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THE GRAFTON, VERMONT, FLOOD 12-13 JUNE 1996

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1. INTRODUCTION

Flooding is the number one weather-related killer in the United States with an average of 140 fatalities each year.[1] The severity of a flood depends on many factors including the amount and intensity of rain, antecedent soil conditions, the degree of foliation, the river channel's base flow and ice content, and the basin's topography (LaPenta et al. 1995). Flooding from convective systems is especially difficult to forecast for several reasons. Convective storms that produce flooding can occur on a small scale, can last for only a short time, and are usually poorly forecast by operational numerical models. During the afternoon and evening of both 12 and 13 June 1996, a series of thunderstorms which moved over northern Windham County, Vermont, produced severe flooding. The worst flooding was along the Stiles Brook, about 4 miles south-southwest of Grafton (Fig. 1). Flooding on 13 June was especially severe with many roads washed out and some areas completely isolated by flood waters. There were no fatalities, but Grafton was declared a federal disaster area with damage in excess of a million dollars.

Rainfall observations from the flood area were limited. The Albany, New York, (KENX) Weather Surveillance Radar - 1998 Doppler (WSR-88D) radar, located 70 n mi west-southwest of Grafton, estimated 6.2 inches of rain on 12 June southwest of Grafton with 3.2 inches in one hour. The radar estimated 5.5 inches northeast of Grafton on 13 June. Figure 2 is a 2-day (1400 UTC 12 June 1996 to 0000 UTC 14 June 1996) storm total WSR-88D precipitation estimate based on a pixel by pixel (1.2 n mi^2 resolution) summation of rainfall for the two events. The radar indicated a maximum 2-day total of 10.5 inches about 5 mi southwest of Grafton. During the same period, an observer located 4 miles south-southwest of Grafton (Fig. 1) measured 6.75 inches of rain (Fig. 2), 3.75 inches on 12 June and 3 inches on 13

[1]1960-1989 average based on statistics from the National Weather Service Office of Meteorology, Warning and Forecast Branch, Silver Spring, MD.

June. The radar estimated about 8 inches of rain at this point. This suggests a maximum 2-day rainfall of between 8 and 9 inches. Small hail may have contributed to the WSR-88D's over estimation of the rainfall.

The meteorological conditions that produced the flash flooding will be summarized. The analysis will identify the meteorological and topographical factors that influenced the thunderstorms. In addition, we will examine the performance of the National Centers' for Environmental Prediction operational numerical models.

2. METEOROLOGICAL CONDITIONS

The weather pattern responsible for the flooding became established nearly a week before the events as a 500 hPa low cut off over the center of the country (not shown). This upper-level low moved very slowly over the next few days before opening up across the northeastern states late on 13 June. A moist, southwest flow was established across the northeastern states on 8 June with localized severe thunderstorms and flash flooding, which continued through 13 June.

a. 12 June 1996

At 1200 UTC 12 June, the 500 hPa (Fig. 3a) low was centered near the western end of Lake Erie, according to the Aviation model (AVN). A vorticity maximum of $2.0 \times 10^{-4}\ s^{-1}$ indicated a short wave was swinging around the south end of the 500 hPa low. At 850 hPa, warm air (greater than 15°C) was located over eastern New England with progressively cooler air to the west (Fig. 3b). Dewpoints at 850 hPa (not shown) over central New England were near 10°C.

A surface analysis at 1400 UTC (Fig. 3c) showed a weak pressure gradient across the northeast U.S. A weak area of low pressure was centered over northeastern Pennsylvania with a trough extending southwest. A second trough extended east-northeast, across central New England, from the low. This trough became less discernable during subsequent hours, while the trough across Pennsylvania persisted.

The Albany 1200 UTC RAOB was modified based on observed surface temperatures and dewpoints to estimate conditions over southern Vermont during the afternoon of 12 June, assuming the rest of the sounding would not have changed. The modified sounding (Fig. 4a) indicated unstable air with Convective Available Potential Energy (CAPE) of 3059 J/kg and relatively weak flow. CAPE is the vertically integrated positive buoyancy of an adiabatically rising parcel. It was calculated by using the Skew-T Hodograph Analysis and Research Program (SHARP, Hart and Korotky 1991) and lifting a surface parcel to its equilibrium level. The wind veered from south at the surface to southwest at 700 hPa and was 15 kt or less in that layer. The flow remained southwest up to the tropopause with a maximum wind speed of 47 kt at 250 hPa. Precipitable water was 1.46 inches (about 150% of normal).

Precipitation efficiency is one factor that contributes to a thunderstorms's ability to produce very heavy rain. Chappell (1992) suggested the following factors are favorable for producing thunderstorms with a high precipitation efficiency:

1) moderate CAPE (1500-3000 J/kg); 2) a vertically elongated distribution of CAPE; 3) a moist environment with high precipitable water; and 4) light to moderate vertical wind shear. Chappell (1992) noted that a sounding

with these characteristics would favor a more slowly accelerating updraft, allowing more time for condensate to be converted to rain through the collision-coalescence (warm-rain) process. This type of sounding would reduce entrainment into the updraft, and a deeper warm cloud layer would favor a more efficient warm-rain process (Chappell 1992). Based on the modified Albany sounding (Fig. 4a), CAPE was just over 3000 J/kg with an elongated distribution from about 850 hPa to about 200 hPa. Precipitable water was about 150% of normal and the shear was small, less than $4 \times 10^{-3}\ s^{-1}$. A rising parcel would not reach 0°C until nearly 550 hPa and -10°C at about 450 hPa, indicating the warm-rain process could operate for a significant length of time. While dry air was evident above 700 hPa, overall the sounding would favor thunderstorms with a high precipitation efficiency.

Terrain forcing was most likely an important factor in determining where thunderstorms formed. The Green Mountains run north to south through the state of Vermont with the Connecticut River Valley to the east (Fig. 1). Differential heating between the ground over the mountains and the free atmosphere at the same elevation some distance away can produce a convergent upslope flow. Peilke and Segal (1986) suggest that with a light synoptic flow, daytime upslope flows are about 6-12 kt with a depth of several hundreds meters, or more. This convergent upslope flow likely caused thunderstorms to form along the spine of the Green Mountains. The storms moved slowly northeastward in the ambient southwest flow and eventually weakened. Visible satellite imagery showed thunderstorm formation was along and just east of the axis of the Green Mountains. At 1215 UTC (Fig. 5a) there was considerable low cloudiness in the Connecticut River Valley. While solar heating quickly dissipated this cloudiness, it likely delayed surface heating in that area, further enhancing the thermal boundary between the Connecticut River Valley and the mountains to the west. Figures 5b-d are visible images at 1702, 1815, and 2002 UTC respectively; showing thunderstorms developing and persisting along and just east of the mountains. The WSR-88D Storm Total Precipitation from 1400 UTC 12 June to 0100 UTC 13 June (Fig. 6a) indicates that much of the heavy precipitation in Vermont fell along and just east of the spine of the Green Mountains.

While satellite and radar imagery strongly suggest terrain induced convergence along the Green Mountains was an important forcing mechanism for thunderstorm generation on a regional scale, it is not clear why specific locations along the Green Mountain convergence zone (hereafter referred to as the GMCZ) were favored for cell development. KENX WSR-88D data were used to investigate this question, although there are limitations in the radar data. First, a full volume scan takes 5 to 6 minutes to complete. As a result, it was sometimes difficult to track individual cells from one volume scan to another. Subjective decisions were made as to whether an echo on a subsequent volume scan was actually a new cell or a cell from the previous scan. There also may have been a time lag (and resultant displacement of a cell) between the time convergence initiated cell development and the appearance of the first 30 dBZ reflectivity. Finally, the radar data vary in elevation. The 0.5° elevation scan samples data at about 4700 ft over the southwest corner of Vermont with the beam elevation increasing to 16000 ft in northeast Windsor County.

