PICTURE OF THE MONTH

A Lake-Effect Snowband over Lake Champlain

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1. Introduction

A number of recent studies have indicated that lake-effect snows can occur in association with lakes of significantly smaller size than the Great Lakes (e.g., Steenburgh et al. 2000; Huggins et al. 2001; Schultz et al. 2004; Sobash et al. 2005). As an example, Lake Champlain lake-effect storms can generate localized snowfalls that are comparable to synoptic winter storms and on rare occasions produce snow squalls with visibilities less than ¼ mile and up to 33-cm (13 in.) of snow in a 12-h period (Tardy 2000). The current article presents observations of a unique lake-effect snowband over Lake Champlain, a north–south-oriented lake situated along the border of northern New York and Vermont. The snowband occurred during a time period of weak southerly surface winds, a distinctive and undocumented condition for a lake-effect event. A description of the evolution of the lake-effect snowband and the environmental conditions that produced the band are provided and compared with those of lake-effect events previously observed over both the Great Lakes and other small lakes.

Lake Champlain is nearly 200 km long and has a maximum width of 19 km, with the complex topography of the Adirondack Mountains to the west and the Green Mountains to the east. The focus of this article is on a lake-effect snow event that began shortly before 0800 UTC 18 January 2003 and dissipated by 1800 UTC, lasting about 10 h. During the event, the snowband extended northward after initially forming over the southern portion of Lake Champlain and was observed by both the National Weather Service (NWS) Weather Surveillance Radar-1988 Doppler (WSR-88D) located near Burlington, Vermont (KCXX), and a high-resolution camera within the CAMNET network (see online at http://www.hazecam.net; Figs. 1 and 2). The camera overlooks downtown Burlington and across Lake Champlain to the Adirondack Mountains in New York State (Fig. 2a). The images, captured every 15 min during daylight hours, show the visible evolution of the snowband over the southern portion of Lake Champlain (Fig. 1) and provide an excellent complement to the WSR-88D radar data (Fig. 2).

2. Synoptic conditions

On the day prior to the event, 17 January 2003, a cold front extending from southwestern Pennsylvania into northern Maine moved through the region (not shown). The front passed through the Lake Champlain valley at approximately 1800 UTC 17 January 2003 and ushered in temperatures well below freezing to a broad area of the Northeast and mid-Atlantic states.

Early on 18 January 2003, a surface trough moved slowly into the Great Lakes region, accompanied by a weak warm front extending to the northeast of Lake Huron. During the same period, a well-defined ridge was positioned northward along the Appalachian Mountains helping cold air to remain in place along the East Coast and in the Lake Champlain valley (Fig. 3). The position of the two pressure systems produced a weak pressure gradient in the vicinity of Lake Cham-

