Lake-Effect Snowstorms in Northern Utah and Western New York with and without Lightning

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ABSTRACT

Lake-effect snowstorms in northern Utah and western New York with and without lightning/thunder are examined. Lake-effect snowstorms with lightning have significantly higher temperatures and dewpoints in the lower troposphere and significantly lower lifted indices than lake-effect snowstorms without lightning. In contrast, there is little difference in dewpoint depressions between events with and without lightning. Surface-to-700-hPa temperature differences (a surrogate for lower-tropospheric lapse rate) for events with and without lightning differ significantly for events in northern Utah, but not for those in western New York. Nearly all events have no convective available potential energy, regardless of the presence of lightning. These results are discussed in the context of current models of storm electrification.

1. Introduction

Observations of lightning and thunder occurring during snowstorms (also known as *thundersnow*) have been reported at least as early as the nineteenth century in Western literature (e.g., Herschel 1888). Thundersnow in China, however, was viewed as a precursor to military advance by the enemy, with reports as early as 1099 A.D. (Wang and Chu 1982). Despite this long observing record, research on thundersnow is nearly nonexistent. MacGorman and Rust (1998, p. 292) state, ''We are aware of no thorough scientific investigation of causal relationships between the electrical state of winter storms and their snowfall. Extensive tests to evaluate the proposed hypotheses concerning possible links between lightning and the mesoscale and synoptic-scale meteorology associated with winter storms have yet to be performed.'' As a result of this lack of scientific information on thundersnow events, the National Oceanic and Atmospheric Administration (NOAA) Storm Prediction Center currently does not issue operational thundersnow forecasts.

The lack of scientific inquiry into thundersnow is probably a result of a number of factors. First, thundersnow in the contiguous United States is relatively uncommon: 1.3% of cool-season (October–May) thunder reports from surface observing stations are associated with snow, and 0.07% of snow reports are associated with thunder (Curran and Pearson 1971). Second, the perceived threat of lightning-related casualties and damage during the winter is less than during the rest of the year. For example, only 0.6% of lightning deaths and 0.7% of lightning injuries occur during December, January, and February (Holle et al. 1997). Third, thunderstorms with snow tend to produce fewer lightning strikes than thunderstorms with rain. For example, in the eastern U.S. Superstorm of March 1993, the majority of observed cloud-to-ground lightning was associated with thunderstorms with rain along the Gulf Coast, not with thundersnow farther north (Orville 1993). Fourth, snow absorbs more sound[1] and more light (e.g., Fraser and Bohren 1992) than rain does, reducing the likelihood of reports of audible thunder or visible lightning. Finally, the colder temperatures may serve a dual purpose by keeping people indoors: injuries due to lightning are minimized and lightning strikes may be less likely to be observed directly. Thus, the impetus to understand thunderstorms with snow is less imperative than that for thunderstorms with rain, despite several documented examples of injurious and damaging wintertime lightning strikes (e.g., Herschel 1888; Holle et at. 1997; Cherington et al. 1998).

Among this dearth of interest in thundersnow, two quantitative studies have been published. Curran and Pearson

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[1] Several authors have found that sounds decrease in intensity and travel less distance over snow-covered ground than bare ground (e.g., Kaye and Evans 1939; Embleton 1996; Albert 1998). Whereas similar damping is to be expected for falling snow, literature quantifying this effect has not been found. Other properties that affect sound propagation in the atmosphere include the lapse rate, humidity content, presence of hydrometeors, wind direction, and wind profile.

TABLE 1. Lake-effect events in northern Utah (a) without lightning and (b) with lightning. Evidence for electrification: TS = surface thundersnow report, CG = cloud-to-ground lightning. Data source for electrification: WWD = World Weather Disc (WeatherDisc Associates, Inc. 1994) thunderstorm beginning and ending times dataset; HC = Holle and Cortinas (1998) surface reports of thunder dataset; BLM = Bureau of Land Management lightning network dataset; NLDN = National Lightning Detection Network dataset. Surface and 700-hPa temperatures are obtained from the sounding data. The surface-to-700-hPa temperature difference is ΔT, a proxy for lower-tropospheric lapse rate. LI = Lifted index. CAPE = convective available potential energy. Source of event in literature: C = Carpenter (1993, Table 1); S = Steenburgh et al. (2000, Table 1); HC = Holle and Cortinas (1998) dataset.

