Though wind directions over the GSL were similar, wind speeds over the GSL were generally slower and thus less convergent (c.f. Figs. 5.1a, 5.1d). Fig. 5.2c shows that the absence of surface sensible heat fluxes had a strong impact on low-level wind circulations over and around the GSL. This impact, however, likely was magnified slightly by reduced latent heat release in the atmosphere. Thus, surface sensible heat fluxes appear to be important for organizing convection over and downwind of the GSL, and their effects may be magnified by convective feedbacks due to latent heat release in precipitation processes. It should also be noted that the effect of sensible heating at the surface in the model is the combination of cooling over land and warming over water (see Figs. 4.21, 4.28), but only sensible heating over water was turned off here for lateral boundary consistency in the model. The effects of turning off sensible heat fluxes in the model would likely be somewhat larger if it was done for the entire domain.

**Effects of Latent Heat Release**

Once initiated due to localized heating or moistening over the GSL, latent heat release due to condensation in convection may form a positive feedback mechanism that may intensify snowbands by strengthening convective circulations (Ballentine 1982, Hjelmfelt and Braham 1983, Hjelmfelt 1990). To examine the effects of latent heat release, a simulation was run which eliminated latent heat release in the atmosphere, though precipitation processes and surface moisture fluxes still occurred (NOLHR). Fig.
Figure 5.2. Two-km resolution MM5 simulations averaged from 0000 to 1500 UTC 7 December 1998. Difference fields of 10-m wind (scale at lower left) and altimeter setting (every .1 hPa, zero contour not shown). Model terrain shaded according to scale at upper right. (a) FRESH minus CTL, (b) NOLHFLX minus CTL, and (c) NOSHFLX minus CTL.
Figure 5.3. Two-km resolution MM5 simulations from 0000 to 1500 UTC 7 December 1998. Total accumulated precipitation (every 2 mm) in solid contours and vector averaged 10-m wind (full barb = 5 m s$^{-1}$). Model terrain shaded according to scale at upper right. (a) CTL and (b) NOLHR.
5.3 contrasts CTL with NOLHR, and shows that precipitation was considerably lower near all three precipitation maxima, though the total precipitation averaged over the grid was only reduced by 18%. This simulation had similar near surface wind flow characteristics and precipitation patterns, though precipitation bands were of much lower intensity, as the velocity and depth of updrafts was reduced due to the absence of latent heating. Fig 5.4 shows an example of the effect of latent heating on updraft organization, intensity, and depth. The resulting precipitation mixing ratio and updraft height were significantly reduced in NOLHR. Wind fields in NOLHR were generally similar to CTL, though weaker convergence occurred near regions where precipitation fell (Figs. 5.3, 5.5). Thus, latent heat release likely plays an important role in strengthening horizontal and vertical convective circulations induced by surface heating, orographic uplift, and other processes.

Effects of Topography

Many studies have noted the effects of orographic uplift on lake-effect snowfall in the Great lakes region (e.g. Hjelmfelt 1992, Muller 1966, Niziol et al. 1995). As described in Chapter 2, the GSL is surrounded by mountains several times higher than those surrounding the Great lakes. In order to examine the effects of local topography on GSLE precipitation processes, two sensitivity experiments were run in which topographical features surrounding the GSL were changed. In the first experiment (FLAT), the surface elevation was set to 1279.5 m, the same elevation as the GSL, except near the lateral boundaries where the terrain matched that of the mother domain (domain 3). Five grid points near the lateral boundaries of the nest were left unchanged, and the flat topography in the interior was blended with these grid points to avoid creating any steep slopes. The
Figure 5.4. Cross-sections of model simulation at 1900 UTC 7 December along line CD of Fig. 4.2. Winds in plane of cross-section according to scale at upper right. Solid lines denote virtual potential temperature every 2 °C. Total cloud (water + ice) mixing ratio (every 0.2 g kg⁻¹), denoted by dashed contours. Total precipitation (snow + rain) mixing ratio (g kg⁻¹) shaded according to scale at upper left. (a) CTL (b) NOLHR.
Figure 5.5. Two-km resolution MM5 simulation averaged from 0000 to 1500 UTC 7 December 1998. Difference fields of 10-m wind (scale at lower left) and altimeter setting (every .1 hPa, zero contour not shown). Model terrain shaded according to scale at upper right. (a) NOLHR minus CTL.
second experiment was run with the Oquirrh, Stansbury, and Cedar Mountains, which form the downstream topography in this case, removed from the terrain (NODST).

