The Lake Effect of the Great Salt Lake: Overview and Forecast Problems

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ABSTRACT

A lake-effect snow phenomenon along the shore of the Great Salt Lake (GSL) in Utah is documented and related to a similar, well-documented lake effect along the shores of the Great Lakes. Twenty-eight cases of GSL lake-effect snowfall are examined for common parameters that can be used to forecast the occurrence of the lake effect and the location of the heaviest snowfall. Each of the cases produced at least 4 in. (10 cm) of snow at some point in the Salt Lake Valley, and nine of them produced over a foot (30 cm) of snow. Upper-air data at 700 mb provide information useful in forecasting both the occurrence of the lake effect and the location of the heaviest snowfall. A temperature difference of at least 17°C between the GSL and 700 mb is common in the heaviest snowfall cases. A method for real-time diagnosis of the temperature of the GSL is discussed. The 700-mb wind direction is useful for predicting the location of heaviest snowfall. The 18 October 1984 case is highlighted as an example of a GSL lake-effect storm and as an example of the forecast problems presented by the GSL lake effect.

1. Introduction

Lake-effect snowstorms near the Great Lakes of North America have been the subject of research and forecast studies for several decades (Wiggins 1950; Peace and Sykes 1966). Meteorologists have noted conditions that are favorable for the development of the lake effect (Wiggin 1950), have developed forecast decision trees to determine the potential for its occurrence in certain locations (Niziol 1987), and have modeled the processes that produce it (Lavoie 1972).

Along the shores of the Great Salt Lake (GSL) in Utah, a similar phenomenon is observed. The GSL lake effect typically occurs in cold northwest flow, following the passage of an upper-level trough. In most areas, the passage of the upper-level trough signals improving conditions and decreasing precipitation. However, downwind of the GSL shoreline, persistent shallow snow bands can extend the period of precipitation significantly and generate large snowfall accumulations. The combination of cold air following trough passage and a boundary layer warmed by a relatively warm lake surface presumably generate enough instability to produce the persistent snow bands. The warm lake surface may also act as a moisture source for the precipitation.

In general, the GSL lake effect affects only a very localized area near the shores of the GSL. However, a major portion of the Utah population lives within the Salt Lake Valley, which is just downwind of the GSL in these northwesterly flow situations. Thus, the public interest in this local phenomenon is quite high. The percentage of annual snowfall in the Salt Lake Valley that can be attributed to the lake effect is not known. Informal studies at the Salt Lake City Weather Service Forecast Office (WSFO) estimate that the average number of lake-effect cases per year is between six and eight. Other studies indicate that the lake effect probably occurs only a small percentage of the time (Dunn 1983). The current study shows that rather restrictive temperature and wind flow regimes are necessary to produce a GSL lake effect, and thus also indicates that it is relatively rare. Nevertheless, the ability of the lake effect to produce large snowfall amounts in localized areas that also coincide with the major population centers of northwest Utah makes it an interesting and challenging forecast problem for forecasters at the WSFO in Salt Lake City.

The purpose of this paper is to give an overview of the forecast problems posed by the GSL lake effect, to show some of the features common to significant GSL lake-effect snowstorms, and to show a case study example of a GSL lake-effect event that had a disastrous impact on the Salt Lake City metropolitan area.

2. GSL lake-effect forecast problems

A major portion of the population of Utah lives within the Salt Lake Valley between the Wasatch
Mountains and the Great Salt Lake (Fig. 1). The location of the lake, northwest of the Salt Lake City area, means that trajectories into the metropolitan area are over water for large distances when flow is from the northwest. For example, at the WSFO, located at the Salt Lake City International Airport, winds blowing from the 320° radial travel approximately 120 km over water before reaching the station. More northerly flows have the longest over-water trajectory into areas of the Tooele Valley, west of Salt Lake City.

Small-scale surface and topographical effects probably play a significant role in lake-effect storms. For example, frictional convergence probably plays a role in initiating the convective bands that form in the unstable air downwind from the lake. Boundary-layer winds over the relatively smooth water surface that impinge upon the rough urbanized land areas of the nearby valleys probably slow due to the influence of increased surface friction. A simple consideration of the balance of forces (for example, Hess 1959, p. 179) reveals that an area of low-level convergence should be produced along the lee shoreline. This has been identified as a factor in Great Lakes lake-effect storms (Lavoie 1972), and it is logical to assume that it is also important in GSL lake-effect storms.