Time of sounding (UTC)	Evidence for electrification	Data source for electrification	Surface temp (°C)	700-hPa temp (°C)	ΔT (°C)	LI (°C)	CAPE (J kg^{-1})	Source of event
(a) Without lightning								
1200 15 Mar 1971	None	WWD	3.0	−10.7	13.7	4.5	0	C
1200 17 Mar 1971	None	WWD	3.4	−5.9	9.3	1.9	0	C
1200 26 Nov 1973	None	WWD	−2.4	−12.7	10.3	3.8	0	C
1200 6 Feb 1974	None	WWD	−7.3	−14.4	7.1	5.8	0	C
1200 10 Apr 1974	None	WWD	1.1	−9.5	10.6	6.4	0	C
1200 23 Oct 1975	None	WWD	1.1	−11.7	12.8	0.7	3	C
1200 12 Mar 1976	None	WWD, HC	−4.4	−14.8	10.4	14.6	0	C
1200 29 Mar 1976	None	WWD, HC	−0.6	−14.1	13.5	1.3	0	C
1200 26 Nov 1976	None	WWD, HC	0.6	−14.8	15.4	5.6	7	C
1200 1 Feb 1977	None	WWD, HC	−0.6	−8.2	7.6	5.9	0	C
1200 2 Mar 1977	None	WWD, HC	−2.8	−13.1	10.3	3.1	0	C
1200 1 Apr 1979	None	HC	−1.7	−13.6	11.9	0.6	5	C
1200 10 Apr 1979	None	HC	2.8	−10.8	13.6	7.4	0	C
1200 25 Nov 1981	None	HC	0.6	−11.6	12.2	6.2	0	C
1200 22 Dec 1981	None	HC	0.0	−12.7	12.7	0.8	0	C
1200 25 Mar 1983	None	HC	1.1	−10.8	11.9	−0.3	0	C
1200 3 Apr 1983	None	HC	2.2	−13.2	15.4	0.1	29	C
1200 8 Nov 1983	None	HC	1.7	−10.2	11.9	4.3	0	C
1200 17 Feb 1984	None	HC	0.6	−10.1	10.7	3.9	0	C
1200 18 Oct 1984	None	HC	2.2	−10.9	13.1	1.6	0	C
1200 26 Nov 1984	None	HC	0.0	−15.0	15.0	1.2	0	C
1200 31 Jan 1985	None	HC, BLM	−20.6	−25.7	5.1	10.8	0	C
1200 19 Nov 1985	None	HC, BLM	−4.4	−18.2	13.8	6.1	0	C
1200 12 Dec 1987	None	HC, BLM	−3.3	−17.6	14.3	6.6	0	C
1200 8 Apr 1988	None	HC, BLM	3.3	−10.4	13.7	13.3	0	C
1200 1 May 1988	None	HC, BLM	3.3	−10.5	13.8	4.8	0	C
0000 27 Nov 1994	None	NLDN	−0.9	−14.5	13.6	−1.6	6	S
1200 8 Dec 1994	None	NLDN	−4.9	−16.3	11.4	2.2	0	S
1200 27 Nov 1995	None	NLDN	0.5	−12.1	12.6	−0.4	60	S
0100 18 Jan 1996	None	NLDN	−3.1	−15.8	12.7	7.0	0	S
0000 26 Jan 1996	None	NLDN	−4.4	−14.8	10.4	−0.5	0	S
1200 20 Oct 1996	None	NLDN	1.6	−11.3	12.9	1.6	0	S
1200 16 Nov 1996	None	NLDN	−0.2	−11.5	11.3	2.6	0	S
1200 14 Dec 1996	None	NLDN	0.0	−15.1	15.1	9.6	0	S
1200 28 Feb 1997	None	NLDN	−0.9	−12.6	11.7	4.3	0	S
1200 1 Apr 1997	None	NLDN	−0.5	−14.0	13.5	2.1	0	S
0000 9 Dec 1997	None	NLDN	0.8	−11.0	11.8	0.9	0	S
1300 25 Feb 1998	None	NLDN	−2.3	−13.1	10.8	5.5	1	S
1200 30 Mar 1998	None	NLDN	0.0	−10.1	10.1	0.3	0	S

TABLE 1. *(Continued)*

Time of sounding (UTC)	Evidence for electrification	Data source for electrification	Surface temp (°C)	700-hPa temp (°C)	ΔT (°C)	LI (°C)	CAPE (J kg^{-1})	Source of event
Avg			−0.9	−12.9	12.0	4.0	2.8	
Std dev			4.1	3.2	2.2	3.7	10.4	
(b) With lightning								
0000 26 Apr 1976	1 TS report	HC	2.8	−8.6	11.4	3.4	0	HC
1200 29 Mar 1977	2 TS reports	WWD	−2.8	−16.5	13.7	0.1	22	C
0000 19 Nov 1979	1 TS report	HC	7.2	−8.5	15.7	1.1	29	HC
0000 1 Apr 1983	1 TS report	HC	7.2	−7.9	15.1	−2.2	0	HC
0000 3 Feb 1986	3 TS reports	HC	10.6	−3.8	14.4	−1.8	0	HC
1200 15 Mar 1986	6 CG strokes	BLM	0.0	−10.3	10.3	1.6	0	C
	1 TS report	HC						
1200 24 Nov 1988	1 TS report	HC	0.0	−11.7	11.7	−1.6	0	HC
1200 4 Oct 1995	12 CG strokes	NLDN	7.3	−6.8	14.1	−2.9	103	S
0000 2 May 1997	1 CG stroke	NLDN	6.0	−9.9	15.9	−0.2	33	S
0000 12 Oct 1997	4 CG strokes	NLDN	6.7	−7.3	14.0	2.6	0	S
Avg			4.5	−9.1	13.6	0.0	18.7	
Std dev			4.0	3.2	1.8	2.0	30.9	