Fig 5.6 contrasts CTL with the two sensitivity experiments. FLAT produced only 7% less domain-average precipitation averaged than CTL, but the precipitation distribution was changed considerably. Instead of a narrow band of precipitation with a 19.3 mm maximum near Tooele, as was produced by CTL, a more broad region of moderate accumulations with a maximum of 11.1 mm was found north of Tooele. The band of precipitation which fell downwind from the west shoreline of the GSL also featured a reduced precipitation maximum in FLAT, though its distribution was comparable to CTL. Interestingly, 8 mm of precipitation fell near the northeast portion of Stansbury island in FLAT, while only 2-4 mm fell in CTL. Most of this precipitation in FLAT, however, fell between 12 and 15 UTC when the precipitation band pivoted near Stansbury Island, rather than moving eastward as occurred in CTL (cf. Fig 5.7 a,b). As shown in Figs 5.7d,e, total precipitation during this time period was similar between these two simulations, however the heaviest precipitation in CTL fell near the south GSL shoreline, though it was located near Stansbury Island in the flat topography run. These results are due mainly to subtle differences in snowband position influenced by topography, though not directly related to orographic uplift, and illustrate that local snowfall amounts are related not only to snowfall rate, but also to the residence time of the snowband over a localized region.

Also shown in Fig. 5.6 is NODST. This simulation produced precipitation amounts similar to FLAT, though the grid averaged precipitation was 14% less than CTL (slightly less than FLAT). The maximum precipitation in this experiment occurred near the city of
Figure 5.6. Two-km resolution MM5 simulations from 0000 to 1500 UTC 7 December 1998. Total accumulated precipitation (every 2 mm) in solid contours and vector averaged 10-m wind (full barb = 5 m s\(^{-1}\)). Model terrain shaded according to scale at upper right. (a) CTL, (b) FLAT, (c) NODST.
Figure 5.7. Depiction of snowband evolution from 1200 UTC to 1500 UTC 7 December 1998. (a) - (c): Axis of snowband at 1-h intervals from 1200 UTC to 1500 UTC. Model terrain shaded according to scale at upper right. (a) CTL, (b) FLAT, (c) NODST. (d) - (f): Accumulated precipitation from 1200 to 1500 UTC. (d) CTL, (e) FLAT, (f) NODST.
Tooele, as in CTL, though a maximum of 10.8 mm fell there, roughly half as much as in CTL, indicating significant enhancement of precipitation in Tooele due to orographic ascent. Comparison of averaged winds from CTL and NODST show the effects of nocturnal drainage flows and topographic blocking. Drainage flows over the Stansbury and Cedar Mountains can be seen in CTL, while light west-northwesterly winds occurred in NODST. Furthermore, winds tended to be more northerly in the valleys, paralleling the mountain ranges, in CTL, while this behavior did not occur in NODST.

The results of these topography experiments show that orographic uplift and channeling were not directly responsible for snowband generation. The lake-induced circulations caused by localized surface energy fluxes were sufficient since FLAT and NODST both produced a west shore and a midlake snowband. However, orographic enhancement of precipitation totals was found in some regions, such as near the city of Tooele. Finally, topography only tended to increase total precipitation slightly. In the FLAT and NODST experiments, precipitation extended farther downstream, and covered a broader area, though maximum amounts were reduced. Similarly, Hjelmfelt (1992) found that overall precipitation rates did not increase significantly when topography was included in simulations of lake-effect snow from Lake Michigan, though local precipitation rates were significantly increased in regions of strong orographic ascent. Furthermore, overall precipitation rates increased least when synoptic flow was parallel to orography, as was the case in the present study.

**Effects of Shoreline Roughness Contrasts**

Several studies have found that shoreline roughness contrasts may enhance lake-effect snowfall. Nicosia et al. (1999) documented a lake-enhanced rainband near the Lake
Erie shoreline which may have been enhanced by lake–land frictional contrasts. It was shown theoretically that frictional convergence may occur near a shoreline due to horizontal speed shear in the low level wind flow and due to horizontal directional shear resulting from a smaller angle between isobars and wind direction over water than over land. Hjelmfelt (1990) showed that precipitation rates and vertical velocities increased dramatically in a numerical model when surface roughness lengths were increased over Lake Michigan, enhancing heat, moisture, and momentum fluxes. Lavoie (1972) found that the effect of shoreline frictional contrasts alone produced elevated inversion heights and enhanced upward velocities over the lee shore of Lake Erie in a numerical model.