A number of tributaries extend west and northwest from the Connecticut River to their source regions in the mountains. These streams create smaller valleys (tributary valleys), which cut into the Green Mountain range. The tributary valleys might produce an enhanced convergence by increasing the strength of the convergent flow because of steeper terrain or a channeling of the upslope flow, thus providing specific locations along the GMCZ especially favorable for storm development. Once these storms developed, they generated outflow boundaries. Additional sites for convective development were created where the outflow boundaries intersected the GMCZ. Figure 7 shows the location of the first observed 30 dBZ echoes (0.5° elevation scan) with individual cells, between 1448 and 1736 UTC on 12 June, superimposed on a terrain map. Many of first 30 dBZ returns were noted over the upper (western or northwestern) portions of the tributary valleys.

While many cells developed in or near tributary valleys, there were exceptions. Two cells developed along the Bennington County and Windham County border over high terrain at 1730 UTC and 1736 UTC (Fig. 7). However, when these cells formed, convection had been ongoing for several hours. At that point, outflow boundaries may have been playing a more important role in the location of new cell formation. The importance of outflow boundaries in cell development along the GMCZ can be inferred in the pattern of cell development across northern Windsor County. At 1448 UTC a cell formed in a tributary valley, near the northern border of the county (Fig. 7) where the White River (Fig. 1) extends northwest into the Green Mountains. While the initial storm developed over a tributary valley, the outflow boundary to the rear of this storm initiated a series of cells to its southwest at 1454, 1517, 1546, 1610, and 1632 UTC.

Thunderstorms frequently produce high rainfall rates. However, total rainfall at a particular point depends on the areal coverage (or size) of the storm and the speed at which the storm moves. If thunderstorms move slowly or remain stationary, rainfall may be heavy enough to produce flooding. Often, convection is multicellular in nature, with individual convective elements moving with a speed related to the flow through the depth of the storm and evolving continuously (Chappell 1986). The formation and decay of individual elements results in motion of the multi-cellular system that is significantly different than the winds through the storm depth. For example, if re-generation of convective elements occurs on the rear storm flank, there is slower forward speed or even backward propagation of the multi-cellular convective system (Chappell 1986). Storm motion of a multi-cellular system can be considered to be the sum of the mean cell motion plus the propagation. For this case, radar showed the cell motion was northeast averaging 10-15 kt. Cells continuously developed on the southwest flank of the system, which resulted in a very slow movement and at times backward propagation of the system.

KENX radar reflectivity data showed that scattered cells formed over northern Windham County shortly before 1600 UTC (Fig. 7). Between 1736 and 1900 UTC most of the thunderstorms remained just north of Grafton. The area of reflectivity greater than 50 (50+) dBZ increased markedly after 1830 UTC (not shown). New storms formed on the rear (southwest) flank of decaying storms, and between 1900 and 2100 UTC there was growth back toward the southwest. After 2105 UTC, 50+ dBZ returns were concentrated between Grafton and Jamaica (not shown). Large areas of 50+ dBZ reflectivity were observed between 2145 and

2226 UTC and the radar estimated 3.2 inches of rain just southwest of Grafton between 2100 and 2200 UTC. The thunderstorms weakened significantly between 2226 and 2306 UTC.

b. 13 June 1996

On 13 June, the 500 hPa low that had remained over the center of the U.S. for nearly a week opened up and began to move northeastward. At 1200 UTC, the 500 hPa vorticity maximum (Fig. 8a) was just east of Lake Huron and a ridge of vorticity extending southward along the Appalachian Mountains, according to the AVN. By 0000 UTC 14 June, it had moved into southern Quebec and New York (not shown). At 1200 UTC 13 June, an axis of moist air at 850 hPa (Fig. 8b) with dewpoints of 10 to 12°C lay across eastern New York and western New England.

Although the surface pressure pattern remained weak, a more organized trough approached the Hudson Valley by 1600 UTC (Fig. 8c). It moved into New England by 0000 UTC 14 June.

The Albany 1200 UTC 13 June RAOB (Fig. 4b) was modified based on observed surface temperatures and dewpoints to estimate conditions over southern Vermont during the afternoon of 13 June, assuming the rest of the sounding had little change. The CAPE was 2630 J/kg with a precipitable water of 1.58 inches. Winds were south to southwest throughout the troposphere and were 15 kt or less below 700 hPa. Winds increased to 61 kt at 250 hPa. The sounding again appeared favorable for producing thunderstorms with high precipitation efficiency. CAPE was within the range indicated by Chappell (1992) and fairly evenly distributed from the LCL to about 200 hPa. Precipitable water was high (greater 150% of normal) and shear was small (about 4 x 10^{-3} s^{-1}). The airmass was nearly saturated through 700 hPa and the temperatures of rising parcels favored the warm-rain process to operate through a deep tropospheric layer.

Satellite imagery (not shown) indicated widespread high level cloudiness across southern Vermont through 1400 UTC. The high cloudiness moved off to the east after 1400 UTC, and increased sunshine heated the lower troposphere. Thunderstorms developed between 1500 and 1600 UTC. Terrain again played a role as storms developed along the spine of the Green Mountains in the GMCZ. In addition, heavy rain on the previous day may have contributed to increased low-level moisture across the Green Mountains. Figure 9 shows the location of the first observed 30 dBZ echoes with individual cells between 1500 and 1600 UTC on 13 June, superimposed on a terrain map. Once again there was a tendency for tributary valley locations to be favored sites along the GMCZ for thunderstorm genesis. By 1553 UTC, small 50 dBZ cells (not shown) had developed between Grafton and Jamaica. Redevelopment on the rear (southwest) flank of decaying thunderstorms produced a quasi-stationary convective system with 50+ dBZ returns located between Grafton and Jamaica through 1840 UTC. By 1903 UTC, the 50+ dBZ echoes propagated eastward, and by 1934 UTC they had moved well east of the Grafton area and weakened considerably. Figure 6b shows the radar estimated storm total precipitation for 13 June 1996, which again illustrates the importance of the Green Mountains in the thunderstorm development.

3. NUMERICAL MODEL PERFORMANCE

Model performance for the two events varied considerably. Gridded output from the Aviation (AVN), Regional (RFS), and early ETA (ETA) models were analyzed.

a. 12 June 1200 UTC Forecast Cycle

Model forecasts for 1200 UTC 12 June were available from the AVN and RFS Models. The 500 hPa initialization showed only very small differences between the two models. Both produced a trough over the eastern Great Lakes and Northeast U.S. through 48 hours. The trough was forecast to move slowly east and open up as two shortwaves rotated through the base of the trough. The AVN resolved two distinct shortwaves initially and merged them as they exited the base of the trough. The RFS had one broader and slower moving shortwave. The RFS was slightly faster with the trough; however, both models predicted slow movement. The AVN forecast a 500 hPa jet maximum over West Virginia to approach eastern Pennsylvania by evening on 12 June.

Both models initialized a weak surface low in southeastern Michigan with a weak surface tough from northwest to southeast across Pennsylvania (not shown). There was weak ridging over the Hudson Valley. The AVN had all three surface features more distinct and with a bit more amplitude than the RFS. The AVN even showed a weak warm front extending across the St. Lawrence River Valley. By forecast hour 12 (0000 UTC 13 June), the RFS deepened the surface low by 1 hPa and placed it in western Pennsylvania. The AVN no longer identified this feature. Both models placed a moderate southwesterly gradient across southern New York and northeast Pennsylvania.

Throughout the forecast cycle, the AVN had more moisture at 1000 hPa. At forecast hour 12 (0000 UTC 13 June), the AVN had 1000 hPa mixing ratios between 17 and 18 g/kg across the Green Mountains and the northeast Catskills as compared to the RFS, which predicted 14 g/kg.

Model quantitative precipitation forecasts (QPFs) showed significant differences. The AVN 12 hour QPF (1200 UTC to 0000 UTC, Fig. 10a) shows an area of 0.3 inches or greater precipitation from near Albany to the mountains of western New England with a 0.5 inches maximum over the southern Green Mountains. During the same period the RFS forecast a maximum of 0.5 inches over northeastern Pennsylvania and less than 0.1 inches of rain from the Hudson Valley to the Connecticut River Valley (Fig. 10b).

