The Madison County, Virginia, Flash Flood of 27 June 1995

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ABSTRACT

Between 25 and 27 June 1995, excessive rainfall and associated flash flooding across portions of western Virginia resulted in three fatalities and millions of dollars in damage. Although many convective storms occurred over this region during this period, two particular mesoscale convective systems that occurred on 27 June were primarily responsible for the severe event. The first system (the Piedmont storm) developed over Madison County, Virginia (eastern slopes of the Blue Ridge Mountains), and propagated slowly southward producing 100–300 mm of rain over a narrow swath of the Virginia foothills and Piedmont. The second system (the Madison storm) developed over the same area but remained quasi-stationary along the eastern slopes of the Blue Ridge for nearly 8 h producing more than 600 mm of rain.

Analysis of this event indicates that the synoptic conditions responsible for initiating and maintaining the Madison storm were very similar to the Big Thompson and Fort Collins floods along the Front Range of the Rocky Mountains, as well as the Rapid City flood along the east slopes of the Black Hills of South Dakota. In all four events, an approaching shortwave aloft coupled with high-level difluence/divergence signaled the presence of local ascent and convective destabilization. A postfrontal ribbon of relatively fast-moving high-θ_e air, oriented nearly perpendicular to the mountain range, provided a copious moisture supply and helped focus the convection over a relatively small area. Weak middle- and upper-tropospheric steering currents favored slow-moving storms that further contributed to locally excessive rainfall.

A conceptual model for the Madison–Piedmont convective systems and their synoptic environment is presented, and the similarities and differences between the Madison County flood and the Big Thompson, Fort Collins, and Rapid City floods are highlighted.

1. Introduction

During the week of 25 June 1995 excessive rains fell over much of the mid-Atlantic region. Weak steering currents and very moist air contributed to periodic, slow-moving thunderstorms. The heavy rainfall saturated the ground and set the stage for severe flash flooding. On 27 June, a weak cold front drifted slowly southward into the mid-Atlantic states, triggering additional thunderstorms with heavy rain. Early in the morning, a convective system developed over Madison County and drifted southward over the Virginia Piedmont. Soon after this first system exited the county, a second system formed over northeastern Madison County, intensified, and propagated southwestward along the eastern slopes of the Blue Ridge (Fig. 1). This system remained quasi-stationary over the western portions of the county, feeding off moist, conditionally unstable air flowing westward from the Atlantic Ocean. Heavy downpours inundated the Rapidan River's watershed throughout the morning and early afternoon causing record flooding. According to Smith et al. (1996), basin-average rainfall reached 344 mm.

The flood waters receded slowly on 28 June, revealing a grim scene of devastation. Massive scars, visible for miles, highlighted the paths where mountainsides had liquefied and flowed into the valleys. These ''debris flows,'' which consisted of boulders, trees, mud, and water, swept downhill and destroyed everything in their path. Over 400 roads were closed, and 80 bridges and 2000 homes were damaged or destroyed (Virginia Department of Emergency Services 1998). There were three fatalities, 20 people sustained injuries, and nearly 800 residents were evacuated (NOAA 1995). The economic toll was equally staggering. Total losses in Virginia, excluding cleanup costs, exceeded 200 million dollars (Virginia Department of Emergency Services 1998). The hardest hit area was Madison County where damage to property and agriculture was estimated at 93 million dollars. Terrain scars and property damage were still evident a full year later.