In the control simulation of the 7 December 1998 GSLE event, the planetary boundary layer was represented with the Blackadar model, which determines surface fluxes of heat, moisture, and momentum based on similarity theory (Blackadar 1976, 1979; Zhang and Anthes 1982). Surface roughness length affects not only the calculation of surface momentum fluxes, but also heat and moisture fluxes as described in Zhang and Anthes (1982). Over land, a surface roughness length of 10 cm was used, while over water roughness length was calculated as a function of friction velocity (Delsol et al. 1971). To study the influence of surface friction on GSLE snowbands, two simulations were run altering frictional characteristics over the GSL. The first experiment used a surface roughness length of 10 cm over both land and water to eliminate contrasts in friction across the GSL shoreline (ROUGH1). This treatment, however, approximately doubled moisture fluxes and quadrupled heat fluxes over the GSL. Resulting precipitation ranged up to 65.1 mm compared with 19.3 mm in CTL from 0000 to 1500 UTC, while the domain average precipitation was 135% more than CTL (Fig 5.8a,b). The wind speeds
Figure 5.8. Two-km resolution MM5 simulations from 0000 to 1500 UTC 7 December 1998. Total accumulated precipitation (every 2 mm) in solid contours and vector averaged 10-m wind (full barb = 5 m s⁻¹). Model terrain shaded according to scale at upper right. (a) CTL, (b) ROUGH1, (c) ROUGH2
over and surrounding the GSL actually increased with the increased friction of a larger roughness length in ROUGH1, likely due to thermally driven convergence due to stronger pressure troughing induced by increased surface heat fluxes over the GSL, as well as latent heat release in the intensified precipitation. Interestingly, the band of precipitation evident along the west shoreline extending toward the Stansbury Mountains in CTL was oriented more zonally in ROUGH1. This appears to be a result of intensified convergence over the GSL (Fig 5.9a). Hjelmfelt (1990) found an order of magnitude intensification of precipitation rate over Lake Michigan, also using a roughness length of 10 cm for land and water.

Since the goal of this experiment is to examine the role of friction in generating the low-level wind convergence zone, a second experiment was run in which surface momentum fluxes were calculated based on a 10 cm roughness length over both land and water, though heat and moisture fluxes were calculated as in the control run (ROUGH2). This eliminated the dramatic impact of increased surface fluxes on the simulation, while isolating the impact of friction on low-level convergence. ROUGH2 featured reduced low level wind speeds compared to CTL over the GSL due to greater surface friction, though only minor changes in direction occurred (Fig. 5.8c, 5.9b). Due to reduced low level wind speeds over water, both heat and moisture fluxes were decreased approximately 12% over the GSL (not shown). Grid averaged precipitation from 0000 to 1500 UTC was reduced by 11% from CTL, though the maximum near Tooele was reduced to 10.1 mm, barely more than half of CTL (19.3). The precipitation pattern produced by the west shoreline snowband from 0200 to 0700 UTC was changed little from CTL. Since this snowband occurred near the shoreline, it might be expected that this band would be affected the most
Figure 5.9. Two-km resolution MM5 simulations averaged from 0000 to 1500 UTC 7 December 1998. Difference fields of 10-m wind (scale at lower left) and altimeter setting (every .1 hPa, zero contour not shown). Model terrain shaded according to scale at upper right. (a) ROUGH1 minus CTL, (b) ROUGH2 minus CTL.
by changes to surface frictional parameterizations. However, this band did form and produced nearly identical precipitation to CTL during this time period with little change in wind directions over and surrounding the lake. These results suggest that contrasts in surface friction across the shorelines did not play a primary role in forming GSLE snowbands, though the parameterization of surface properties can have significant consequences on numerical forecasts such as this.