The complex topography of the region probably also plays a role in the lake-effect storms. The low-level flow is probably modified by the channeling effect of the north-south-oriented mountain ranges in northwest Utah. The mountain barrier likely steers the northwest flow southward in the low levels. The combination of this channeling and the frictional convergence should lead to strong low-level convergence in the southeastern portions of the Tooele Valley and along the eastern side of the Salt Lake Valley. However, the complexity of the nearby topography may create many unusual wind features that are important for fueling the lake-effect snow in addition to simple channeling.

The mesoscale nature of the GSL lake effect can produce large differences in weather conditions across a small area. Conventional large-scale forecasting tools available to forecasters cannot effectively deal with the small-scale phenomena. For example, numerical models that are routinely available to forecasters, such as the Nested Grid Model from the National Meteo-

Fig. 2. Least-squares regression curves for predicting Great Salt Lake temperature from the WSFO Salt Lake City (a) previous 7-day mean air temperature and (b) normal mean air temperature.
Table 1. Dates and various parameters for each case of GSL lake-effect snowfall investigated. Temperatures are in degrees Celsius. The lake temperature listed in column 3 was computed using the normal curve in Fig. 2b. The temperature difference listed in column 5 is between 700 mb and the lake. The greatest snow accumulation (in cm) associated with each storm is listed in column 7.

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Rheological Center (Hoke et al. 1989), do not have sufficient resolution to capture the effect. At the current resolution of the numerical models, only a few grid points are located over the waters of the GSL, and there is no attempt to account for the surface energy fluxes produced by the shallow lake.

Fig. 3. Water equivalent precipitation for (a) normal October–April period and (b) 28 lake-effect cases. Values in inches (2.54 cm), with contour interval of 2 in.
The current observing network is also inadequate to
detect when the small-scale effect is occurring. Often,
the most reliable indicator of a GSL lake-effect storm
is provided by visual observations of shallow convective
clouds forming over the lake. Although these observa-
tions are valuable, they are limited by darkness, ob-
scurations, and the availability and reliability of human
observers. Visual and infrared satellite imagery of lake-
effect storms often lack the resolution needed to follow
changes in the snow bands. Furthermore, at night, the
cloud tops of the lake-effect snow bands are nearly the
same temperature as the surrounding high topography,
and are not easily discernable in infrared imagery.
Weather radar data that is currently available to fore-
casters in Salt Lake City has not been effective at de-
tecting GSL lake-effect snow. It is hoped that the pend-
ing installation of more sensitive and higher-resolution
Doppler radars by the National Weather Service (NWS)
will provide a continuous and high-resolution obser-
vation of the snow bands.

Since the ability to detect and forecast the lake effect
using conventional tools is inadequate, forecasters must
rely heavily on experience and historical analogs.
However, since GSL lake-effect snowstorms may occur
only a few times in a year, or not at all in some years,
it is difficult for forecasters to achieve a sufficient level
of experience with the phenomena. Thus, many cases
of lake-effect snowfall are investigated here to deter-
mine any similarities among the historical events.

The data for this study are taken from lake-effect
cases that occurred in northwest flow following the
passages of upper-level troughs. Twenty-eight cases
from March 1971 to May 1988 are considered in this
study and listed in Table 1, along with values for several
lake and atmospheric variables during each event. All
of the cases produced at least 4 in. (10 cm) of snow at
some point in the valley, and nine of them produced
over a foot (30 cm). Occasionally, it appears that lake-
effect snow may occur in southwest flow that enhances
or produces snowfall close to the northeast shoreline
of the GSL. Data for these cases were not available for
this study.

As discussed above, the observational network is not
adequate to always identify when a lake effect is oc-
curring. The cases considered here were obtained from
an informal log kept at the Salt Lake City WSFO. When

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Fig. 4. Water equivalent precipitation for lake-effect cases with
700-mb wind direction from (a) 340° to 350°, (b) 330° to 340°, (c)
320° to 330°, (d) 310° to 320°, (e) 300° to 310°, (f) 290° to 300°,
(g) 280° to 290°. Values in inches (2.54 cm), with contour interval
of 0.5 in.