Figure 2 shows the precipitation amounts that fell in the region of most severe flooding. Several observers recorded total precipitation in excess of 500 mm, with 250 mm falling in only 2 h (NOAA 1995). Radar es-

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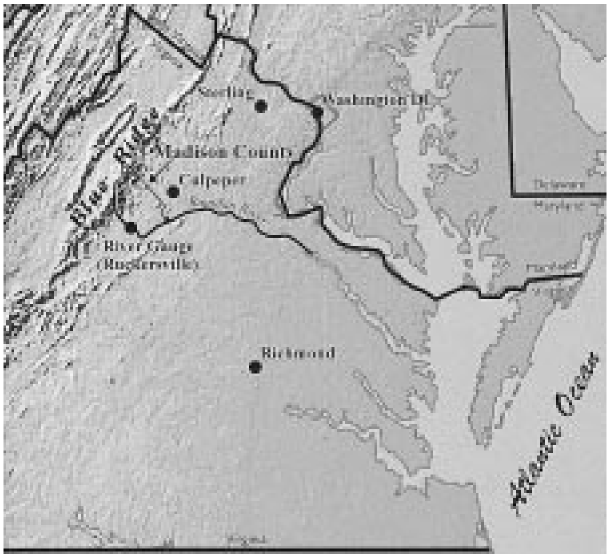


FIG. 1. Map of Virginia showing physiographic features, political boundaries, and selected observing sites.

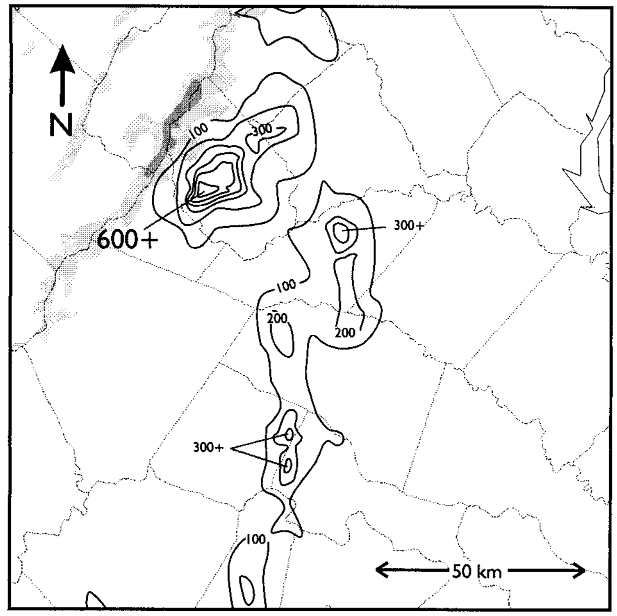


FIG. 2. Estimated storm total precipitation (contour interval 100 mm) ending 2200 UTC 27 Jun 1995. Light and dark shading indicate the approximate areas of terrain heights >750 m and >1000 m, respectively.

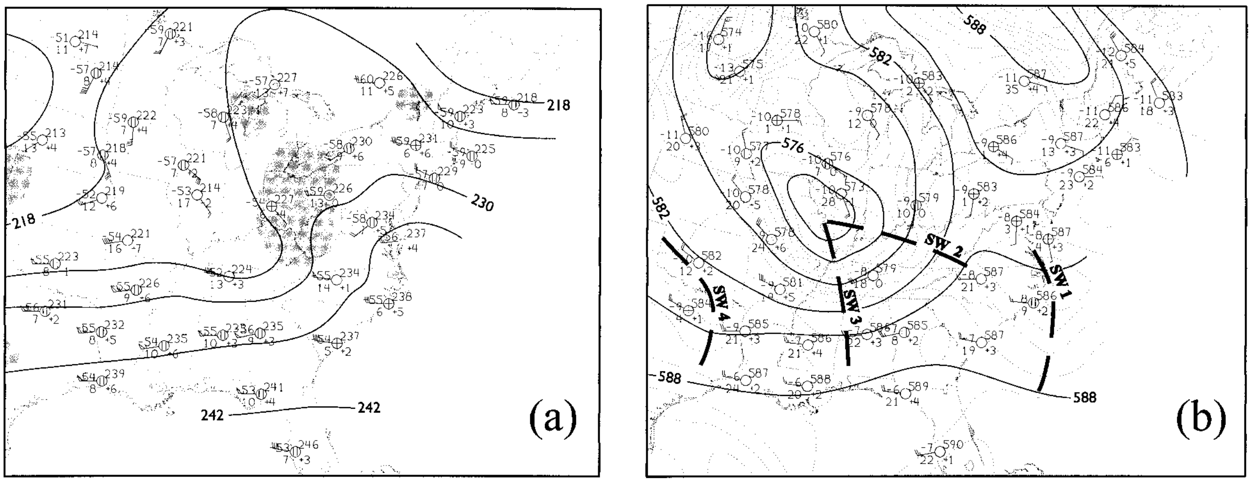
timates indicated that some regions received more than 600 mm (Smith et al. 1996). As a result of the extremely heavy rainfall, the Rapidan River rose over 5 m in 1 h near the town of Culpeper. It is estimated that the return period for a flood of this magnitude is nearly 500 years along portions of the Rapidan River in Madison County (Prugh 1995). Many stream gauges along the Rapidan experienced record discharges, with an incredible discharge of 3000 $m^3\ s^{-1}$ at Ruckersville, Virginia (Smith et al. 1996).