**Effects of Lake Temperature**

As shown earlier in this chapter, sensible heat fluxes from the relatively warm waters of the GSL appear to be of primary importance in driving convergent low-level wind circulations over the lake. Thus lake temperature, which partially determines lake–land and lake–700 hPa temperature difference, should be an important variable for GSLE snow events. Carpenter (1993) and Steenburgh (2000) reached similar conclusions while studying the climatology of GSLE snow. Hjelmfelt (1990) found lake–land temperature difference to be of primary importance in producing land-breeze induced vertical velocities and lake-effect precipitation. Since GSL temperature observations are not currently sufficient to determine the distribution GSL surface temperatures, some uncertainty of GSL surface temperature still exists. Fortunately, Lavoie (1972) and Hjelmfelt and Braham (1983) found that numerical simulations using detailed analyses of lake temperature did not differ significantly from simulations using a mean lake temperature.

To examine the effects of moderate changes in GSL temperature, comparable to the magnitude of diurnal lake temperature oscillations and uncertainties in mean lake temperature due to spatial variations in lake surface temperature, simulations were run
with GSL temperature set 2 °C higher (280K; T+2) and 2 °C lower (276K; T-2) than in
CTL. The results of this experiment are summarized in Fig 5.10. T+2 (T-2) produced a
maximum of roughly 34.7 mm (12.0 mm) of precipitation near Tooele and 32% more
(24% less) precipitation averaged over the grid. The precipitation pattern was similar in
each case as was the evolution of the model simulation (not shown). Fig 5.11 reveals that
T+2 had a more convergent low-level wind flow toward a region of lower pressure
centered over southern Gilbert Bay, likely driven by the additional sensible heat fluxes
generated by the warmer lake in T+2. The T-2 simulation had a nearly equal but opposite
difference from CTL. These results confirm that GSLE is sensitive to lake surface
temperature and that an accurate representation of GSL temperature is critical in accurate
simulation of GSLE precipitation. However, since only the magnitude of precipitation
accumulations and land breeze strength was altered significantly by this experiment, small
errors in GSL temperatures, which are inevitable with the current observation network, are
unlikely to cause a complete misrepresentation of the dynamics of GSLE snowbands for
research applications.

Effects of Upstream Moisture

Given the small size of the GSL and the reduction in saturation vapor pressure due
to salinity, it is possible that upstream moisture is an important variable in GSLE events.
Steenburgh et al. (2000) found that most GSLE events were associated with high values of
700-hPa relative humidity downstream of the GSL at SLC (see Fig. 2.1 for location),
though no representative soundings were available upstream of the GSL. Hjelmfelt (1990)
found that vertical velocities, precipitation, and land breeze strength increase with higher
Figure 5.10. Two-km resolution MM5 simulations from 0000 to 1500 UTC 7 December 1998. Total accumulated precipitation (every 2 mm) in solid contours and vector averaged 10-m wind (full barb = 5 m s$^{-1}$). Model terrain shaded according to scale at upper right. (a) CTL, (b) T+2, (c) T–2.
Figure 5.11. Two-km resolution MM5 simulations averaged from 0000 to 1500 UTC 7 December 1998. Difference fields of 10-m wind (scale at lower left) and altimeter setting (every .1 hPa, zero contour not shown). Model terrain shaded according to scale at upper right. (a) T+2 minus CTL and (b) T-2 Minus CTL.
upstream moisture in simulations of lake-effect snowfall over Lake Michigan. This effect was more pronounced with a higher lake–land temperature difference and lower stability.

In order to determine the effects of upstream moisture for this GSLE event, sensitivity experiments were run with varying amounts of moisture. The first increased relative humidity by 10% (up to a maximum of 100%) in the initial and boundary conditions for the 2-km resolution domain 4 simulation (RH+10). The second decreased relative humidity by 10% down to a minimum of 5%\(^2\) (RH-10). Fig. 5.12 shows the results of these two experiments compared to CTL. RH+10 developed precipitation bands farther upstream than CTL, though the evolution of the snowband and low-level wind circulations were otherwise similar to CTL. As discussed in Hjelmfelt (1990), lake-effect cases with higher ambient moisture require less moisture from surface fluxes to moisten the boundary layer, thus more moisture is available for precipitation. Kristovich and Laird (1996) found that lake-effect clouds on Lake Michigan formed farther upwind in cases where the boundary layer had previously been modified by Lake Superior, though the influence of moisture was not specifically studied. In this case, it appears that this mechanism is responsible for forming precipitation farther upstream than in CTL. RH+10 also produced approximately 27.0 mm of precipitation in the maximum near Tooele, and 91% more domain averaged precipitation. The greater average precipitation can be attributed in part to a greater areal extent of precipitation. The additional latent heat release in precipitation appears to have caused additional convergence in the time-averaged RH+10 simulation compared to CTL (Fig 5.13a).