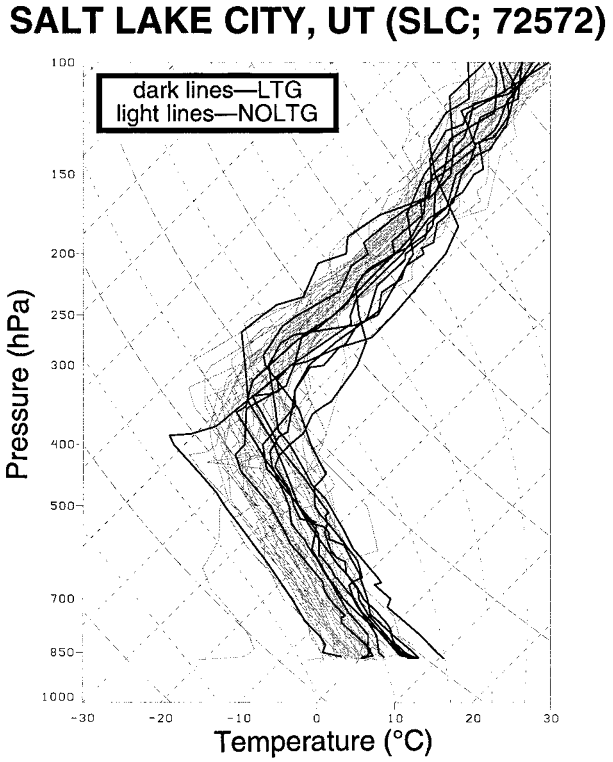

FIG. 1. Skew T–log$p$ plot of temperature profiles (°C) at Salt Lake City for lake-effect snowstorms for LTG (dark solid lines) and for NOLTG (light solid lines) events.

(1971) compiled a climatology of 76 thundersnow reports from 536 surface stations in the United States and Canada for the cool seasons (October–May) 1968–71. Of those reports, proximity soundings could be obtained for 13 events. They found that the lower troposphere was stable and capped by a moist layer of near–convective neutrality (near-constant θ_e), similar to the elevated thunderstorms of Colman (1990a,b, 1991). More recently, Holle and Cortinas (1998) constructed a climatology of 402 thundersnow reports from 476 U.S. and Canadian hourly reporting stations at surface temperatures less than or equal to 10°C for 1976–89. In agreement with Curran and Pearson (1971), these reports occurred primarily over the central Great Plains, with several secondary peaks elsewhere, in particular, areas downwind of large lakes.

This paper focuses upon two of those downwind lake areas known for lake-effect snowstorms: northern Utah (e.g., Carpenter 1993; Steenburgh et al. 2000) and western New York (e.g., Niziol et al. 1995). The desire is to concentrate on lower-tropospheric environments of surface-based convection, as in lake-effect snowstorms, rather than that of elevated convection (Colman 1990a,b, 1991). Preliminary exploration using broader criteria for thundersnow events that included elevated convective events revealed that limiting this study to surface-based lake-effect convection produced more robust results.

Lake-effect snowstorms of the Great Salt Lake and

TABLE 2. Lake-effect events in western New York (a) without lightning and (b) with lightning. The sounding from 22 Nov 1956 was taken from Niagara Falls, NY, rather than Buffalo. Evidence for electrification is as in Table 1, except no BLM evidence and the addition of evidence from SALDN = State University of New York at Albany Lightning Detection Network [plotted data in Moore and Orville (1990)]. Surface and 700-hPa temperatures, ΔT, LI, and CAPE as in Table 1. Source of event in literature: W = data used in Westcott et al. (1999) study [dates obtained from N. Westcott (1999, personal communication)]; MO = Moore and Orville (1990); HC = Holle and Cortinas (1998) dataset.