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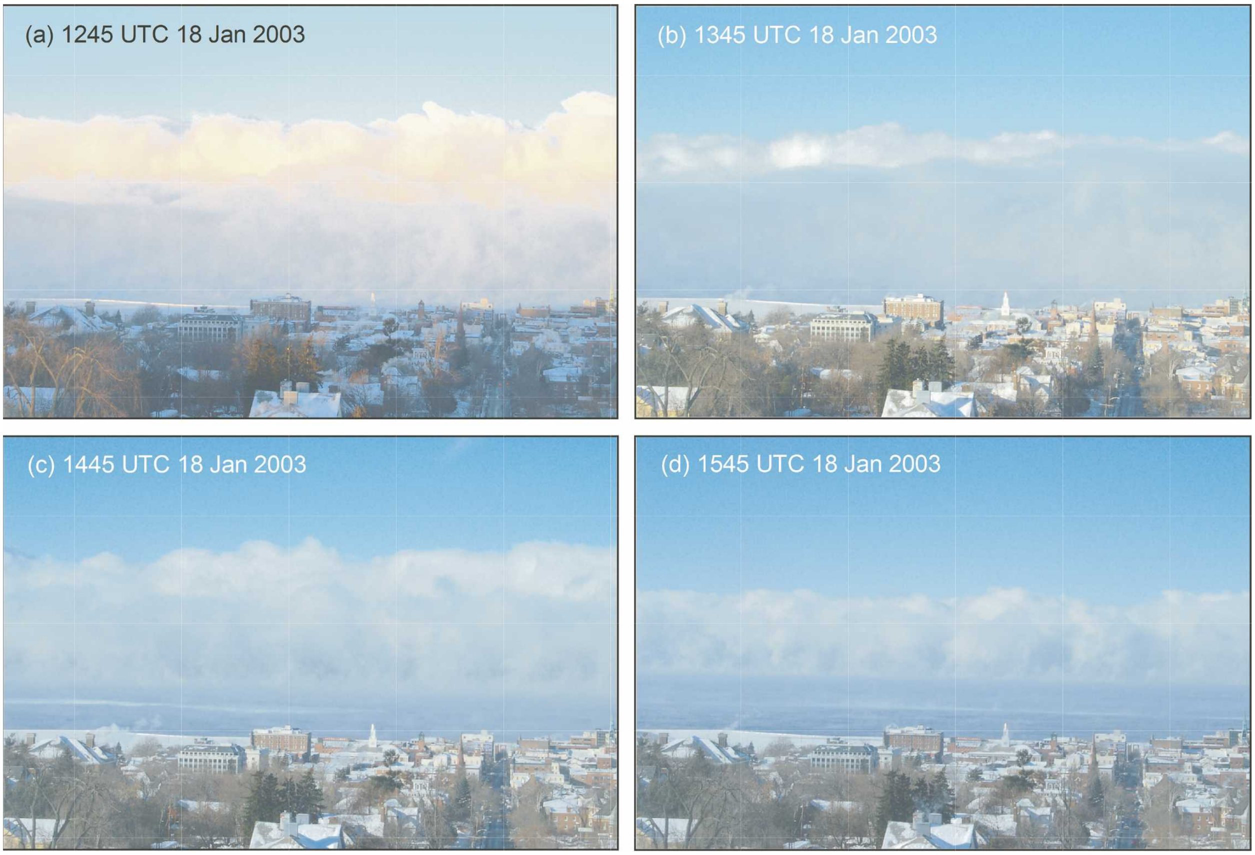

FIG. 1. Photographs of a lake-effect snowband over southern Lake Champlain. High-resolution, real-time photographs were taken by a westward-viewing camera overlooking downtown Burlington and Lake Champlain on 18 Jan 2003. [Photos courtesy of CAMNET webcam network operated by the Northeast States for Coordinated Air Use Management (NESCAUM).]

plain that aided land-breeze development and led to light southerly ambient winds, both important factors for snowband development in this case. The polar air mass extending southward along the Appalachian Mountains was an important factor that allowed for continued advection of cold air into the Lake Champlain valley following a strengthening of the southerly winds. In section 3, we will discuss the mesoscale winds observed in the vicinity of Lake Champlain and their contribution to the initiation and evolution of the lake-effect snowband.

The water temperatures on 18 January at the King Street Ferry Dock in Burlington were about 0.5°C, resulting in a lake-to-air temperature difference of nearly 26°C for locations along the eastern shore (i.e., Burlington) and western shore (i.e., Plattsburgh, New York). This lake-to-air temperature difference that existed during the initial 4 h of the case appreciably exceeded observations presented by Passarelli and Braham (1981) (10°–13°C) and Steenburgh et al. (2000) (6°–14°C), which were shown to have led to the development of land breezes, low-level convergence, and lake-effect snowbands over Lake Michigan and the Great Salt Lake, respectively.

Upper-air soundings from Albany, New York (ALB) and Maniwaki, Quebec, Canada (WMW) at 1200 UTC 18 January (not shown) indicated that the atmosphere was stable throughout the lower troposphere. A surface inversion was in place at WMW and an isothermal lapse rate from the surface to about 900 hPa existed at ALB. Temperatures at 850 hPa were −13.5° and −17.3°C for WMW and ALB, respectively. The lake to 850-hPa temperature difference was nearly 18°C in the vicinity of Lake Champlain, providing an absolutely unstable lapse rate that exceeded the dry-adiabatic lapse rate criteria (≥13°C) suggested as necessary for lake-effect snowbands to develop by numerous observational (e.g., Rothrock 1969; Niziol 1987) and numerical modeling studies (e.g., Hjelmfelt 1990; Laird et al. 2003). Steenburgh et al. (2000) found the dry-adiabatic lapse rate criteria applied for cases over the Great Salt Lake; however, the metric used was in exceedance of 16°C for the lake to 700-hPa temperature difference because of the higher altitude of the lake.