2. The minimum threshold of 5% was used in order to prevent numerical problems in the model’s explicit moisture scheme.
Figure 5.12. Two-km resolution MM5 simulations from 0000 to 1500 UTC 7 December 1998. Total accumulated precipitation (every 2 mm) in solid contours and vector averaged 10-m wind (full barb = 5 m s\(^{-1}\)). Model terrain shaded according to scale at upper right. (a) CTL, (b) RH+10 simulation, (c) RH-10 simulation.
Figure 5.13. Two-km resolution MM5 simulations averaged from 0000 to 1500 UTC 7 December 1998. Difference fields of 10-m wind (scale at lower left) and altimeter setting (every .1 hPa, zero contour not shown). Model terrain shaded according to scale at upper right. (a) RH+10 minus CTL and (b) RH-10 minus CTL.
RH-10, unlike RH+10, formed the precipitation band farther downstream than in CTL and often did not form a precipitation band, though RH-10 featured a convergence zone which evolved and moved similar to CTL. RH-10 produced only 13.8 mm of precipitation in the maximum near Tooele, and 54% less domain averaged precipitation. Fig 5.13b shows that RH-10 was slightly less convergent than CTL, likely due to reduced latent heat release feedbacks.

These experiments demonstrate that upstream moisture is of primary importance in determining the intensity and areal coverage of precipitation, as a reduction of only 10% relative humidity was able to change a heavy snowfall event into a moderate one of limited areal extent. Conversely, an increase of 10% produced precipitation capable of covering most of Tooele valley with 10 to 50 cm of snowfall accumulations. Since present day numerical-model root-mean-squared errors in 12-h (36-h) forecasts of 700-hPa relative humidity average 18-25% (22-28%) (White 1997, Cook 1998, White et al. 1999), upstream moisture is clearly a potential problem for forecasting GSLE snow.
CHAPTER 6

SUMMARY AND CONCLUSIONS

This dissertation has examined a lake-effect snow event associated with the Great Salt Lake (GSL) that occurred on 6-7 December 1998. Using surface and upper-air observations, regional-scale analyses from the National Centers for Environmental Prediction (NCEP) Rapid Update Cycle (RUC), radar observations from the KMTX WSR-88D radar, upper-air soundings from the Salt Lake City International Airport, and surface observations from the Utah Mesonet, the synoptic and mesoscale features of this event were described. It was shown that prior to the onset of lake-effect snow, an upper-level trough axis passed from west to east across the GSL, causing winds below 500 hPa to veer from southwesterly to northwesterly. Lapse rates also destabilized and higher relative humidity air moved into northern Utah. Environmental conditions during the event were characterized by a lake-700-hPa temperature difference of up to 22.5 °C, a lake-land temperature difference as large as 10 °C, and a lake-modified surface parcel CAPE as high as 611 J kg⁻¹, values that are favorable for the development of lake-effect precipitation (Steenburgh et al. 2000).

Lake-effect precipitation began ~2200 UTC 6 December when a series of disorganized cells formed over the lake and moved downstream to the south and east. At 0400 UTC 7 December, an organized snowband began to form near the west shoreline of the GSL. This band was aligned parallel to the steering-layer wind (~800 m above lake-
level) and was associated with an abrupt wind shift and line of confluence at the surface. This kinematic structure was analogous to that found during similar events over the Great Lakes (e.g., Peace and Sykes 1966, Passarelli and Braham 1981, Braham 1983). As the event progressed, a second region of precipitation formed over the southern GSL and eastern Tooele Valley, and by 0815 UTC merged with the original snowband to form a mature, solitary, mid-lake snowband. The mature snowband appeared to be aligned along the surface confluence zone, which was now located near the mid-lake axis. In general, the snowband was also oriented parallel to the steering-layer flow, although the snowband orientation briefly became more meridional during a period when the formation point moved rapidly eastward.