Time of sounding (UTC)	Evidence for electrification	Data source for electrification	Surface temp (°C)	700-hPa temp (°C)	ΔT (°C)	LI (°C)	CAPE ($J\ kg^{-1}$)	Source of event
(a) Without lightning								
0900 22 Nov 1956	None	WWD	0.7	−18.4	17.7	5.3	0	W
1200 21 Nov 1964	None	WWD	−3.9	−20.5	16.6	12.5	0	W
1200 23 Nov 1970	None	WWD	−3.9	−23.6	19.7	13.8	0	W
1200 29 Nov 1976	None	WWD, HC	−6.1	−12.7	6.6	26.2	0	W
1200 10 Dec 1977	None	WWD, HC	−13.3	−26.0	12.7	17.5	0	W
1200 27 Jan 1978	None	HC	−7.1	−20.6	13.5	23.9	0	W
1200 29 Nov 1979	None	HC	−2.6	−22.0	19.4	11.3	0	W
1200 3 Jan 1981	None	HC	−15.0	−25.1	10.1	26.0	0	W
1200 24 Dec 1983	None	HC	−15.0	−29.9	14.9	14.5	0	W
1200 4 Dec 1984	None	HC	−3.9	−22.0	18.1	16.3	0	MO
1200 19 Jan 1985	None	HC	−6.1	−21.5	15.4	13.7	0	MO
1200 20 Jan 1985	None	HC	−15.6	−32.4	16.8	16.1	0	W
1200 18 Dec 1985	None	HC	−11.1	−31.4	20.3	14.0	0	W
1200 1 Jan 1986	None	HC	−5.0	−18.9	13.9	16.4	0	MO
1200 27 Jan 1986	None	HC	−5.6	−15.8	10.2	13.5	0	W
1200 13 Nov 1986	None	HC	−5.3	−23.8	18.5	24.1	0	MO
1200 10 Dec 1988	None	HC	−7.4	−23.8	16.4	17.1	0	W
1200 20 Dec 1989	None	NLDN	−9.6	−27.1	17.5	15.0	0	W
1200 4 Dec 1991	None	NLDN	−2.9	−20.6	17.7	9.7	0	W
1200 25 Dec 1993	None	NLDN	−7.5	−21.1	13.6	14.5	0	W
Avg			−7.3	−22.9	15.5	16.1	0	
Std dev			4.5	4.8	3.5	5.2	0	
(b) With lightning								
1200 1 Dec 1966	1 TS report	WWD	0.0	−13.7	13.7	17.2	0	W
1200 5 Nov 1967	1 TS report	WWD	−1.1	−20.4	19.3	13.8	0	W
0000 6 Nov 1982	2 TS reports	HC	0.6	−18.5	19.1	6.9	0	HC
0000 30 Nov 1983	4 TS reports	HC	1.1	−17.2	18.3	9.5	0	MO
1200 30 Nov 1983	1 TS report 3 CG strikes	HC SALDN	−1.7	−16.3	14.6	13.1	0	MO
1200 4 Dec 1986	4 CG strikes	SALDN	−0.1	−18.7	18.6	14.1	0	MO
1200 10 Nov 1996	1 CG strike	NLDN	0.0	−16.2	16.2	3.1	0	W
Avg			−0.2	−17.3	17.1	11.1	0	
Std dev			0.9	2.0	2.1	4.5	0	

the eastern Great Lakes typically have many similarities (e.g., Steenburgh et al. 2000): a westerly or northerly component to the lower-tropospheric winds and a lower-tropospheric lapse rate that is nearly dry adiabatic. Other areas where similar surface-based electrified snowstorms occur include the Sea of Japan (e.g., Maekawa et al. 1992; Kitagawa 1992; Michimoto 1993; Kitagawa and Michimoto 1994; also reviewed in MacGorman and Rust 1998, pp. 286–291) and the North Sea (e.g., Halsey and Patton 1999). Lake-effect snowstorms in northern Utah and western New York are classified as those that are associated with lightning/thunder (LTG) and those that are not associated with lightning/thunder (NOLTG).

2. Data

Dates of 49 lake-effect snowstorms of the Great Salt Lake were obtained primarily from the 28-event climatology of Carpenter (1993) and the 16-event climatology of Steenburgh et al. (2000). Five additional events were collected from the thunder climatology of Holle and Cortinas (1998) in the following manner. Initially, all thunder reports at temperatures less than or equal to 10°C for the Utah reporting stations of Hill Air Force Base, Salt Lake City, and Ogden were obtained. Then, reports occurring during an unfavorable synoptic pattern for lake-effect snow [i.e., patterns that were not consistent with those of Steenburgh et al. (2000)] were eliminated. Data from those five events of Holle and Cortinas (1998) so obtained, plus the 44 other events, are listed in Table 1.

Dates of 26 lake-effect snowstorms of western New York were obtained: 6 events (seven different sounding times) from Moore and Orville (1990) and 19 events from Westcott et al. (1999). An additional event was obtained from the thunder climatology of Holle and Cortinas (1998) at Buffalo, New York, in a manner similar to that described above. Data from the 26 events (27 different sounding times) are listed in Table 2.