Fig. 5. Storm total snowfall for 17–18 October 1984 storm.
Values in inches (2.54 cm). Topography as in Fig. 1.
visual observations and spotter reports indicated a concurrence of convection over the GSL and localized heavy snowfall nearby, the case was logged as a possible lake-effect case, and atmospheric data were saved for further analysis. Thus, this dataset is not intended as a complete study of all lake-effect cases that occurred from 1971 to 1988, but probably considers many of the most significant storms.

3. Meteorological elements of a GSL lake effect

a. Temperature of the lake

The temperature of the GSL is an important factor for lake-effect snowfall for two reasons. As has been shown for Great Lakes lake effect (Ellenton and Danard 1979), the temperature of the GSL is an indicator of the amount of energy available in the lake that may be transferred to the lowest layers of the atmosphere and increase instability. The temperature of the lake also determines the maximum saturation vapor pressure of the air mass over the lake. The vapor pressure over the high-salinity water of the GSL is somewhat lower than that over fresh water (Dickson et al. 1965), but the difference is usually not large enough to prevent a transfer of moisture into the overlying air mass during typical lake-effect conditions.

For both forecasting and research purposes, it would be ideal to have daily GSL temperature measurements at several points around the lake. And for operational forecasting, such information is needed in real time to assess instability potential. However, lake temperature is measured only twice a month by the United States Geological Survey (USGS), at one location near the southern shore of the lake.

To develop a more real-time estimate of the lake temperature, the twice-monthly USGS lake temperatures for the months of October–April from 1980 to 1985 were compared to atmospheric temperatures at the Salt Lake City WSFO. The GSL is a rather shallow lake, and can respond fairly rapidly to periods of abnormally warm or cold atmospheric conditions. Figure 2 shows the least-squares regression curves relating the lake temperature to (a) the average of mean air temperatures over the 7 days prior to the lake temperature measurement and (b) the 30-year normal mean air temperature at the time of the lake temperature measurement. The correlation coefficient for the curve in Fig. 2a is .88, with a standard error of 2.7°C. For Fig. 2b, the correlation coefficient is .91, with a standard error of 2.3°C. As shown in both figures, the temperature of the GSL tends to be a few degrees warmer than the mean temperatures at the WSFO.

There are some slight differences in individual predicted lake temperature values using the two different methods. For example, using a temperature of 5°C for the normal mean produces a lake temperature estimate of 8°C, while a 7-day mean temperature of 5°C results
in a lake temperature estimate of 6°C. Even though the difference is small in this example, a period of abnormally warm or cold temperatures could cause a large difference in the two estimates. This could be a significant difference when evaluating instability criteria.

Considering the size of the standard error for these two curves and considering that real-time data on the temperature of the lake is not available, an average of the two predictions should normally produce an acceptable lake temperature estimate. Averaging the two predictions allows for an abnormally warm or cold period to be taken into account, while still considering the rather consistent annual cycle of lake temperatures. The estimate of GSL temperature shown for each case in Table 1 was calculated from the normal mean temperature (Fig. 2b).

b. Instability criteria

One of the necessary conditions for a significant GSL effect snowfall is an unstable layer of air of sufficient depth to allow upward vertical motion. This instability is achieved when the lake is warm enough and/or the air mass aloft is cold enough to create a lapse rate that is moist adiabatic, or greater. For Great Lakes lake-
effect storms, this lapse rate is evaluated using the 850-mb temperature and the lake temperature (Rothrock 1969). For the higher elevations around the GSL, the most convenient and representative way of measuring this lapse rate is to use the 700-mb temperature along with lake temperature. For the Salt Lake City WSFO—where the typical surface pressure is near 870 mb—and using a lake temperature of 5°C, a moist-adiabatic lapse rate results when the temperature difference between the lake and 700 mb is about 10°C, and a dry-adiabatic lapse rate results with a difference of about 17°C.