By 1445 UTC, the snowband had deteriorated into an area of precipitation with embedded convective cores, drifting northeastward over the GSL. Although surface confluence was still evident over the GSL, steering-layer winds were veering to westerly and temperatures were increasing aloft as an upper-level ridge developed over the region. Significant lowering of the equilibrium level for convection occurred during this period as the base of a strong inversion that was located near 500 hPa at 1200 UTC 6 December, lowered to 700 hPa by 0000 UTC 7 December. As a result, environmental conditions were becoming less favorable for GSLE snowfall due to the shorter over-water fetch and reduced depth of convection (Carpenter 1993, Steenburgh et al. 2000) and by 1900 UTC precipitation cells were no longer forming over the GSL.

The heaviest storm-total snowfall was found in a 10 km wide band that extended from the south shore of the GSL to the city of Tooele. The maximum observed storm-total
snowfall of 36 cm (18.8 mm liquid equivalent) occurred in the city of Tooele. Only trace amounts of snow were reported 30 km from the accumulation band.

A nonhydrostatic numerical model simulation with a maximum horizontal resolution of 2 km was used to further examine this lake-effect event. The simulation, which employed analysis nudging on the 54-km domain for the entire simulation and on the 18-km domain for the first 12 h, closely matched the large-scale evolution of the event, with only small timing or placement errors of synoptic systems. The model run also captured the general characteristics of the observed soundings at SLC. The 2-km resolution domain produced snowbands that were similar in structure to radar reflectivity patterns observed by the KMTX WSR-88D, although errors in timing of up to 5 h were observed and the simulated snowbands appeared to be located farther downstream than the observed reflectivity bands. Surface winds and temperatures were also well simulated compared to observations from the Utah Mesonet, with the exception of a stronger than observed nocturnal inversion over the Great Salt Lake Desert. The development of the strong nocturnal inversion in the simulation may have resulted from errors in the specification of surface properties over the Great Salt Lake Desert. The simulated storm-total precipitation agreed well with radar reflectivity composites and snowfall observations. The maximum simulated precipitation was 19.3 mm, slightly greater than the observed 18.8 mm, and was found in approximately the same location as observed.

Analysis of the formative stage of the event (0300 UTC 7 December 1998) showed that the snowband developed along a land-breeze front near the west shoreline where offshore flow was convergent with northerly flow over the GSL. The development of the land-breeze front and associated convergence zone was associated with localized heating
over the lake surface, which produced pressure troughing near the west and east shorelines. The former was associated with the west shoreline convergence zone and snowband while the latter did not produce a precipitation feature. The complex dual trough structure was produced by the geometry of the upwind (northern) shoreline, which determined the amount of over-water fetch and boundary-layer modification. Specifically, isotherms paralleled the upwind shoreline, with low-level warm anomalies and pressure troughing located downstream of the major bays separated by an intermediate tongue of colder air that was located downstream of Promontory Point, a peninsula that extends southward into the GSL. Similar observations were made along the upwind shore of Lake Ontario by Phillips (1972). Trajectories ending at 0300 UTC showed that the warmest low-level temperatures over the GSL were associated with longest over-water fetches.

Boundary-layer modification by the GSL was illustrated by a series of soundings taken along a selected trajectory as it moved over the GSL. Near-surface temperature and boundary-layer depth increased due to sensible heat fluxes. When the trajectory was located over the hyper-saline north arm near-surface dewpoints actually dropped due to reduced latent heat fluxes and the entrainment of drier air from aloft into the boundary layer. Over the less saline south arm, where latent heat fluxes reached as high as 170 W m\(^{-2}\), the boundary-layer dewpoint again decreased 0.2 °C prior to being ingested into the snowband. This decrease, though small, appeared to be the result of upward moisture transport and mixing through a relatively deep layer of a well-mixed boundary layer.

By 1300 UTC, the snowband was aligned along the mid-lake axis, coincident with a low-level convergence zone and a narrow tongue of locally high near-surface temperatures. Nocturnal cooling over land during the previous 10 h had intensified the
lake–land temperature gradient, near-shoreline pressure gradients, and offshore flow from both shorelines. Because of the stronger offshore flow, trajectories ending at 1300 UTC experienced a shorter over-water fetch compared with those ending at 0300 UTC. Nevertheless, soundings along an over-lake trajectory showed that sensible heating rapidly eroded away a shallow surface inversion and all stable layers up to 550 hPa, increased the near-surface temperature 5.1 °C, and increased the near-surface dewpoint 2.0 °C, resulting in significant boundary-layer destabilization.