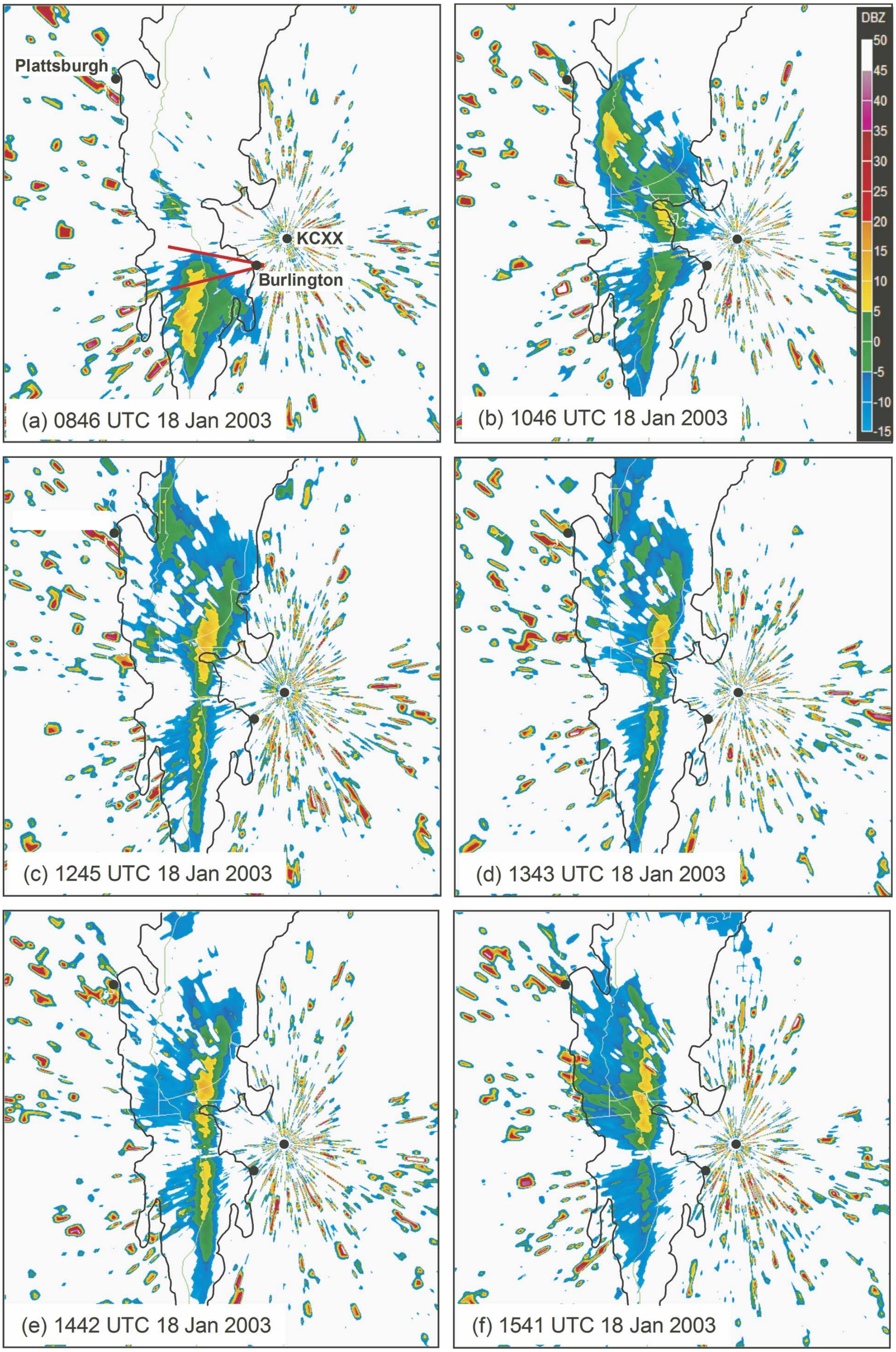

FIG. 2. WSR-88D radar reflectivity from KCXX on 18 Jan 2003. (a), (b) The earlier evolution of the snowband. (c)–(f) Corresponding times of the images shown in Fig. 1. The directional view and sector of the camera used to collect the images in Fig. 1 are shown in (a) along with locations of Burlington and Plattsburgh.