Four datasets were examined for evidence of light-

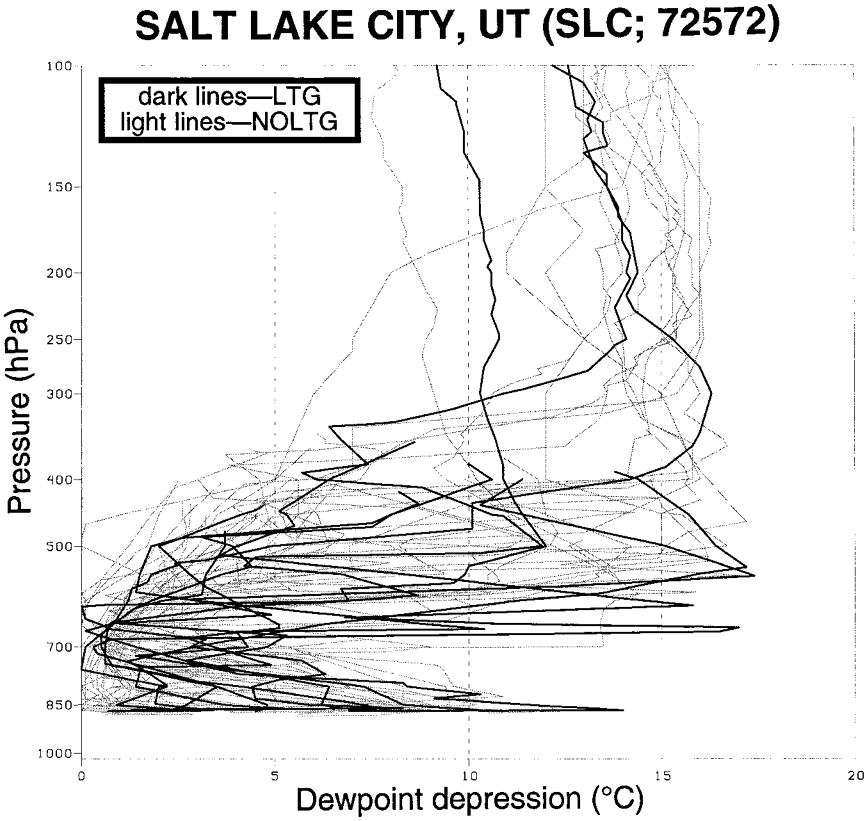

FIG. 2. Vertical profiles of dewpoint depression (°C) at Salt Lake City for lake-effect snowstorms for LTG (dark solid lines) and for NOLTG (light solid lines) events.

ning or thunder to determine whether these lake-effect snowstorms were electrically active. Cloud-to-ground lightning data detected by the Bureau of Land Management lightning network over the western United States for 1985–93 and from the National Lightning Detection Network [NLDN; e.g., Cummins et al. (1998)] over the majority of the contiguous United States for 1989–98 were used. [An overview of lightning detection systems can be found in Holle and López (1993).] Although the networks collect data of generally high quality, detection capability is less than 100%, especially early in the period of record. Thunderstorm beginning and ending times extracted from the original records of surface weather observations for 1948–77 (National Climatic Data Center dataset TD-9945) were obtained on the World WeatherDisc (WeatherDisc Associates, Inc. 1994) for Salt Lake City, Hill Air Force Base, Ogden, and Buffalo. Finally, thunder data from the hourly surface observing network for 1976–89 (Holle and Cortinas 1998) were examined for the same four stations. The Automated Surface Observing System became the official observation at Salt Lake City on 1 March 1998 and at Buffalo on 1 December 1995. Both stations always have been augmented manually for thunder.

Events from Tables 1 and 2 were examined for evidence of electrical activity using these four datasets. Determining conclusively that NOLTG storms did not produce lightning is not possible considering the vagaries of detecting and reporting intracloud and cloud-to-ground lightning either manually or with networks of sensors. In addition, exhaustive climatologies of lake-effect thundersnow events over northern Utah and western New York do not exist. Therefore, the results of this study are tentative, awaiting a more comprehensive database with uniform reporting procedures.

For each of the lake-effect events, one or more soundings were obtained from the North American Radiosonde Database on compact disc (Schwartz and Govett 1992) or from the NOAA/Forecast Systems Laboratory Web site (http://raob.fsl.noaa.gov/Raob_Software.html). If the time of lightning/thunder was known, the sounding or soundings closest in time was used. If the time of lightning/thunder was unknown or if lightning/thunder did not occur, then the sounding closest in time to the start of the lake-effect event was used. If the starting time of the lake-effect event was not known, then the 1200 UTC sounding for the first day of the event was used.[2] These soundings are displayed in sections 3 and 4. All derived values (e.g., lifted index, convective available potential energy) were computed using the Generalized Meteorological Analysis Package (GEMPAK). The Wilcoxon–Mann–Whitney rank-sum test (e.g., Wilks 1995, 138–140) was employed to test for differences between the populations represented by LTG and NOLTG events.

3. Northern Utah

The temperature profiles from Salt Lake City (Fig. 1) indicate considerable range in lower-tropospheric temperatures associated with lake-effect snow (e.g., surface temperatures generally range from −7° to 10°C). The highest temperatures, however, tend to be associated with LTG events. For the Salt Lake City LTG and NOLTG events, the surface temperatures differ at the 1% level according to the Wilcoxon–Mann–Whitney test (i.e., there is a 99% chance that LTG and NOLTG events are derived from separate populations). The 700-hPa temperatures also differ at the 1% level. One event (29 March 1977) is an outlier, being the coldest Salt Lake City sounding associated with thundersnow (Fig. 1 and Table 1b). It is not evident why this event bears a thermodynamic profile drastically different from the others; speculation is offered in section 5.