A series of sensitivity studies were conducted to examine the influence of surface properties, local terrain features, and upstream moisture on the structure and intensity of this event. The influences of surface heat and moisture fluxes from the GSL were examined with a series of simulations in which the saturation vapor pressure over the GSL was set to that of fresh water (FRESH), latent-heat fluxes over the lake were neglected (NOLHFLX), and sensible-heat fluxes were neglected (NOSHFLX). Compared to the control simulations, FRESH produced 20% more domain-averaged precipitation than the control simulation (CTL), illustrating that reduction in saturation vapor pressure by the hyper-saline composition significantly reduced the magnitude of this event. FRESH also featured stronger convergence over the GSL, presumably a secondary effect related to additional latent heat release due to condensation within the snowband (e.g., Ballentine 1982, Hjelmfelt and Braham 1983, Hjelmfelt 1990). In the absence of moisture fluxes from the GSL (NOLHFX), lake-effect precipitation was still generated, but domain-averaged precipitation was 49% lower than in CTL. Thus, moisture fluxes from the GSL were not necessary for snowband generation, but did significantly enhance snowfall. Compared to CTL, over-lake convergence was weaker in NOLHFX. As with the contrast
in convergence between FRESH and CTL, this appeared to be a secondary effect related to differences in condensational heating in the snowband.

NOSHFLX was used to examine the impact of localized sensible heating over the GSL on thermally-induced convergence and snowband initiation and evolution. Compared to CTL, domain-averaged precipitation in this simulation was 28% lower. In addition, banded precipitation structures that developed over the GSL occurred relatively infrequently. Instead, the bulk of the precipitation in NOSHFLX fell over orographically-favored regions. This simulation thus illustrates the importance of thermally-driven circulations induced by the GSL in organizing low-level convergence and the initiation of convection in the post-frontal environment. In the absence of these thermally-driven circulations, the primary mechanism for precipitation generation is orographic uplift.

The effects of latent heat release due to condensation was examined using a simulation in which the latent heat release was ignored (NOLHR). Compared with CTL, the domain-averaged precipitation was reduced just 18%, although the precipitation maximum was reduced by 45%. Thus, the intensity of the band was diminished in the absence of latent heat release. Surface convergence was also weaker in NOLHR, and the updraft velocity and height were lower, similar to modeling studies of lake-effect snowbands over Lake Michigan (e.g., Ballentine 1982, Hjelmfelt and Braham 1983, Hjelmfelt 1990).

Several studies have examined the importance of frictional convergence on lake-effect precipitation over the Great Lakes (e.g., Lavoie 1972, Hjelmfelt 1990, Nicosia et al. 1999). To examine the role of friction in this event, a simulation was performed with the surface roughness length for all flux calculations over the GSL set to that specified over
land (ROUGH1). Compared to CTL, ROUGH1 featured enhanced precipitation rates, primarily due to increased surface fluxes of heat and moisture. Wind speeds, mesoscale pressure troughing, and over-lake convergence were also much stronger due to the impact of the heat and moisture fluxes. It was concluded that the use of land values of roughness length increased surface moisture and heat fluxes to values that were unrealistic. Since the primary goal of this component of the study was to determine the role of friction in generating the convergence zone, a second simulation was run where only the momentum flux calculations over the lake used the same roughness length as specified over land (ROUGH2). This simulation produced a zone of low-level convergence and precipitation pattern that were similar to CTL. Over the GSL, wind speeds and convergence were weaker than in CTL, though low-level convergence and snowbands similar to CTL still occurred, illustrating that contrasts in surface friction were not a primary mechanism for snowband generation in this event.