Looking ahead to the second day of the event on the afternoon of 13 June, the AVN outperformed the RFS on the 24 to 36 hour QPF. The AVN showed a band of precipitation of about 0.5 inches across much of Vermont and the Berkshire Mountains of Massachusetts (Fig. 10c). The RFS forecast a general 0.1 to 0.2 inches (Fig. 10d).

b. 13 June 0000 UTC Forecast Cycle

The AVN, RFS, and ETA were all available for the 0000 UTC 13 June forecast cycle. The 500 hPa initialization showed several differences among the three models. The ETA had the weakest shortwave, it was initially located over northeast Ohio. The AVN predicted the strongest shortwave. The AVN also had the trough somewhat negatively tilted. All three models had the core of strongest 500 hPa winds just off the New York and New Jersey coast. The 500 hPa shortwave was forecast to move into the region between 0000 UTC 13 June and 0000 UTC 14 June.

The initial surface analyses for all three models had a trough extending from eastern Lake Erie to northern Virginia. There was weak ridging across eastern New York and Vermont. The AVN and ETA defined both features better than the RFS did. By forecast hour 12 (1200 UTC 13 June), the AVN and RFS had a well defined trough lifting north and east across the Hudson Valley. The ETA had a much weaker trough than the RFS or AVN. This trough continued to lift north and east moving into central New England by 0000 UTC. All models had a southwesterly gradient ahead of the trough. The AVN and RFS were initialized with large amounts of low level moisture. The ETA had an ambiguous and diffuse 1000 hPa moisture pattern. Again, the initialized AVN was the most moist with 1000 hPa mixing ratios between 16 and 18 g/kg across the Green Mountains. At the same time ETA had mixing ratios between 12 and 14 g/kg.

Model QPFs had significant differences. The AVN 12-24 hour QPF (1200 UTC 13 June to 0000 UTC 14 June, Fig. 11a) showed a band of precipitation in excess of 0.6 inches along the ridges of the northeast Catskills, the Berkshires, and the Green Mountains. There was a 0.8 inches maximum located in extreme southern Vermont, near the flood area. The RFS and ETA forecast 0.1 to 0.3 inches rainfall (Figs. 11b and 11c respectively). The AVN outperformed the ETA and RFS by predicting more rain and better pinpointing the rainfall maximum. The QPFs of all models were underdone.

c. 13 June 1200 UTC Forecast Cycle

The 1200 UTC 13 June forecast data were available from the AVN and RFS. The 500 hPa initialization showed only small differences between the two models. Both models forecast the 500 hPa trough located over the eastern Great Lakes to open up slightly and become somewhat negatively tilted.

The initial AVN surface analysis had a better defined and sharper trough, especially across south-central New York and northern Pennsylvania. Both models forecast the trough to move east during the day. By forecast hour 12 (0000 UTC 14 June), the AVN still had a much sharper and better defined surface feature. There was also a much stronger southwesterly gradient ahead of the trough in the initial and 12 hour AVN forecast. At 1000 hPa both models showed pooling of low level moisture along the southern Green Mountains with the RFS slightly more moist ahead of the trough. The RFS was much drier west of the surface trough.

Model QPFs were markedly different. The AVN 12 hour forecast (through 0000 UTC 14 June, Fig. 12a) showed a band of heavy rain (greater than 0.5 inches) across most of Vermont and the northern Berkshire Mountains. Although the QPF was underdone, the AVN placed a maximum of 0.7 inches in southern Vermont. The RFS 12 hour QPF forecast (Fig. 12b) was quite poor, showing a broad area of rain greater than 0.25 inches extending from northeast Vermont and New Hampshire northward.

4. DISCUSSION

During the afternoons of both 12 June and 13 June 1996, thunderstorms repeatedly moved over parts of Windham County producing severe flooding. The thunderstorms formed in a moist (precipitable water greater than 1.4 inches) and unstable (CAPE greater than 2500 J/kg) airmass. As solar insolation heated the unstable airmass, convection developed with terrain forcing likely a major factor in determining where thunderstorms formed. Thunderstorms were concentrated along the spine of the Green Mountains in the GMCZ. Tributary valleys may have produced areas of enhanced convergence in the GMCZ creating areas especially favorable for cell development. Once cells developed, thunderstorm outflows played a role in the evolution of the convective system. Regeneration on the southwest flanks, where thunderstorm outflows intersected the GMCZ, resulted in nearly stationary convective systems that dumped excessive rains over the Grafton area on successive days.

Maddox et al. (1979) discussed the meteorological patterns associated with quasi-stationary convective events that produce flash flooding east of the Rockies. They grouped these flash floods into three categories: synoptic, frontal, and mesohigh. The Grafton event did not fit into any of these distinct categories.

Forecasters during the events recognized the potential for heavy rain producing thunderstorms, but were unable to pinpoint the threat area prior to storm development. Numerical models showed little skill in forecasting the heavy rain with only the AVN model predicting a rainfall maxima over the Green Mountains on both 12 June and 13 June.

Experimental mesoscale numerical models have shown increased ability to resolve small-scale precipitation events. Two reasons for this improvement are the better resolution of terrain features and better parameterization schemes (Black 1994, Wesley et al. 1996, and Kalnay et al. 1996). Since this event was terrain enhanced, improved definition of terrain features may produce improved precipitation forecasts, both quantitatively and spatially, by resolving mesoscale phenomena such as mountain-valley circulations and terrain induced convergence (Wesley et al. 1996). Future research efforts on this storm will include analyzing its predictability using a high resolution mesoscale numerical model. Figure 13 shows the 48-km early ETA model terrain which depicts a broad area of high terrain (greater than 400 m) from north-central Pennsylvania across east-central New York to northern New England. Features such as the Hudson River Valley, the Green Mountains and the Connecticut River Valley are not resolved.

In addition, convection is not explicitly calculated by operational numerical models but estimated through convective parameterization. As computer power increases, newer models will be able to include improved parameterization schemes with better physics and fewer approximations (Perkey 1986). A portion of the model's QPF is generated by convective parameterization schemes. RFS output from this event indicated almost all of the model precipitation was produced by convection parameterization. Even with improved terrain definition and better parameterization, the forecast skill of mesoscale models may be limited by the lack of high quality mesoscale meteorological data (Perkey 1986).

5. CONCLUSION

The flash flooding which occurred across the southern Green Mountains on 12 June and 13 June, 1996 highlights several factors that meteorologists must be aware of when trying to identify potential convective flash flood events. In an unstable atmosphere with little synoptic scale forcing, we need to be cognizant of the mesoscale features which often control the development and evolution of convection. Operational numerical models usually do not resolve mesoscale features. In addition they do a poor job of forecasting convectively generated heavy rain since they do not explicitly calculate convective rainfall but approximate it through parameterization. Current operational models do not accurately resolve important terrain features and terrain forcing may play a crucial role in the development and evolution of convective systems, especially during times of weak synoptic scale forcing. In this case, convection formed as a result of various mesoscale features such as terrain forcing and convergence along outflow boundaries.

The movement of a multi-cellular convective system can be significantly different than the movement of individual elements. During this event, individual cells moved northeast in the ambient southwest wind flow. However, the combination of the terrain induced GMCZ and development along thunderstorm outflow boundaries resulted in a quasi-stationary system that concentrated heavy rain over a small area.

Research is a crucial component in our ability to improve the prediction of convectively generated flash floods. The analysis of events (and non-events) can help forecasters improve their understanding of the large-scale conditions that favor flash flooding; and of the mesoscale processes that may play a critical role in the development and evolution of flash flood producing storms. As meso-scale models with improved terrain depiction and better convective parameterization are developed, we must carefully evaluate their ability to accurately depict small scale events.