Table 1 lists the difference between the 700-mb temperature and the estimated lake temperature for each of the 28 cases used in this study. It appears that, within this sample of cases, the heaviest snowstorms occur when the temperature difference is more than 17°C. It has been theorized that an upper limit for temperature differences in lake-effect storms should be evident, since extremely cold air masses are often accompanied by strong subsidence that would oppose the formation of convection. An upper limit for the temperature difference is not clearly evident in the sample of cases examined here. For example, two cases with large temperature differences that also produced large snow accumulations are October 1984 and October 1975. It may be that during the winter and spring, when the GSL is relatively cold, that large temperature differences between the lake and 700 mb are produced by very cold air masses that are quite stable. In the fall,
when the lake is warm, large temperature differences can be produced by air masses that allow for more instability.

c. Inversion height

A limiting factor to lake-effect snowfall and one that is strongly correlated to the end of the snowfall is the height of the subsidence inversion following the 700-mb trough (Rothrock 1969). The height of this inversion serves to limit the vertical extent of the convective cloud tops. A look at “before and after” rawinsonde traces for the cases in this study indicates that an inversion height between 650 and 700 mb is probably the minimum for production of any precipitation. Once the inversion drops below 700 mb, the precipitation generally ends.

d. Upper-level support

Possibly due to the smaller size of the GSL, the lake-effect convection appears to be weaker than the convection observed in the Great Lakes lake effect. The role of upper-level features in helping to initiate convection, or sustain convection that has already begun, may be important in some storms. The heaviest GSL lake-effect snowfalls occur in strong northwest flow where jet streaks embedded in the flow can play a part in enhancing or diminishing the precipitation. Also, the periods of heaviest snowfall in many of these cases seemed to coincide with the passage of a weak 700-mb temperature trough that was embedded in the northwest flow. These forcings for vertical motion from other

![Fig. 9. Schematic diagram of cold-air damming in Salt Lake Valley.](image)

sources must be taken into account when evaluating the evolution of lake-effect convection.

e. Location of snowfall versus wind direction

As has been discussed by others (Elliott et al. 1985), the areas of heaviest precipitation around the GSL can

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TABLE 2. Surface aviation observations at Salt Lake City WSFO, 18 October 1984.
be estimated by the 700-mb wind direction at the Salt Lake City WSFO during a lake-effect event. The 700-mb wind direction has been used because it is routinely observed, and unlike the surface wind at the Salt Lake City WSFO, it is not as affected by frequent fluctuations due to terrain or local drainage effects. Here, the average 700-mb wind direction during each event has been estimated from the routine Salt Lake City rawinsonde ascents (see Table 1) and used to generate maps of the liquid equivalent precipitation at nearby climatological stations for different flow directions.

The location of climatological precipitation stations used in this analysis are shown in Fig. 1, and maps of the normal October–April water equivalent precipitation and the total water equivalent precipitation for all 28 cases are shown in Fig. 3. As mentioned earlier, terrain effects on precipitation in the complex topography surrounding the lake can be great. Note that in Figs. 3a and 3b there are large variations in both the normal and lake-effect precipitation between valley and mountain stations. It should be noted, however, that in every case examined here there were large spatial variations in precipitation amount that were clearly seen above the typical mountain-valley variations.

Figure 4 shows maps of the total precipitation for all of the events where the 700-mb wind direction fell within 10° bands between 350° and 280°. Many cases are included in two maps because their average 700-mb wind direction fell on the boundary between categories. The total precipitation values have not been divided by the number of cases that fell within each category. This is to avoid the temptation to associate an average snowfall amount with a particular wind direction. Rather, the intent is to show that the spatial pattern of snowfall changes with wind direction.

Figures 4a and 4b show the precipitation data for cases with an average 700-mb wind direction from 330° through 350°. The area receiving the greatest precipitation in these cases is in the Tooele Valley west of Salt Lake City. There are indications from local spotter reports that areas along the western side of the Salt Lake Valley also receive large amounts of snowfall in these “northerly” cases, but only a few climatological precipitation stations are found in this area.