The synoptic environment of the Madison County event was similar to the environment that was conducive to other destructive upslope convective events in the western United States. These include the devastating Rapid City, South Dakota, flood along the eastern slopes of the Black Hills on 9 June 1972, which killed 236 people (Thompson 1972; NOAA 1972; Dennis et al. 1973), and the Big Thompson Canyon flood along the Front Range of the Colorado Rockies on 31 July 1976, which resulted in 139 fatalities (Maddox et al. 1977). More recently, another similar upslope event occurred in Fort Collins, Colorado, in July 1997, claiming at least five lives (Petersen et al. 1999).

This paper strives not only to provide a detailed investigation and documentation of the synoptic and mesoscale conditions associated with the Madison County flash flood, but also to provide a comparison of the characteristics of this event to other significant leeside (with respect to the climatological flow in the mountain layer) upslope convective events. A brief description of the data and methodology is presented in section 2, followed by the meteorological analysis in section 3. A comparison to other events is summarized in section 4 and concluding remarks are presented in section 5.

2. Data and methodology

Surface, upper-air, ship, and buoy reports, supplemented by mesonet data purchased from Automated Weather Source, Inc. (Gaithersburg, MD), were used to construct the surface and upper-air analyses. The mesonet data consisted of hourly observations of surface temperature, dewpoint, and wind speed and direction for six stations in northwestern Virginia. These stations provided critical information about the near environment of the convective system that produced the Madison County flood. All data were analyzed subjectively with the exception of the 500-mb vorticity and the 200-mb divergence fields. These fields were analyzed using the objective analysis system described by Cahir et al. (1981) and smoothed using the Barnes analysis technique (Barnes 1964). All radiosonde data and corresponding indices were analyzed using the RAOB PRO software package (Environmental Research Services, Matamoras, PA). Level II Doppler radar data from the National Weather Service (NWS) office at Sterling, Virginia, were displayed and analyzed using the SNAP Weather Information Integrated Forecast Tool (SWIIFT); (Pearce and Hoffert 1997). Precipitation estimates were based on the storm total precipitation algorithm in SWIIFT. Since Smith et. al. (1996) found that the discharge from the Rapidan River was approximately three times larger than the Weather Surveillance Radar-1988 Doppler (WSR-88D) estimated rainfall accumulation observed by the Sterling radar, and that the mean bias of the radar-estimated rainfall compared to five ground-based raingauges was 2.5, the precipitation