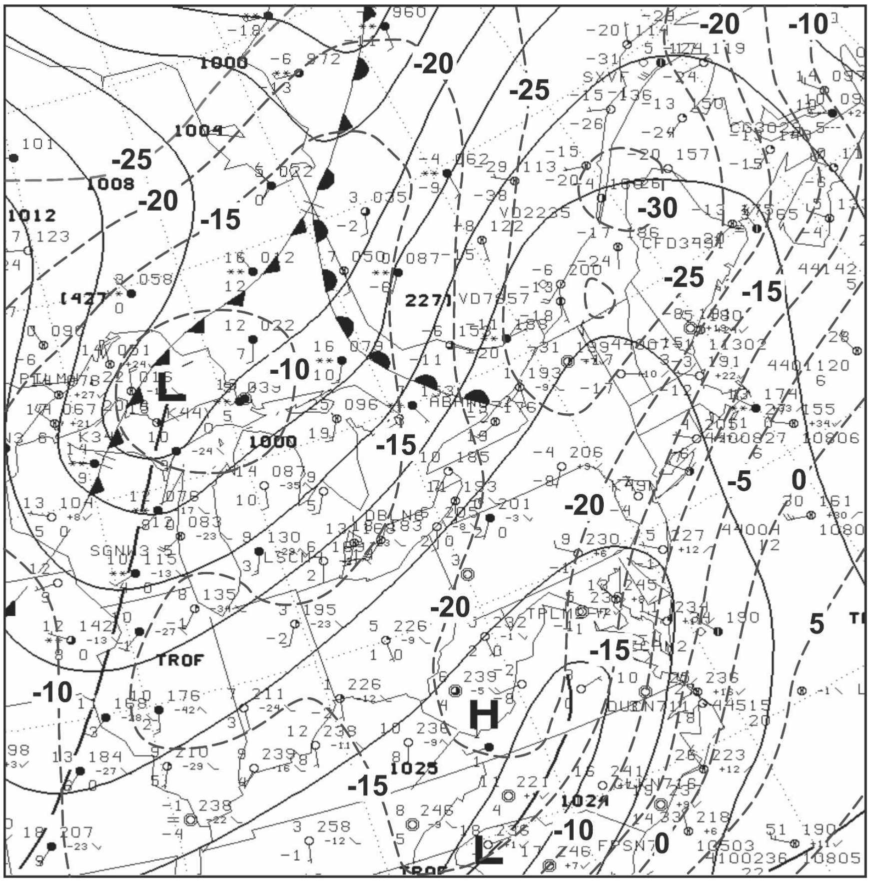

FIG. 3. Surface analysis at 1200 UTC 18 Jan 2003 with isobars (hPa; solid lines) and isotherms (°C; dashed lines) shown. The surface analysis was obtained from the National Centers for Environmental Prediction.

In the final hours of the event, the stationary surface trough over the Great Lakes region strengthened into a low pressure center. The weak warm front extending east of the Great Lakes passed through the Lake Champlain valley region (not shown) and provided a combination of conditions less favorable for maintaining the lake-effect snowband (discussed further in section 3b).