6. ACKNOWLEDGMENTS

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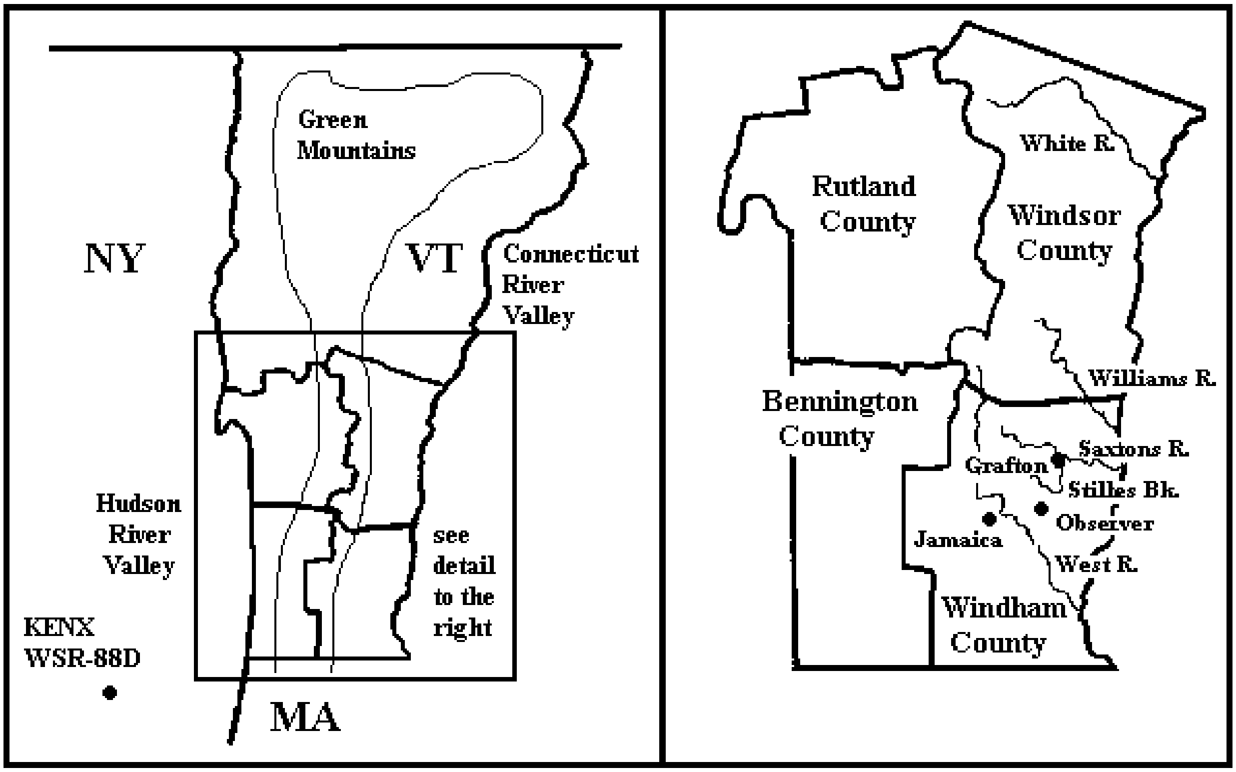

Figure 1. Geographical locations in eastern New York and western New England. Shaded area represents the higher terrain of the Green Mountains.

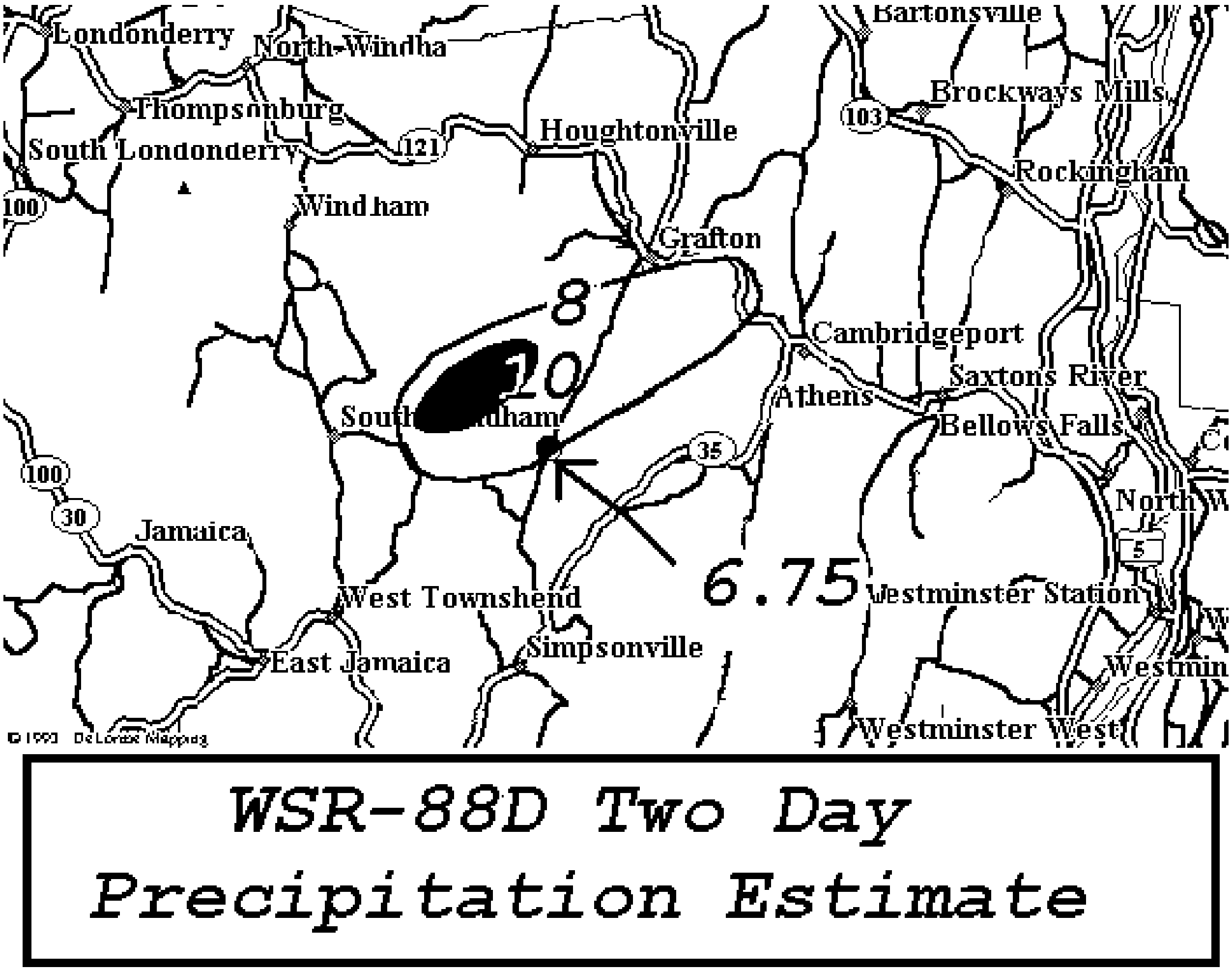

Figure 2. The two day storm total WSR-88D(KENX) precipitation estimate for 1400 UTC 12 June to 0000 UTC 14 June 1996. The radar indicated a maximum of 10.5 inches about 5 mi southwest of Grafton. An observer located 4 mi south-southwest of Grafton had a 2-day total of 6.75 inches.