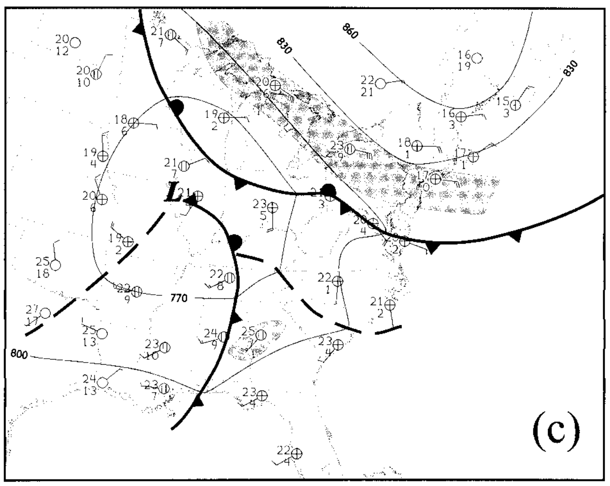


FIG. 3. Synoptic upper-air analyses for 0000 UTC 27 Jun 1995: (a) 200-mb heights (solid lines, 6-dam intervals), divergence $>2 \times 10^{-5}$ s^{-1} shaded; (b) 500-mb heights (solid lines, 3-dam intervals), vorticity (dotted lines, contour interval 1×10^{-4} s^{-1}), and shortwave trough axes (bold dashed lines); and (c) 925-mb heights (solid lines, 3-dam intervals), winds >12.5 m s^{-1} shaded. Fronts and troughs are from surface level analysis. Station plots follow standard convention.

estimation parameters in the SWIIFT algorithm were adjusted until the mean error between the radar and the cooperative observer data was minimized. The resulting estimation of storm total precipitation over Madison County agreed well with the analysis of Smith et al. (1996).

Visible satellite loops were used to estimate cloud motion in the marine boundary layer of the western Atlantic. It was this air that was advected into the Madison system. It was assumed that the shallow cloud motion was representative of the speed and direction of the low-level regional flow. Individual cloud elements that could be tracked through at least three consecutive 15-min satellite images were used. The satellite-derived winds were used to more accurately locate and identify features such as fronts and low-level wind maxima.

3. Meteorological analysis

a. Prestorm environment

The synoptic pattern at 0000 UTC 27 June was dominated by a large, negatively tilted upper-level longwave trough over the Upper Mississippi River Valley and a ridge axis over New England (Fig. 3). The flow around the base of the trough split as it approached the Appalachians, creating pronounced divergence over the Ohio Valley. Similar large-scale divergence/difluent patterns have been associated with excessive rainfall in the mid-Atlantic region (Giordano and Fritsch 1991). Several weak shortwave troughs (labeled 1–4 in Fig. 3b) were propagating through the longwave trough. The third wave, which was then moving through the Tennessee Valley, propagated into the mid-Atlantic region on the day of the Madison flood.

At the surface, a high-pressure system was positioned over New England and was funneling cool air southward into the circulation of a broad low-pressure system, centered almost directly under the upper-level trough (cf. Figs. 3a,b and 4). The southward progress of the cool air was marked by a weak frontal zone stretching from the western Atlantic westward into the southern Great Lakes.

A pronounced easterly flow dominated the region between the synoptic high- and low-pressure systems. This

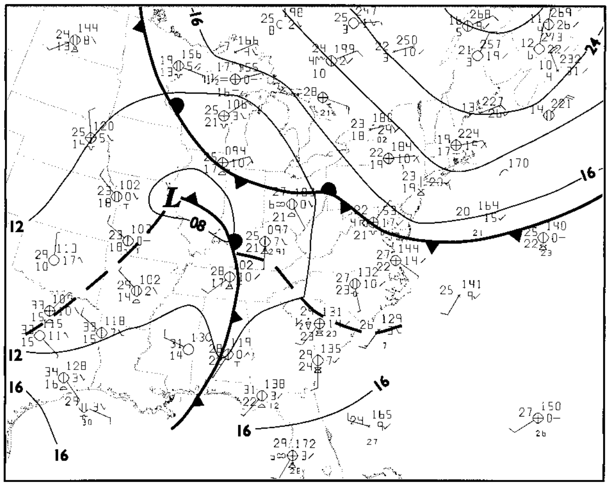


FIG. 4. Synoptic surface analysis for 0000 UTC 27 Jun 1995. Contour interval is 4 mb.