3. Mesoscale conditions

On 18 January 2003, the KCXX radar observed a lake-effect band over Lake Champlain with no other precipitation systems present. During the early evolution of the lake-effect band when air temperatures were reported to be about −25°C at both Burlington and Plattsburgh, radar reflectivity values reached nearly 20 dBZ within the band (Figs. 2c–e). While the snowfall rates within the mesoscale band were likely much less than values typically observed in Great Lakes lake-effect snowbands, a radar study performed by Boucher and Wieler (1985) to relate radar-measured reflectivity factors to snowfall rates suggests the lake-effect band observed on 18 January 2003 may have been producing snowfall rates of 7–14 mm h^{-1} over Lake Champlain. Unfortunately, the snowband remained positioned over Lake Champlain for nearly the entire event and did not intersect the shoreline until the band was dissipating at roughly 1800 UTC, so surface observations of snowfall were not collected.

a. Radar evolution of the snowband

At 0800 UTC the snowband formed over the lake southwest of Burlington and began to extend northward (Figs. 2a,b). The camera first captured the snowband at sunrise (about 1200 UTC, Fig. 1a), 4 h after it had formed. The KCXX radar reflectivity from about 1045–1245 UTC showed a "branched" structure to the developing snowband and segments along the eastern and western shorelines (Figs. 2b,c).

At approximately 1430 UTC, the snowband transitioned from a branched band into a well-defined, single, narrow area of precipitation over the center of the lake (Figs. 1c and 2e). Radar reflectivity within the band peaked near 20 dBZ and radar radial velocities indicated that low-level winds were from the south at ≤ 3 m s^{-1}. The section of the snowband south of Burlington remained stationary over central Lake Champlain throughout the event; while the northern section shifted position between the eastern and western overlake regions. The camera recorded clouds and cloud elements (Lyons and Pease 1972) extending down to the lake surface from about 1200 to 1530 UTC.

Over the last few hours of the event, the reduction in band coherence and radar reflectivity corresponded to the period when the cloud base became disconnected from the lake (Fig. 1d) and a wind shift positioned the precipitation over the western portion of the lake and shoreline (Fig. 2f).

b. Surface observations

Hourly observations were gathered from three stations. Saint Hubert Airport (CYHU) in Montréal was examined to provide information about the regional air mass without modification by Lake Champlain. Burlington International Airport (KBTV), located on the eastern shore in Burlington and Clinton County Airport (KPLB), located farther north on the western shore in Plattsburgh, provided meteorological conditions in the nearshore regions (<4 km) of Lake Champlain (Fig. 2a).

During the early hours of the event, from 0800 UTC to approximately 1500 UTC, the winds were calm or less than 2.5 m s^{-1} (Figs. 4a,b). Wind directions at both KBTV and KPLB showed a lakeward component and the lake-to-air temperature difference was nearly 26°C. Observations at CYHU showed a continual weak southerly wind (Fig. 4c). The weak southerly wind combined with enhanced low-level overlake convergence from approaching land breezes around 0800 UTC to provide the dynamic environment responsible for initiation, extension northward, and the branched structure of the snowband (Figs. 2a,b).

During the later portion of the event, sea level pressure decreased, southerly wind speeds increased, and temperatures steadily warmed at KBTV, KPLB, and CYHU (Fig. 4). For example, from 1400 to 1800 UTC the temperatures at both KBTV and KPLB warmed at a rate of about 2°C h^{-1} and wind speeds increased from 2 m s^{-1} at both stations to 6 m s^{-1} at KBTV and 4 m s^{-1} at KPLB. The land-breeze wind components at both KBTV and KPLB decreased significantly by 1400–1500 UTC and winds became more consistent with wind speeds and directions observed at CYHU throughout the event. This resulted in a transition in the snowband