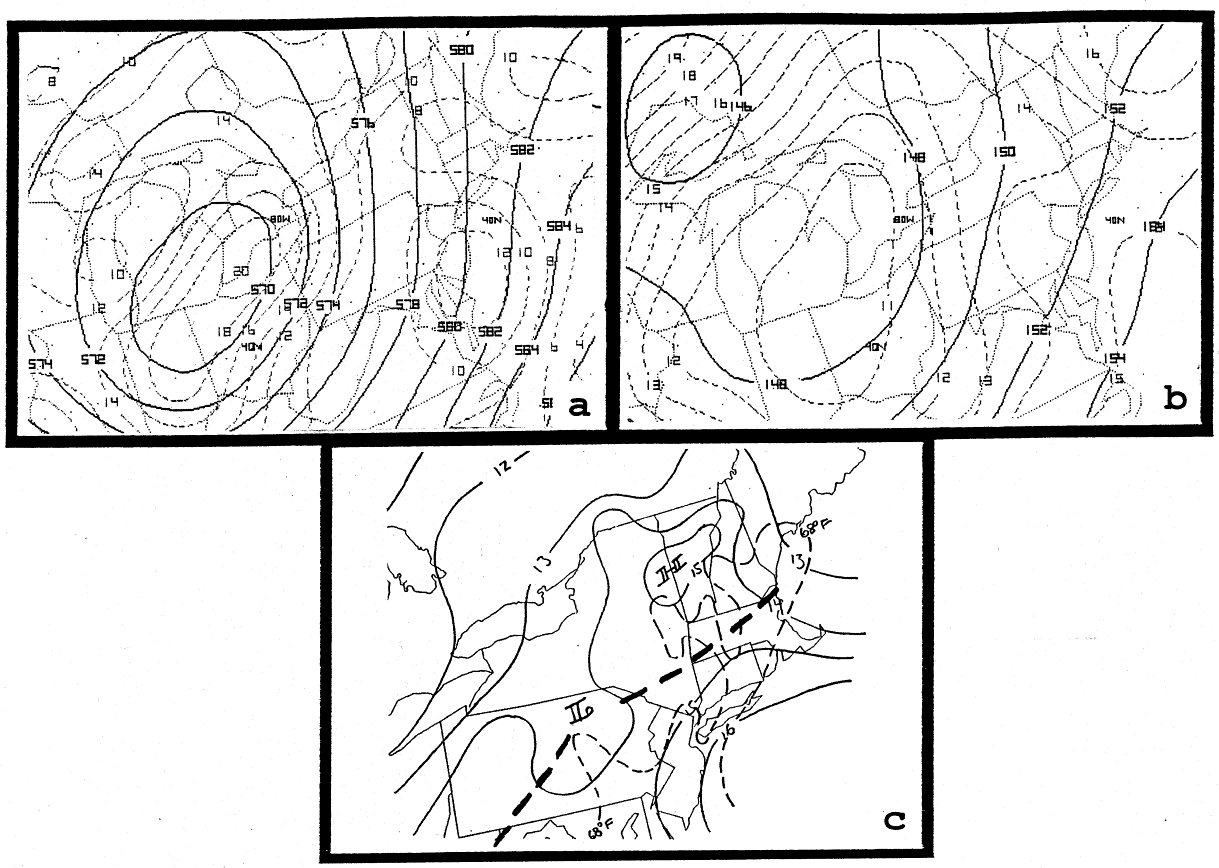

Figure 3. a) AVN analysis for 500 hPa at 1200 UTC 12 June 1996 with solid lines height contours (dm) and dashed lines vorticity (10^{-5} sec^{-1}). b) AVN analysis for 850 hPa at 1200 UTC 12 June 1996 with solid lines height contours (dm) and dashed lines temperature (°C). c) Surface analysis for 1400 UTC 12 June 1996 with solid lines pressure (hPa + 1000), thin dashed line representing the 20°C isodrosotherm, and the thick dashed line the location of the surface trough.

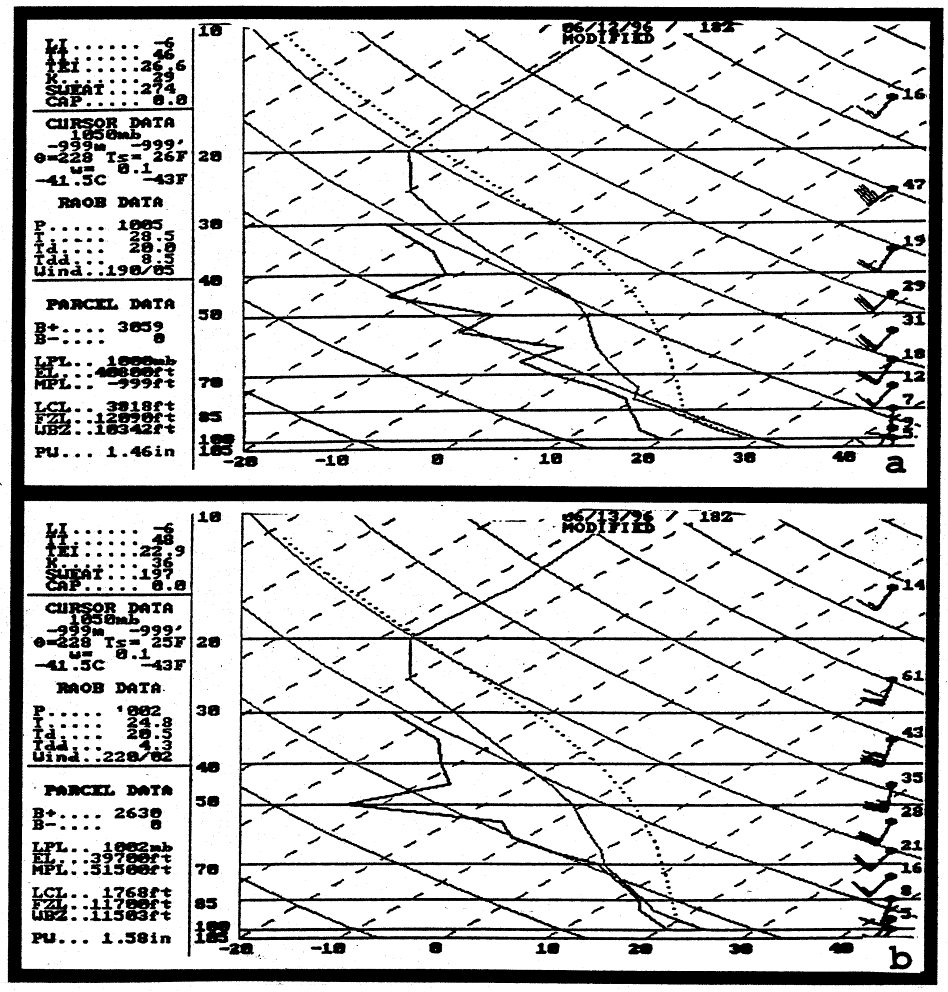

Figure 4. Modified Albany soundings for a) 12 June 1996 and b) 13 June 1996. Soundings were modified for observed surface temperature and dewpoint over southern Vermont near the time of convective development.

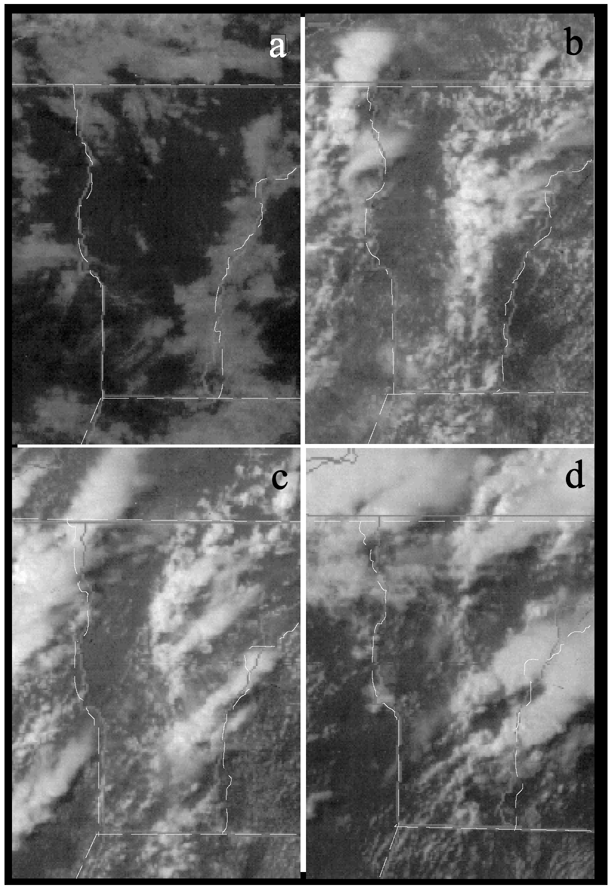

Figure 5. Visible satellite imagery from a) 1215 UTC, b) 1702 UTC, c) 1815 UTC, and d) 2002 UTC 12 June.