easterly flow was transporting cool, moist Atlantic air westward into the east slopes of the Appalachians. Soundings taken at Sterling and Wallops Island, Virginia (Figs. 5a,b), just ahead of the frontal zone, show a deep, moist southeasterly flow with a weak stable layer between 50 and 100 mb above the surface. The total totals index (Peppler and Lamb 1989) ranged between 44 and 47, indicating favorable conditions for the development of thunderstorms if the inversion were broken. Lifted index values (Peppler and Lamb 1989) for surface air were only slightly negative (≈ -1) and convective available potential energy (CAPE) was relatively small, only about 150 J kg^{-1} across the region. Wind speeds were generally light, particularly at midlevels, and weak vertical shear prevailed over most of the troposphere. The air was very moist over the mid-Atlantic region, especially in northern Virginia where the entire troposphere was nearly saturated (Fig. 5a) and precipitable water values (Fig. 6) were nearly 50 mm (over 160% of normal). These values exceeded the maximum precipitable water values observed just prior to the Johnstown, Pennsylvania, flash flood of 1977 (Hoxit et al. 1978). Moreover, the surface and 850-mb dewpoints were the same as the average values found for intense rainstorms in the mid-Atlantic region, that is, 23° and 14°C, respectively (Giordano and Fritsch 1991). The light winds, coupled with the nearly saturated conditionally unstable environment, favored the development of slow-moving convective systems capable of producing prolonged heavy rainfall.

Farther north, in the postfrontal region, the atmosphere was slightly more stable and exhibited a midlevel dry layer (e.g., see the Brookhaven, NY, sounding; Fig. 5c). Nevertheless, precipitable water values were still over 40 mm, which is well above normal (Fig. 6). The low-level layer of easterlies was stronger and deeper at Brookhaven than at Sterling and Wallops Island, reflecting the stronger postfrontal pressure gradient (Fig. 4) and greater potential for orographic lift in the postfrontal zone. Analysis of the 925-mb winds (Fig. 3c) revealed a postfrontal area of relatively fast-moving low-level easterlies with an overwater fetch.

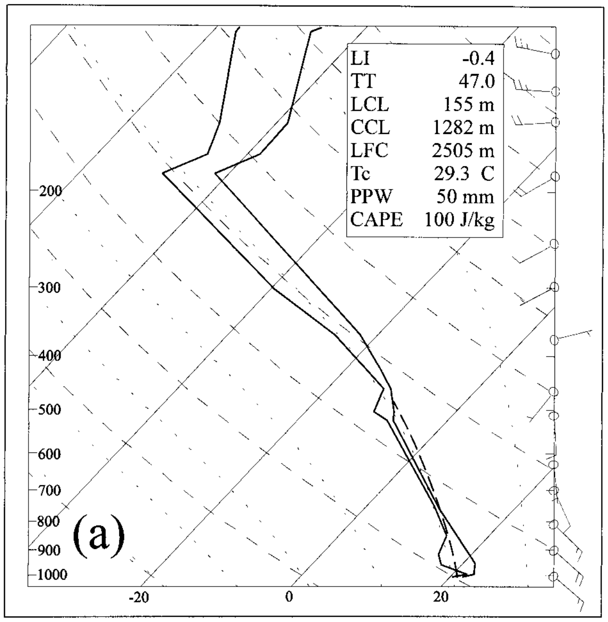
Localized flooding had already occurred from Virginia northward through Pennsylvania and New Jersey during the afternoon of 26 June as slow-moving thunderstorms passed through the region. A rash of severe weather also accompanied the passage of the cold front in Pennsylvania and New Jersey, with numerous reports of wind damage and several confirmed F0 tornadoes. Showers and thunderstorms continued immediately behind the front in the low-level moist ribbon flowing westward from the Atlantic Ocean. South of the front, locally heavy showers and thunderstorms formed during the overnight hours in the tropical-like air mass over Virginia, West Virginia, and Maryland. These storms exhibited a slow northward drift, allowing some areas to experience heavy rainfall on already saturated ground. For example, Cumberland, Maryland, received 63 mm of rain in the 1-h period ending at 0100 UTC 27 June.