A corresponding plot of the dewpoint temperature profiles (not shown) is similar to Fig. 1 (i.e., LTG events tend to have higher dewpoint temperatures than NOLTG events in the lower troposphere). The vertical distribution of dewpoint depression, however, appears to show little difference between LTG and NOLTG events (Fig. 2). For both LTG and NOLTG events, a minimum in the dewpoint depression tends to occur at about 700 hPa, suggesting a snow-producing cloud layer of varying thickness about 1500 m above ground level.

Surface-to-700-hPa temperature differences (proxies for lower-tropospheric lapse rates) are larger for the LTG events than for the NOLTG events (Table 1), results statistically significant at the 2% level. Since most cases, LTG and NOLTG, have zero convective available potential energy (CAPE), this parameter is not useful as a lightning forecasting aid. The lifted indices for the LTG events are more unstable than those for the NOLTG events at the 1% level.

This northern Utah dataset can be used to define critical values of the parameters in Table 1. For example, at a surface temperature of 2°C, approximately half of the lake-effect events will be associated with lightning and half will not be associated with lightning. For surface temperatures greater (less) than 2°C, lake-effect snowstorms are more (less) likely to produce lightning. Similar values of 700-hPa temperature, surface-to-700-hPa temperature difference, and lifted index are −10.5°C, 13.7°C, and 1°C, respectively.

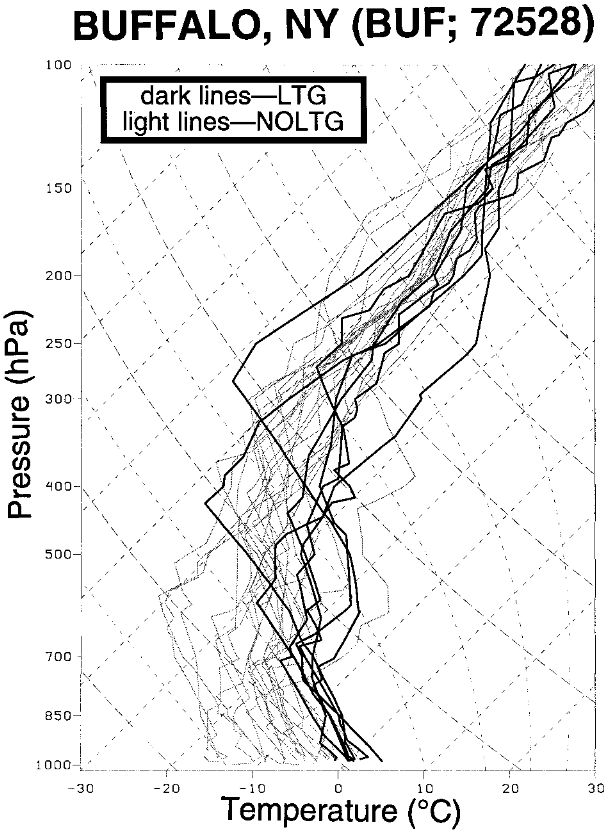

FIG. 3. Skew *T*–log*p* plot of temperature profiles (°C) at Buffalo/Niagara Falls for lake-effect snowstorms for LTG (dark solid lines) and for NOLTG (light solid lines) events.

4. Western New York

The temperature profiles from Buffalo (Fig. 3) are similar to those from Salt Lake City (Fig. 1), with LTG events tending to have warmer tropospheric profiles than NOLTG events. The range of surface temperatures for these events is a little larger for Buffalo than for Salt Lake City (about 20°C compared with 17°C). For the Buffalo LTG and NOLTG events, the surface temperatures differ at the 0.1% level; the 700-hPa temperatures differ at the 1% level.

As at Salt Lake City, profiles of dewpoint depression at Buffalo (Fig. 4) do not show a large difference between LTG and NOLTG events. For both LTG and NOLTG events, a minimum at about 850–800 hPa occurs, consistent with a layer of clouds about 1500 m above ground level.

Surface-to-700-hPa temperature differences are larger for the LTG than for the NOLTG events (Table 2), but

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[2] Because many of the LTG soundings occur at 0000 UTC rather than 1200 UTC, it might be speculated that surface temperatures and lifted indices might be greater for LTG than NOLTG events. Replacing 0000 UTC soundings with the previous 1200 UTC soundings did not appreciably affect these results.