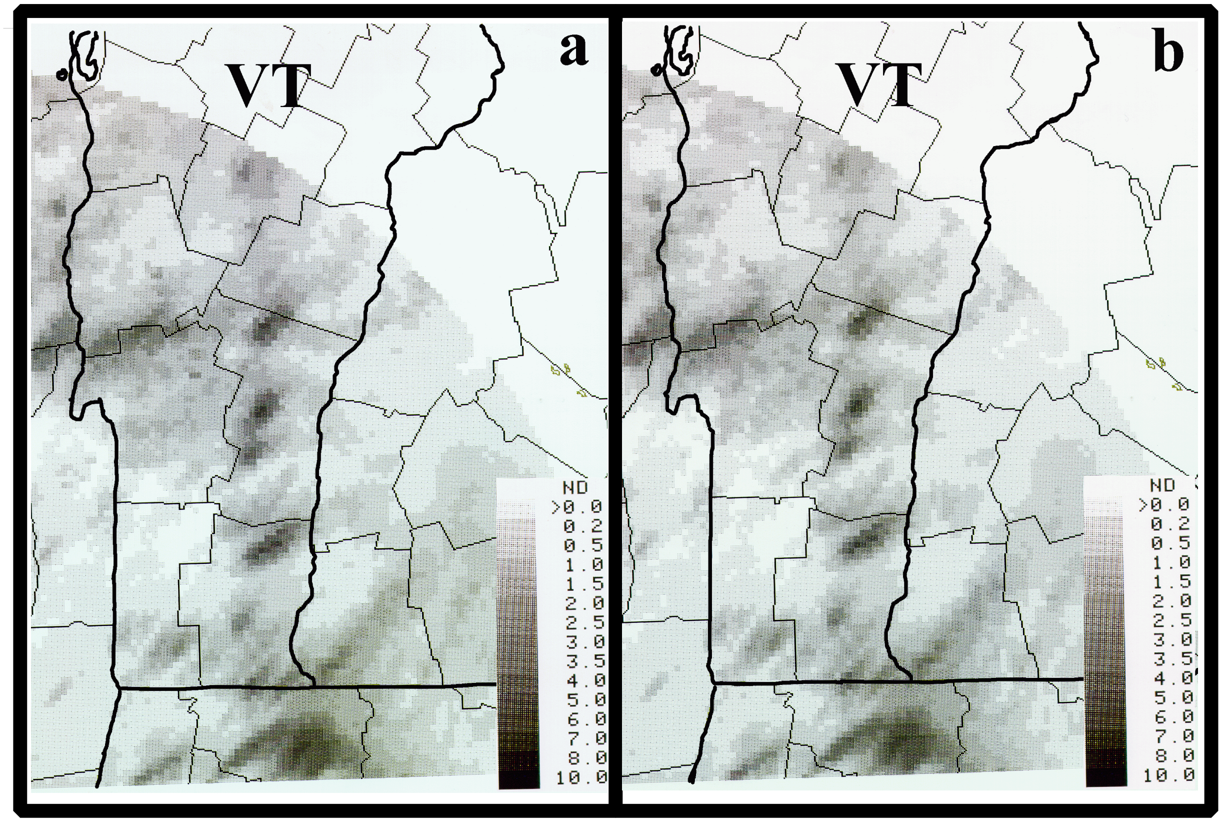

Figure 6. WSR-88D KENX storm total precipitation(inches) for a) 1535 UTC 12 June to 0435 UTC 13 June and b) 1455 UTC 13 June to 0009 UTC 14 June.

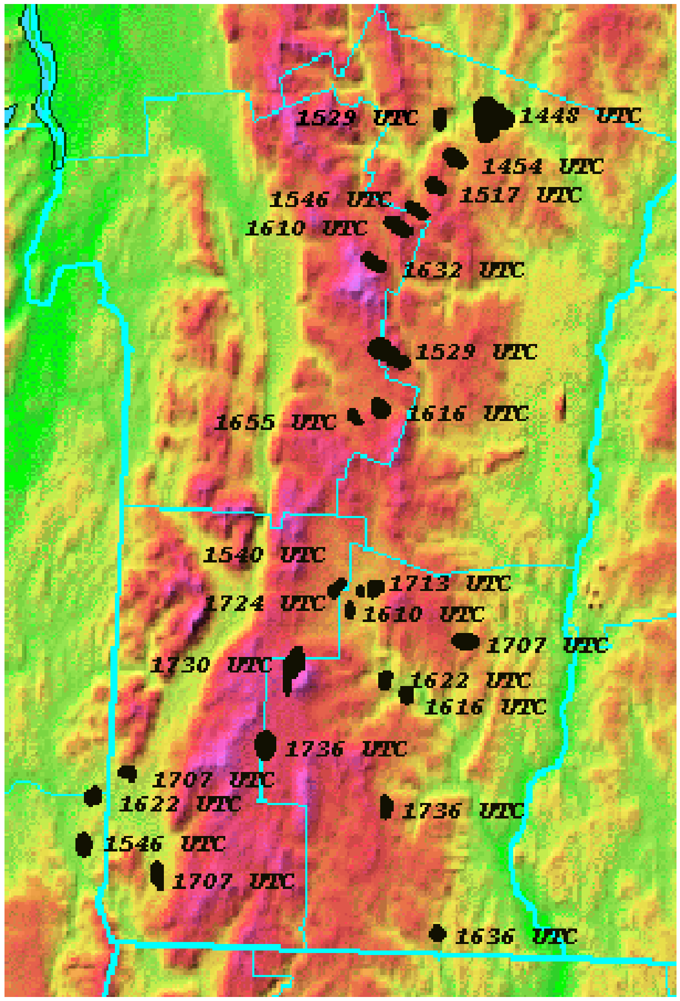

Figure 7. The location of the first observed 30 dBZ echoes (0.5° elevation scan KENX radar) associated with individual cells between 1448 and 1736 UTC on 12 June, superimposed on a terrain map of southern Vermont (see Fig. 1). Dark shades indicate higher terrain, lighter shades lower elevations.