A dramatic switch in the location of the maximum snowfall occurs when the 700-mb wind direction becomes more westerly. This is shown in Fig. 4c where the average 700-mb wind direction is from 320° through 330°. The heavy snowfall maximum in this case is located in the southern portion of the Salt Lake Valley. Average 700-mb wind directions from 290° through 320° produce essentially the same results (Figs. 4d–g). As the wind direction becomes even more westerly the pattern of heaviest precipitation in the Wasatch Mountains appears to move northward. Figure 4g does not show this shift very well, because this group is dominated by the 18 October 1984 storm that produced extremely large precipitation amounts in the

![Figure 10](file)

**Fig. 10.** Divergence of \( \mathbf{Q} \) vectors at (a) 700 and (b) 500 mb for 1200 UTC 18 October 1984. Contour interval is (a) \( 6 \times 10^{-17} \text{ s}^{-3} \text{ mb}^{-1} \) and (b) \( 10 \times 10^{-17} \text{ s}^{-3} \text{ mb}^{-1} \), with negative contours dashed and the zero contour dotted.
mountains southeast of Salt Lake City. A larger dataset of lake-effect cases is needed to average out the unusually heavy snowfall events.

4. An example of the GSL lake effect

On 17–18 October 1984, a record 18.4 in. (47 cm) of snow fell at the Salt Lake City WSFO. All but 1 in. (2.54 cm) of this record amount fell between 0030 UTC and 1830 UTC on 18 October. The heaviest snow fell in a relatively small area in the eastern portion of the Salt Lake Valley, while little or no snow fell on the west side of the valley (Fig. 5). The storm resulted in approximately $1 million damage in the Salt Lake City area, mostly due to downed power lines and broken limbs from trees that had not yet shed their leaves. This storm contained many of the elements of a classic lake-effect snowstorm.

During the event, the temperature of the Great Salt Lake was estimated to be around 10°C, using a combination of the previous 7-day mean air temperature and the normal air temperature discussed earlier and illustrated in Fig. 2. This lake temperature estimate was later confirmed by the USGS, which had measured the lake temperature a few days before the event.

Figure 6 shows the National Meteorological Center

![Figure 11](image)

**Fig. 11.** NMC Limited Fine Mesh (LFM) model analyses of 500-mb height (solid) and vorticity (dashed) analysis for (a) 0000 UTC 18 October 1984, (b) 1200 UTC 18 October 1984, and (c) 0000 UTC 19 October 1984. Height contours at 60-dam intervals, and vorticity contours at $2 \times 10^{-5}$ s$^{-1}$ intervals.
(NMC) 700-mb height and temperature analysis for 1200 UTC 18 October. Note that there was strong cold advection occurring at this time. Temperatures at 700 mb would have maintained a difference greater than 17°C from the lake throughout the time period that the snow fell. The estimated average 700-mb wind direction during the event was 290°.

The precipitable water values (in inches) at 1200 UTC on 18 October are shown in Fig. 7. First, note the difference between Salt Lake City's value and those upstream (.31 vs .19-.25 in.). Second, note that the air mass alone would not be capable of producing the 1.76 in. of water equivalent precipitation that fell at the Salt Lake City WSFO, which indicate the important local effects of the lake.

Figure 8 shows the surface analyses for 0900 UTC and 1800 UTC on 18 October. Surface winds at the airport remained out of the northwest during the storm, then shifted to southwesterly at 1800 UTC. It is a relatively common occurrence for the surface wind at the Salt Lake City WSFO to shift to the southwest and then to the southeast during a lake-effect event, while the wind at nearby surrounding stations remains northwesterly. It is the hypothesis of this author that this is a reflection of a local front or boundary-layer front (Garner 1986). This local front could be considered similar to the New England coastal front (Nielsen 1989; Bosart et al. 1972). In the case of the GSL, the local front could be formed as cold air at the surface becomes blocked by the mountains in the upper portion of the valley, resulting in a small dome of cold air (Fig. 9). Low-level flow coming off the Great Salt Lake (relatively warm air) could ride up and over this cold dome, resulting in localized upward vertical motion. Within the cold dome over the valley, the flow may become detached from the larger-scale flow, resulting in a southerly flow at the WSFO. A table (Table 2) of observations for the 18 October 1984 storm shows this occurring near the end of the event. Without mesoscale observational data to confirm the foregoing scenario, it is mostly conjecture, but not unreasonable.