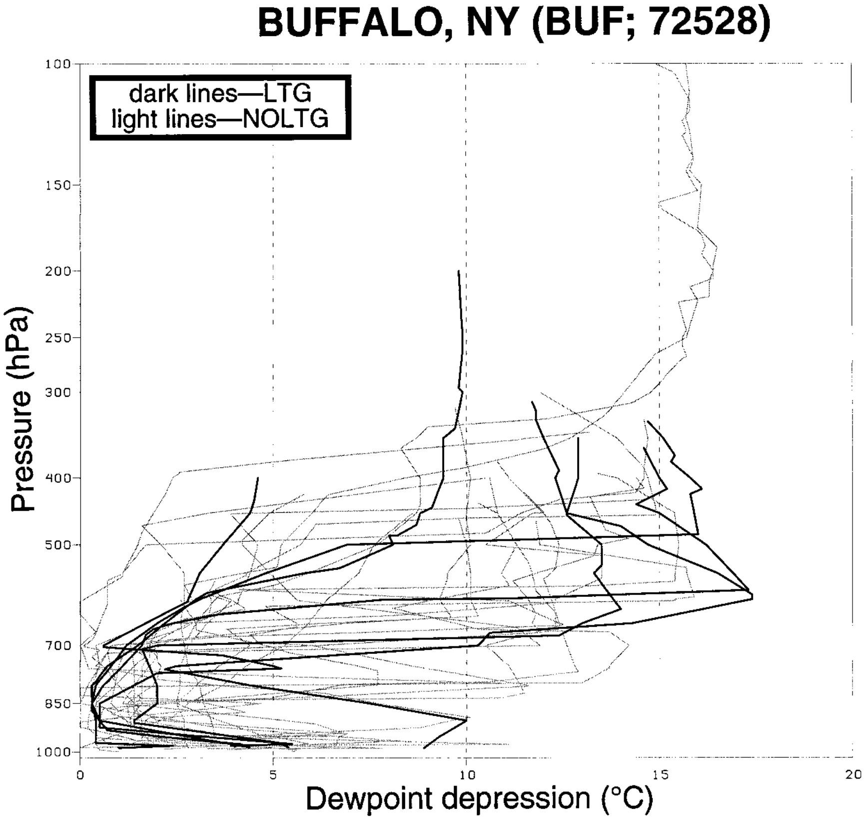

FIG. 4. Vertical profiles of dewpoint depression (°C) at Buffalo/Niagara Falls for lake-effect snowstorms for LTG (dark sold lines) and for NOLTG (light solid lines) events.

the results are not statistically different at the 10% level. Lapse rates over a shallower layer, as measured by the surface-to-850-hPa temperature differences (not shown), also are not statistically different at the 10% level. CAPE for all the Buffalo events (LTG and NOLTG) is zero (Table 2), making this parameter not useful for forecasting lightning in lake-effect snowstorms. The lifted indices for the Buffalo soundings (Table 2) are considerably more stable than those for Salt Lake City (Table 1), an indication of the arctic air masses that affect western New York more frequently than northern Utah. The Buffalo lifted indices for LTG are smaller than those for NOLTG and differ from each other at the 5% level. Critical numbers for lake-effect events associated or not associated with lightning for Buffalo based on this western New York dataset are surface temperature of −2°C, 700-hPa temperature of −20°C, and lifted index of 14°C.

5. Conclusions

The vertical profiles from two different regions affected by lake-effect precipitation (northern Utah and western New York) are examined for differences between lake-effect events that produce lightning and those that do not produce lightning. These statistically significant results indicate that the most useful parameters for forecasting lightning during lake-effect snowstorms are lower-tropospheric temperatures and lifted index, whereas dewpoint depression and CAPE are not useful. Surrogates for the lower-tropospheric lapse rate (e.g., the surface-to-700-hPa temperature difference) are useful for northern Utah events, but not for western New York events.

The present data preclude knowing why the warmer and moister lower troposphere appears to be necessary for lightning. Based on the currently accepted model of storm electrification, charging occurs in the mixed-phase region of the cloud. In this region, generally considered to be −10° to −30°C, strong updrafts (greater than roughly 5 m s^{-1}) supply adequate water vapor for supersaturation and the production of actively riming ice particles, which interact with other unrimed ice particles to produce and separate electrical charge (e.g., Zipser 1994; Solomon and Baker 1994; MacGorman and Rust 1998, pp. 218–220). Given this model, two possibilities could explain the present results. First, at cold temperatures, as in NOLTG events, there may not be enough water vapor in the atmosphere to actively rime ice to promote the transfer of adequate electrical charge in the mixed-phase region, even with sufficient vertical motion. Second, the warmer lower troposphere