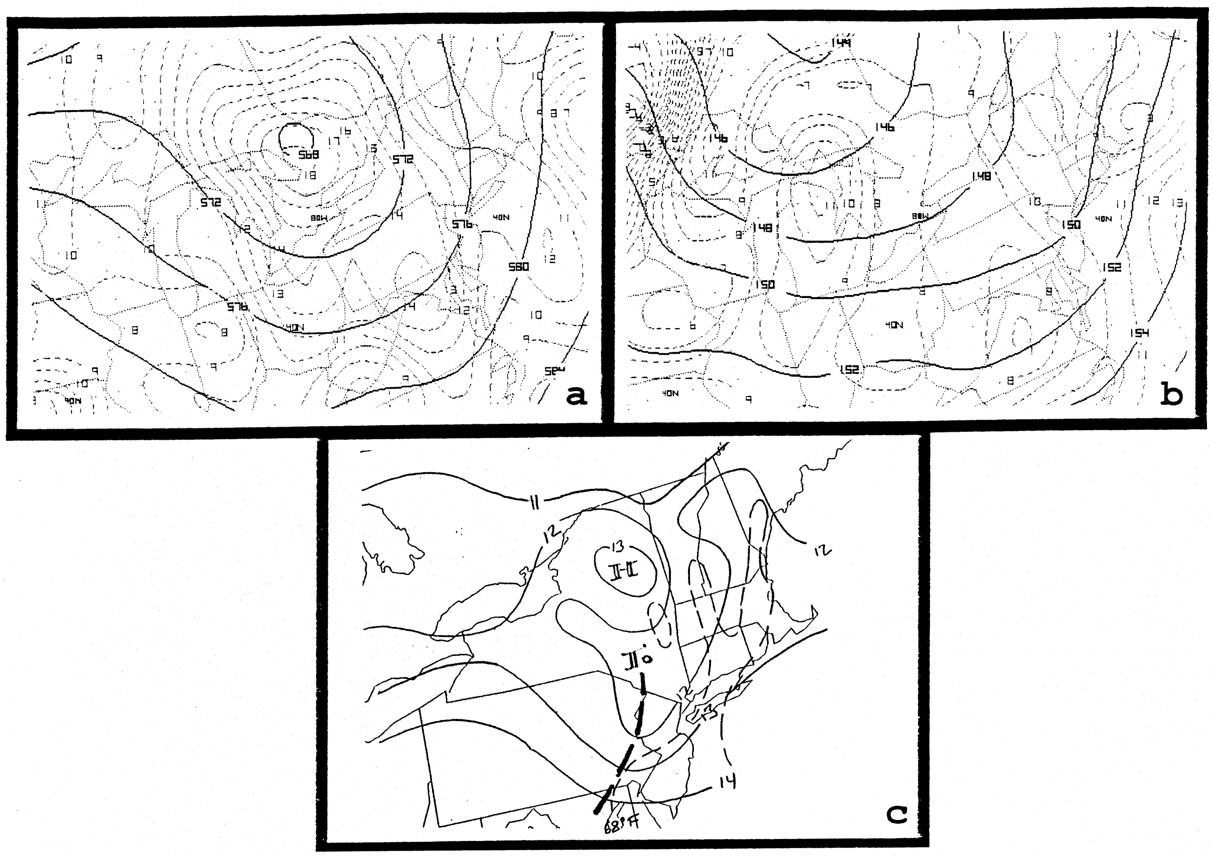

Figure 8. a) AVN analysis for 500 hPa at 1200 UTC 13 June 1996 with solid lines height contours(dm) and dashed lines vorticity (10^{-5} sec^{-1}). b) AVN analysis for 850 hPa at 1200 UTC 13 June 1996 with solid lines height contours (dm) and dashed lines dewpoint temperature (°C). c) Surface analysis for 1600 UTC 13 June 1996 with solid lines pressure (hPa +1000), thin dashed line representing the 20°C isodrosotherm, and the thick dashed line the location of the surface trough.

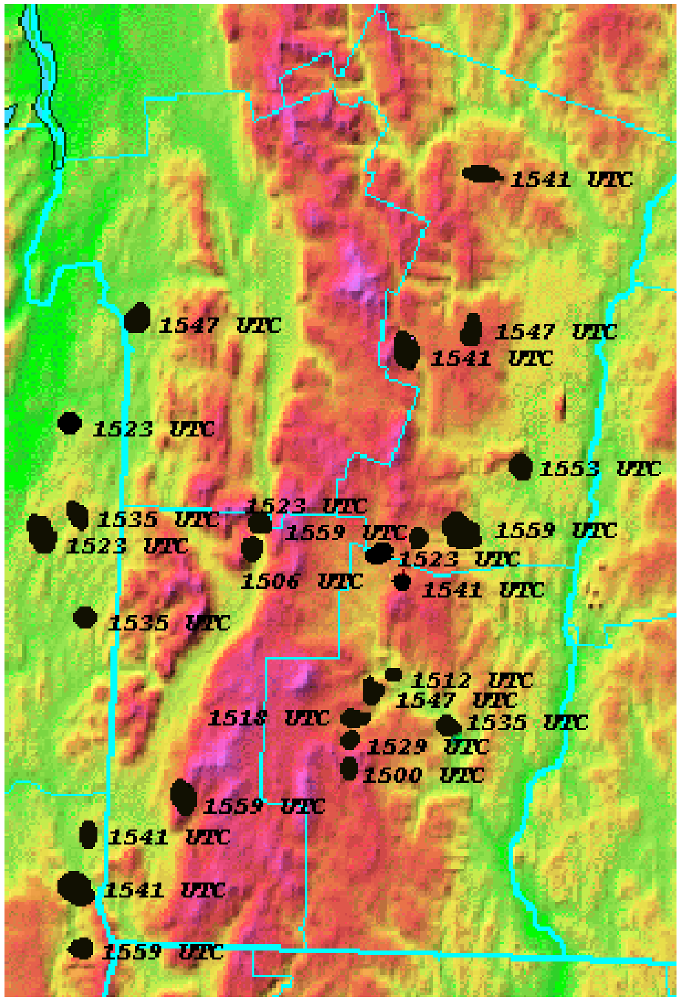

Figure 9. The location of the first observed 30 dBZ echoes (0.5° elevation scan KENX radar) associated with individual cells between 1553 to 1934 UTC on 13 June, superimposed on a terrain map of southern Vermont (see Fig. 1). Dark shades indicate higher terrain, lighter shades lower elevations.

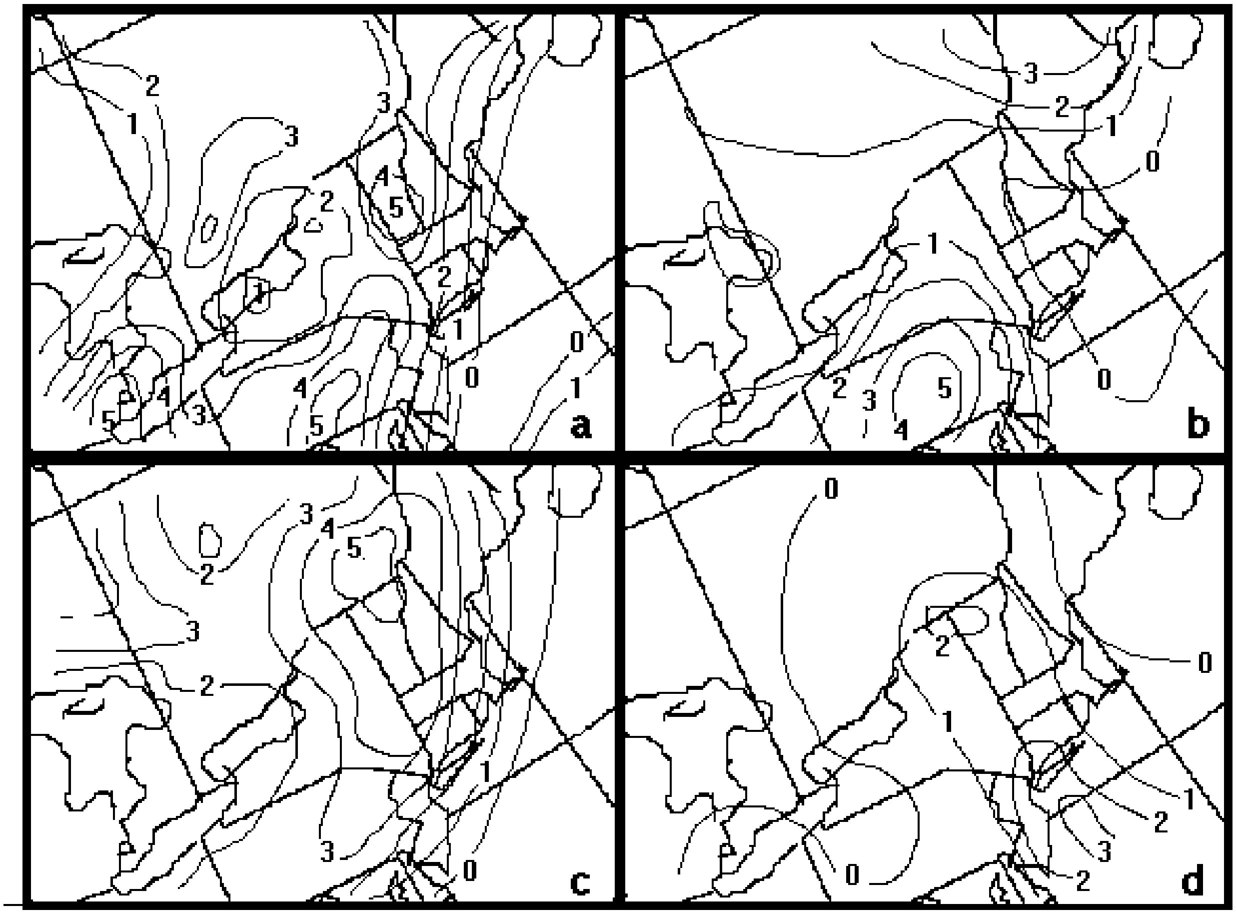

Figure 10. The QPFs (contours 0.1 inches) based on 1200 UTC 12 June numerical model output. AVN (a) and RFS (b) forecasts are shown for the 12 hr period ending 0000 UTC 13 June 1996. The AVN (c) and RFS (d) forecasts for the 12 hr period ending 0000 UTC 14 June 1996.