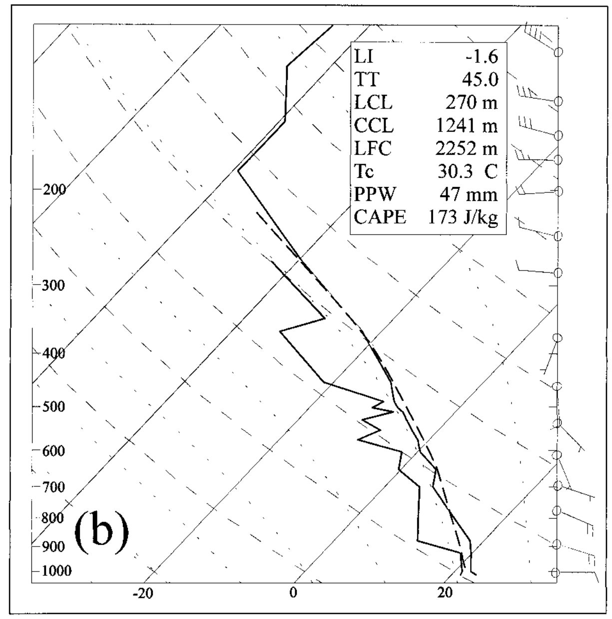
The high-pressure system over New England built southward, pushing the cold front into northeastern Virginia by 0600 UTC 27 June (Fig. 7a). Sterling, Virginia, radar indicated that a broken line of showers attended the front and that a convective system (hereafter termed the Piedmont storm) initiated around 0700 UTC in the prefrontal air in Madison and Orange Counties, just east of the Blue Ridge (Fig. 8). As evident from Figs. 1 and 2, the Blue Ridge is the first significant orographic feature west of the Piedmont plains and Madison County is near the midpoint of one of the tallest unbroken stretches of the ridge. Because of Madison County's geographic position with respect to the Blue Ridge, it is situated in an area that should receive the maximum orographic forcing given an easterly wind component. Flooding was already occurring in Madison County by 0815 UTC, prompting the NWS forecasters at Sterling, Virginia, to issue the day's first flash flood warning for Madison County at 0833 UTC.

The front advanced steadily to the south and west, reaching extreme eastern Madison County around 1000 UTC (not shown). Automated rain gauges and radar estimates indicated that over 75 mm of rain had already fallen across the ridges by 1020 UTC. As the front overtook the Piedmont storm, the convection began moving southward with and along the frontal zone, leaving the county virtually rain free at 1100 UTC (Fig. 8d). This would be a brief respite, however, as a new convective system was forming just northeast of Madison County and was propagating southwestward into the county. This second system, hereafter referred to as the Madison storm, became the main flash-flood-producing convective system.

b. Storm environment

The position of the large upper-level longwave trough changed very little overnight (0000–1200 UTC 27 June;