Interestingly, synoptic-scale forcing during the peak of this storm favored downward motion over northern Utah. As an indicator of quasigeostrophic vertical motions (Barnes 1985), Fig. 10 shows the divergence of the Q vectors at 700 and 500 mb for 1200 UTC 18 October. Software to calculate such fields was not available at the time of this event, but is now available to forecasters at the WSFO (Foster 1988). At 700 mb, there is little divergence or convergence of the Q vectors, while at 500 mb there is divergence of Q vectors, indicating synoptic-scale downward motion. Figure 11 shows the NMC 500-mb height and vorticity analyses during the event. The upper-level trough line is well east of Utah during the heaviest snowfall, and vorticity advection appears very weak. Figure 12 shows the 250-mb isotach analysis and divergence at 1200 UTC 18 October. The jet streak shown in Fig. 12b may have contributed to some upward vertical motion early in the storm (before approximately 0900 UTC), but

Fig. 12. (a) 250-mb divergence and (b) 200-mb isotachs for 1200 UTC 18 October 1984. Divergence contours at $10 \times 10^{-5}$ s$^{-1}$ intervals and isotachs at 10-kt intervals.
much of the snowfall at the WSFO occurred while northwest Utah was under the influence of the convergent left-rear quadrant of the jet streak, an area generally associated with downward vertical motion (Uccellini and Kocin 1987).

Geostationary satellite images in Fig. 13 show the dissipation stage of the cloud cover over the lake during the last few hours of the storm. Early in the day, the lake is nearly covered with clouds. The cloud cover then dissipates from north to south until only a narrow band remains along the southeastern shore in the 1630 UTC image. This gradual shrinking of the cloud layer over the GSL may be a reflection of the surface of the lake being cooled by the cold air mass. Also, an increase in the stability of the air mass and lowering of the subsidence inversion in the strong cold advection are probably occurring during this time.

This lake-effect storm followed 1 day after a more widespread winter storm that produced snow over many sections of northwest Utah. Within the Salt Lake Valley, only about 1 in. (2.54 cm) of snow fell from this earlier storm, and the light flurries falling on the evening of 17 October were characteristic of the final stages of many winter storms that move through northwest Utah. The first problem in forecasting any type of event that occurs infrequently is recognizing the potential for such an event soon enough to make a useful forecast. Since most of the synoptic-scale tools used by forecasters were showing indications for downward motion and the end of precipitation, the
theme of the forecast products issued during the evening and early morning of the event was for the eventual ending of snowfall.

After it became obvious that this was an unusual event, the second problem was forecasting how long the snowfall would continue and what area it would cover. Spotters within the heavily populated Salt Lake Valley were very helpful in determining the areal extent of the snow, but tools for forecasting the ending of the event were not available. By the time the 1200 UTC Salt Lake City sounding was available, 10 in. (25.4 cm) of snow had already fallen, and even though forecasters could examine the sounding and the height of the subsidence inversion, a means of forecasting the rate of descent of the inversion was (and is) not available. Forecasting the end of a lake-effect event remains a difficult forecast problem.

5. Summary

Twenty-eight cases of heavy (greater than 10 cm) GSL lake-effect snowfall were studied to discover parameters that could be used to forecast the occurrence of the lake effect and the location of the heaviest snowfall. It was found that upper-air data taken at the 700-mb level yielded useful information in this regard. A method for predicting the temperature of the GSL was developed. It was found that a difference of at least 17°C between the GSL and 700 mb was common in the heaviest snowstorms. The 700-mb wind direction was also found to be a good predictor of the location of heaviest snowfall.

The conditions that are apparently necessary for the occurrence of GSL lake-effect snowfall in the Salt Lake and Tooele valleys can be summarized as follows. The average 700-mb wind direction should be between 270° and 360°. The difference between the temperature of the Great Salt Lake and the average 700-mb temperature should be at least 17°C. The availability of some kind of upper-level support in the form of quasigeostrophic forcings or a jet streak may aid in the initiation of the lake-effect convection.

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