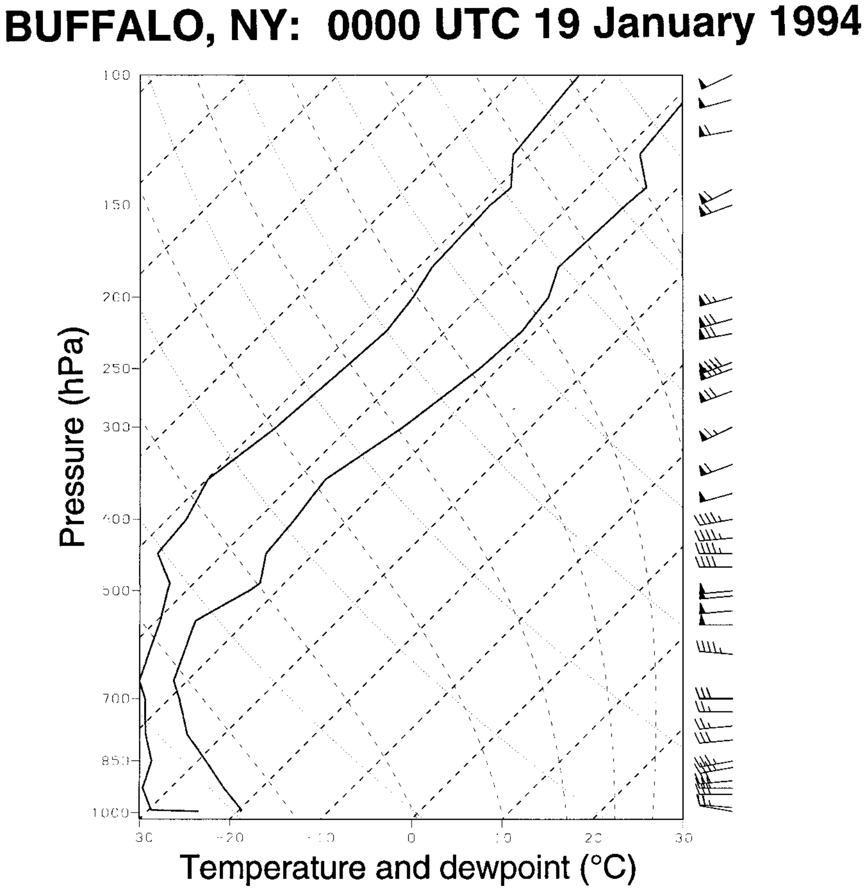

FIG. 5. Skew *T*–log*p* plot of temperature (right) and dewpoint (left) profiles (°C) at Buffalo for a lake-effect snowstorm with lightning on 0000 UTC 19 Jan 1994. Lightning was detected by the National Lightning Detection Network 150 km east of Buffalo at 0304 and 0543 UTC 19 Jan 1994. Horizontal wind profile is displayed along the right side (pennant, full barb, and half barb denote 25, 5, and 2.5 m s^{-1}, respectively).

in the LTG events implies that the height of the -10°C isotherm (the region where charging is believed to occur most efficiently due to the coexistence of high concentrations of ice crystals and supercooled liquid water) is high enough off the ground that vertical motions in the lower troposphere during lake-effect events can become of sufficient intensity to separate charge. This second hypothesis is consistent with results from work on winter thunderstorms over the Sea of Japan (e.g., Kitagawa 1992; Michimoto 1993; Kitagawa and Michimoto 1994).

In addition, this second hypothesis would explain outlier observations of lightning occurring when the lower troposphere is very cold. In northern Utah on 29 March 1977, thunder was reported at a surface temperature of -2.8°C (Fig. 1 and Table 1b). In Oswego, New York, during 19–21 January 1994, numerous intracloud lightning discharges were reported (J. LaDue 1999, personal communication), and two cloud-to-ground discharges were recorded by the NLDN at a surface temperature of -20°C, the warmest temperature in the troposphere for this event (Fig. 5)! In these situations, sensible and latent heating over the lake and the accompanying vertical motions would be extremely large at low levels, perhaps offsetting the effect of much lower temperatures and water vapor mixing ratio on electrification.

Nevertheless, evaluation of these hypotheses awaits field programs designed to address these issues. Specifically, in situ measurements, both downwind of the lake and over the lake, of electrical charge, cloud microphysics, and dynamic/thermodynamic quantities (in particular, vertical velocity) will be needed to understand the charging mechanism in thundersnow and any differences from summertime thunderstorms. If future field experiments were to examine the environment of lake-effect bands with and without lightning and were to find strong updrafts within the mixed-phase region in nonthundering clouds, then cloud microphysics might be implicated as the mechanism limiting lightning production. If, however, high concentrations of liquid water were found in nonthundering clouds, then it might be that strong vertical velocity alone is inadequate to produce lightning. Finally, detailed observations of surface precipitation type should be recorded during lake-effect events, because the occurrence of graupel at the surface has been noted as a precursor to lightning in two cases of wintertime lightning casualties (Cherington et al. 1998) and during fieldwork in Japan (Kitagawa and Michimoto 1994, p. 10 715). Also, Herschel (1888) describes ''snow in hail-like pellets'' falling shortly after a wintertime lightning discharge.

These results can address only the observed vertical structure of the atmosphere for LTG and NOLTG events, not the processes of electrification in such storms. Nevertheless, testable hypotheses can be drawn from these results. Clearly, resolution of such hypotheses will require more detailed climatologies and in situ electrical measurements of lake-effect storms than are currently available.

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