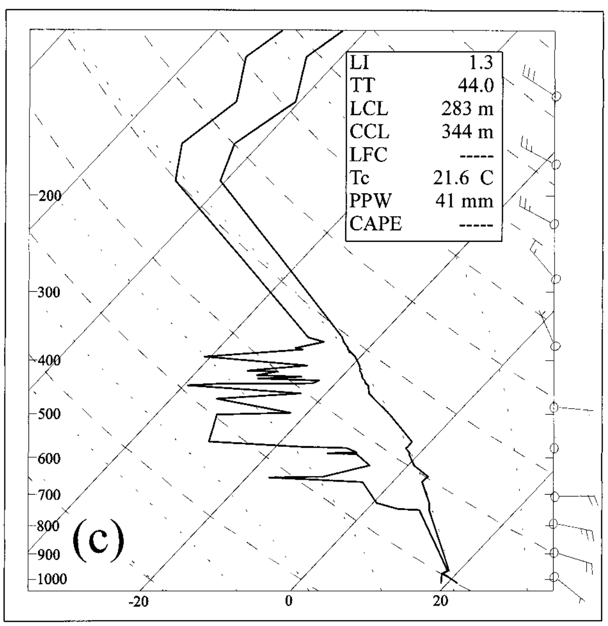


FIG. 5. Soundings for 0000 UTC 27 Jun 1995: (a) Sterling, VA; (b) Wallops Island, VA; and (c) Brookhaven, NY. Listed in each insert are lifted index (LI), total totals (TT), lifting condensation level (LCL), convective condensation level (CCL), level of free convection (LFC), convective temperature (Tc), precipitable water (PPW), and CAPE. Bold dashed lines indicate the thermodynamic path of a parcel lifted from the surface.

cf. Figs. 3 and 9). However, shortwave troughs continued to rotate cyclonically around the main upper-level system, with the axis of the third shortwave trough (SW-3) entering southwestern Virginia. The zone of maximum difluence shifted directly over western Virginia (Fig. 9a) while the flow in the mid- and upper levels remained weak, maintaining the threat for slow-moving thunderstorms.

The New England surface high-pressure system continued to press southward, spreading cooler and drier air into New Jersey and Pennsylvania (Fig. 10). As a result, stability increased (e.g., the lifted index at Brookhaven increased to +6.9; see Fig. 11c), which contributed to a tendency for cold air damming (Forbes et al. 1987; Fritsch et al. 1992) and effectively prohibited any surface-based convective development from central New Jersey northward. Closer to the front, in northern Virginia, surface temperature and dewpoint values remained in the lower 20°Cs, similar to conditions south of the front. The 1200 UTC radiosondes from both Sterling and Wallops Island (Figs. 11a,b) were representative of the postfrontal environment and indicated that very little lifting was needed to initiate convection. For example, at Sterling, the lifting condensation level and

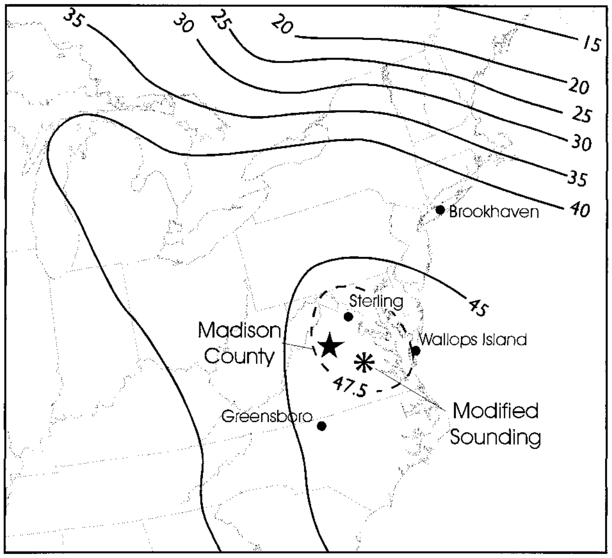


FIG. 6. Precipitable water (mm) for 0000 UTC 27 Jun 1995. Locations of soundings used in the thermodynamic analyses are shown by bold dots. Asterisk indicates the location of the modified 1800 UTC sounding (Fig. 11d later).

the level of free convection for a surface parcel were only 155 and 337 m above ground level, respectively. Moreover, with a convective temperature of 24°C and surface temperatures already in the low 20°Cs, little heating would be required to initiate deep convection. From 0000 to 1200 UTC, CAPE increased to about 600 J kg^{-1}, the lifted index lowered slightly to -2, and the precipitable water remained near 50 mm.