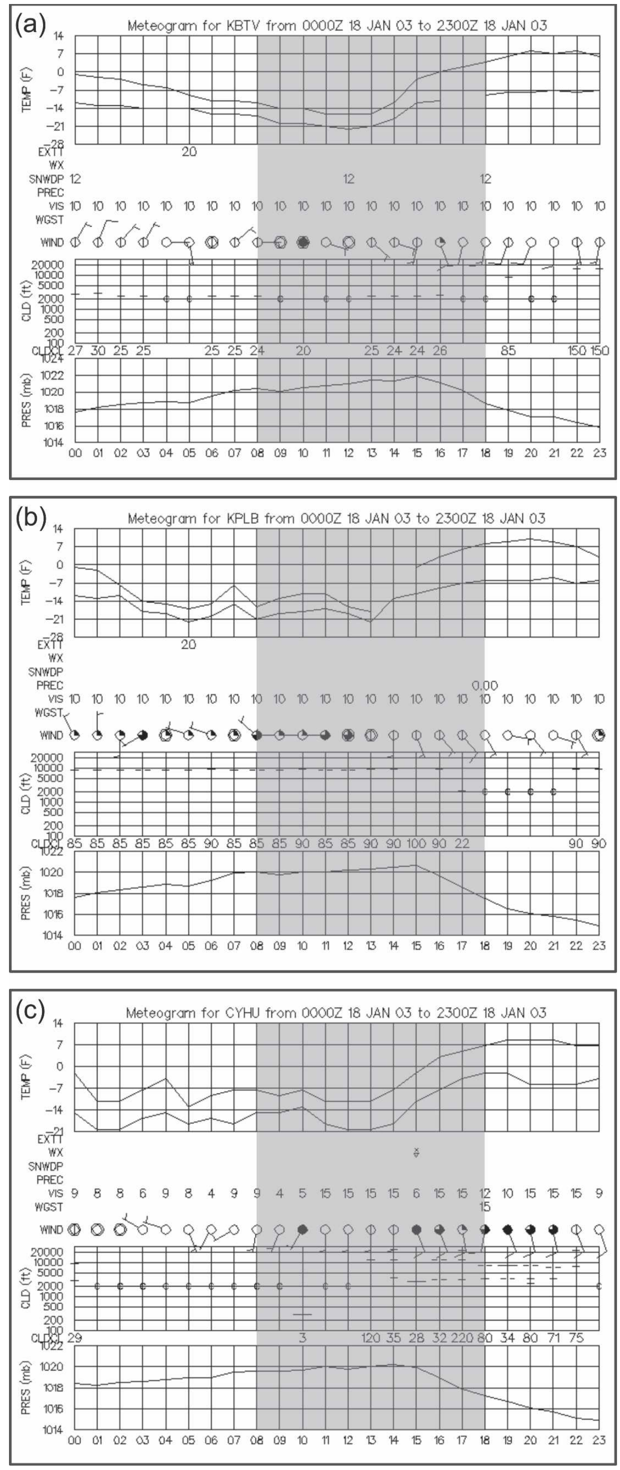

FIG. 4. Meteograms showing surface observations for (a) KBTV, (b) KPLB, and (c) CYHU on 18 Jan 2003. The shaded area designates the time period of the Lake Champlain lake-effect snowband. (Plots courtesy of Plymouth State University Meteorology Program.)

from a branched shore-parallel structure (Fig. 2c) forced primarily by land breezes to a single, wind-parallel, midlake snowband oriented along the major lake axis (Fig. 4e) and forced primarily by strengthening southerly winds and the weakening of land breezes. This change in wind conditions corresponded to the eastward advancement of the low pressure center that had been located in the Great Lakes region and the movement of the associated warm front through the Lake Champlain valley.

Also associated with this shifting synoptic pattern, several other changing environmental conditions combined to cause the dissipation of the snowband. By 1700 UTC, both KBTV and KPLB observations showed winds with an onshore component indicating that low-level divergence had developed over the lake. This change in low-level winds, along with a steady decrease of the lake-to-air temperature difference to about 14°C, provided a less than favorable environment and led to a rapid dissipation of the Lake Champlain lake-effect snowband by 1800 UTC. In addition, the 850-hPa temperatures had increased with warm advection in the Lake Champlain region and by 0000 UTC 19 January the lake to 850-hPa temperature difference had reduced to approximately 13°C, the minimum threshold found to support lake-effect snowbands.

4. Summary

The images of the lake-effect snowband presented provide an excellent complement to the WSR-88D radar data and show the visible evolution of the snowband over the southern portion of Lake Champlain on 18 January 2003 (Fig. 1). The WSR-88D radar and surface data indicate that the lake-effect snowband developed in association with land breezes on both the western and eastern shores of Lake Champlain and transitioned from a shore-parallel to midlake snowband in a unique situation with strengthening southerly winds. A climatological study of Lake Champlain lake-effect events is needed to address the exclusivity of this southerly wind snowband event and provide a greater perspective on the comparison of environments that support lake-effect snow over small lakes relative to large lakes.

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