Topographic features surrounding the GSL were altered to study the influence of terrain in northern Utah on this event. Simulations with no topography (FLAT) and no down-stream topography (NODST) showed that lake-effect snowbands occurred in the absence of orographic uplift and channeling. Both sensitivity experiments, however, featured broader precipitation areas with weaker maxima. The primary impact of the topography was to enhance precipitation rates in the snowbands in regions of orographic uplift, and to reduce precipitation rates in regions of orographic descent. These results suggest that thermal forcing due to the warm lake surface was the primary cause of these snowbands, but that orographic uplift, descent, and channelling altered the distribution of snowfall.
The operational utility of mesoscale model simulations of these events is influenced by a number of factors including inaccurate model numerics and physical parameterizations, boundary condition uncertainty from the lateral boundaries and the specification of surface properties, and initial condition uncertainty. Because of limited observations over the GSL, and the rapid change in lake-surface temperature that frequently accompanies the movement of cold air over northern Utah, the specification of lake temperature is a significant problem for operational numerical weather prediction. To illustrate the sensitivity of mesoscale model forecasts to errors in lake temperature, simulations were performed with the lake temperature increased and decreased by 2 °C (T+2 and T-2, respectively). This value was chosen to approximate a typical-to-large uncertainty in mean lake temperature due to diurnal temperature variations, rapid changes due to cold air outbreaks, and unobserved spatial variations in lake temperature. T+2 (T-2) produced 32% more (24% less) domain-averaged precipitation, but otherwise the distribution of precipitation in the simulations were similar to CTL. This indicates that errors in lake temperature on the order of 2 °C can produce significant differences in the quantitative precipitation forecast amount, but not necessarily in the general character and structure of the precipitation patterns, timing, and location.

A potentially major source of uncertainty in high resolution simulations is the relatively low skill of moisture forecasts over the western United States. Present day models feature average 12 h relative humidity root-mean-squared errors over SLC of 18–25% (White et al. 1999), and events over both the GSL and Great Lakes are known to be influenced by the amount of upstream moisture (e.g., Hjelmfelt 1990, Steenburgh et al. 2000). Sensitivity to upstream moisture was studied using simulations in which relative
humidity was changed in the initial and boundary conditions for the 2-km resolution domain. The moist simulation, in which relative humidity was increased by 10\% (RH+10), produced nearly twice as much domain-averaged precipitation as CTL. The snowband in this case initiated farther upstream and much of the additional precipitation fell over the GSL. In contrast, the dry simulation (RH-10) produced less than half as much domain-averaged precipitation as CTL and at times did not develop the snowband at all. These results illustrate the significant role of upstream moisture in determining the intensity and areal coverage of precipitation.

The results described above show that the structure of snowbands in this event was similar to midlake and shoreline-parallel snowbands found on the Great Lakes (e.g. Peace and Sykes 1966, Passarelli and Braham 1981, Hjelmfelt and Braham 1983, Hjelmfelt 1990, Niziol et al. 1995). Localized heating over the relatively warm GSL induced mesoscale pressure troughing, land-breeze circulations, and low-level convergence, resulting in the development of convective updrafts and a wind-parallel band of clouds and precipitation. The hyper-saline content of the GSL was found to result in greatly reduced moisture fluxes over the lake’s north arm, while moisture fluxes over the south arm were reduced less compared to fresh water, and sensitivity studies illustrated a significant reduction in storm-total precipitation due to these effects. Nevertheless, through direct and secondary effects, moisture fluxes over the lake increased precipitation amounts by about 94\%. Orographic uplift and channelling were not found to be essential for producing the snowbands. In fact, only modest decreases in domain-averaged precipitation occurred in simulations without topography. Topographic effects were, however, shown to affect the
distribution of precipitation with localized enhancement (reduction) of precipitation in regions of orographic uplift (subsidence).

Although the simulation presented in this paper exhibited significant skill, several factors would limit the skill of the modeling system when applied in an operational environment. The present study used observed analyses for boundary conditions and employed data assimilation on the coarse resolution grids to limit large-scale error growth. Real-time prediction would not have such advantages and errors in the upstream moisture forecast, as well as other characteristics of the synoptic environment, may limit model skill. For example, present day numerical model root-mean-squared errors in 12-36 h forecasts of 700-hPa relative humidity average over 18% (White 1997, Cook 1998, White et al. 1999). Such errors would likely reduce forecast skill given the sensitivity found in this study to upstream moisture. Lake temperatures, salinity, and wave height are also not well known and are not necessarily predicted in present-day mesoscale models, although Powers et al. (1997) developed a coupled wave, lake, and atmospheric model for Lake Erie. Since GSL temperatures and wave activity can vary rapidly, such a modeling approach may be needed in some events. In addition, given the inherent errors in large-scale forecasts, probabilistic guidance from mesoscale ensembles, run at a sufficient resolution to resolve lake-effect precipitation processes, may be required before significant gains in forecast skill are achieved.
REFERENCES