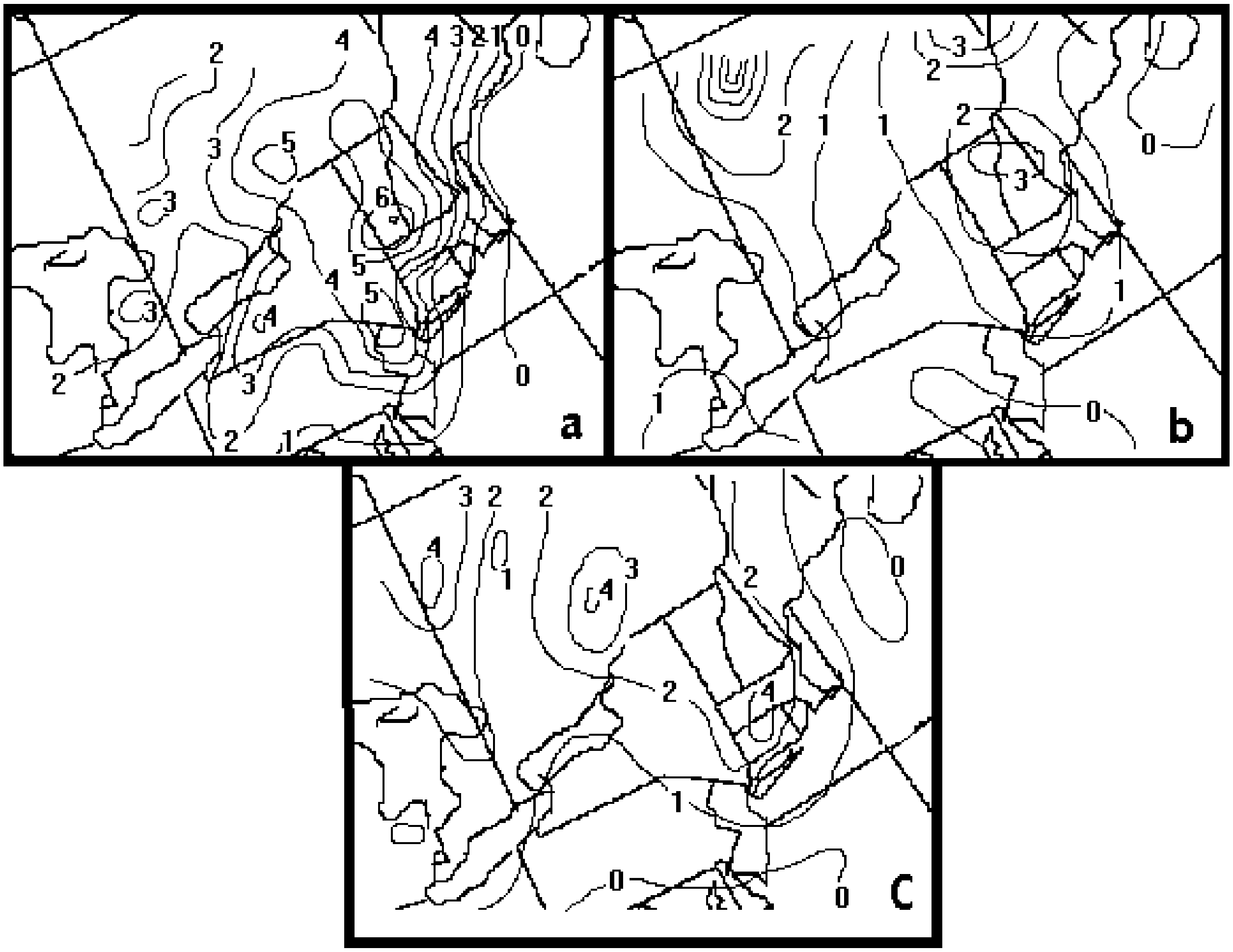

Figure 11. The QPFs (contours 0.1 inches) based on 0000 UTC 13 June numerical model output. AVN (a), RFS (b), and ETA (c) forecasts are shown for the 12 hr period ending 0000 UTC 14 June 1996.

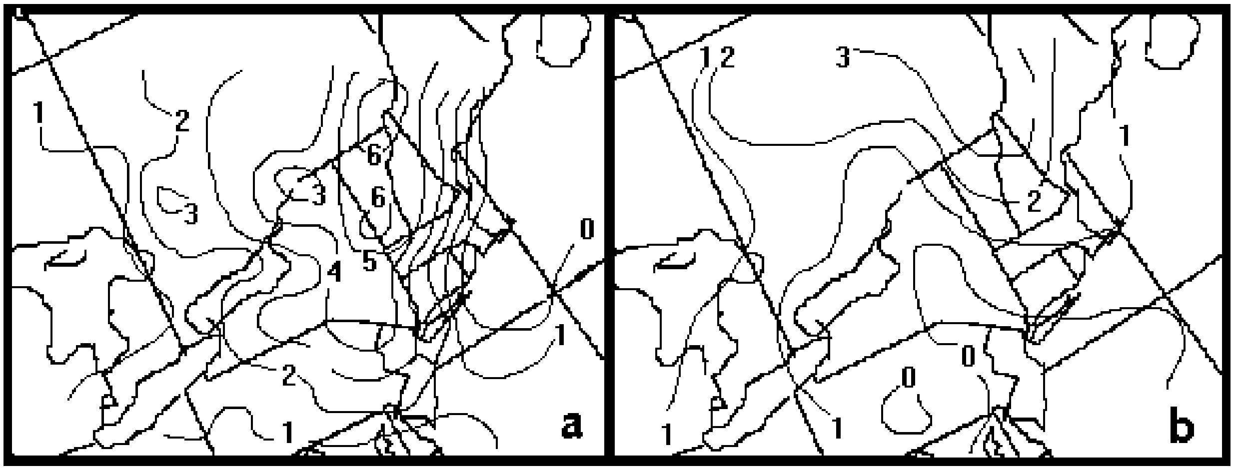

Figure 12. The QPFs (contours 0.1 inches) based on 1200 UTC 13 June numerical model output. AVN (a) and RFS (b) forecasts are shown for the 12 hr period ending 0000 UTC 14 June 1996.

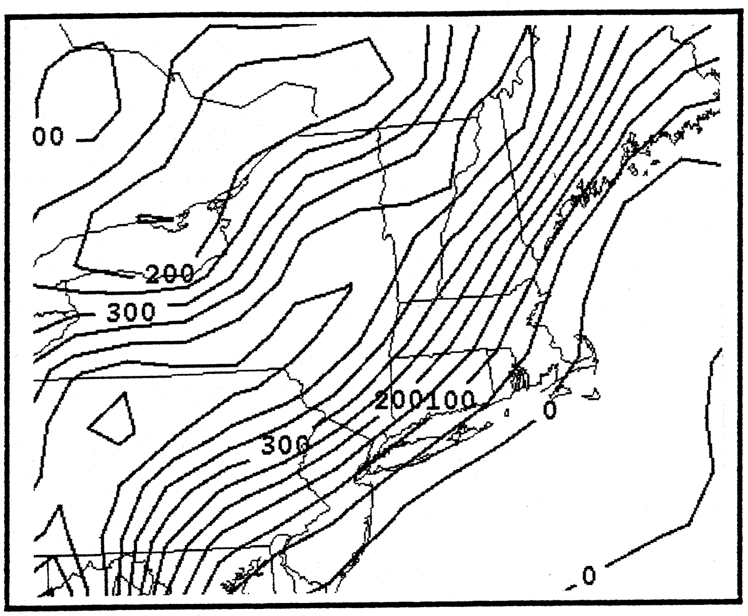

Figure 13. Model terrain for the 48-km early ETA with terrain increments of 50 m.