The soundings also indicate that the zone of relatively fast-moving low-level easterlies evident in Fig. 3c were now firmly entrenched in northern Virginia. Satellite-observed cloud motion vectors derived from the movement of low-level cumulus clouds reveal that the swath of high speed easterlies originated well out over the Atlantic Ocean and was advecting maritime air into the northern region (Fig. 12). Doppler radar velocity azimuth display (VAD) wind profiles from Sterling (Fig. 13) indicate that the depth and strength of the postfrontal high speed easterlies were increasing over northern Virginia as the high-pressure system built southward. Moreover, the satellite and radar loops indicated that the easterly flow was converging with a pronounced southeasterly flow over central Virginia. The combination of high precipitable water, a deep layer of small CAPE, converging low-level inflow, and weak midlevel steering currents was conducive to nonsevere, slow-moving thunderstorms with heavy rain.

To explore further the convective potential, the Sterling sounding was introduced into the one-dimensional cloud model developed by Anthes (1977). Without any modification, the sounding did not generate a deep convective cloud. Analysis of the various physical terms in the model indicate that deep cloud growth was prevented by the weak convective inhibition and by large water loading [cloud droplets and rainwater; see Anthes (1977)] associated with the moist air mass. However, a cloud over 10 km deep was able to form in the model if the surface parcel was provided with an initial vertical velocity of 2.0 m s^{-1} or if the surface temperature was warmed by 2°C. These results suggest that, despite the small convective inhibition, deep clouds were possible only if either the surface air was heated or if some type of feature such as a front, terrain, or an outflow boundary forced the air upward. The necessity for some type of forcing mechanism for *deep* convection to develop agreed with radar observations (Fig. 14), which showed numerous small echoes (typically about 2–4 km deep) in the postfrontal air over eastern Virginia but no deep moist overturning.

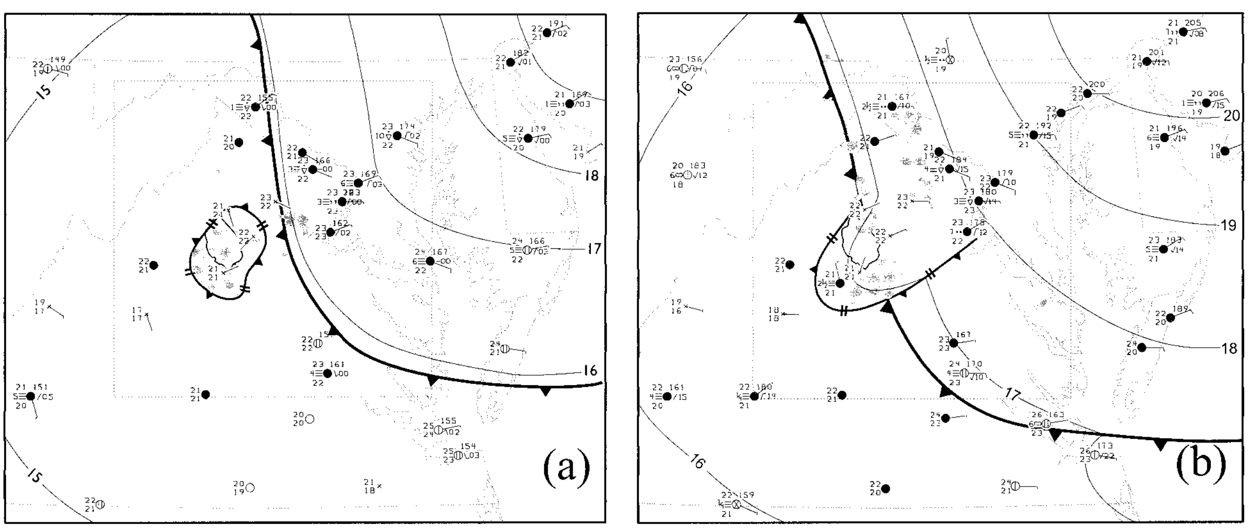


FIG. 7. Mesoscale analyses for 27 Jun 1995: (a) 0600 and (b) 1200 UTC. Convention for mesoscale boundaries is that used by Young and Fritsch (1989). Radar echoes >30 dBZ shaded. Dotted square depicts the SWIIFT analysis domain for the Sterling, VA, WSR-88D data.

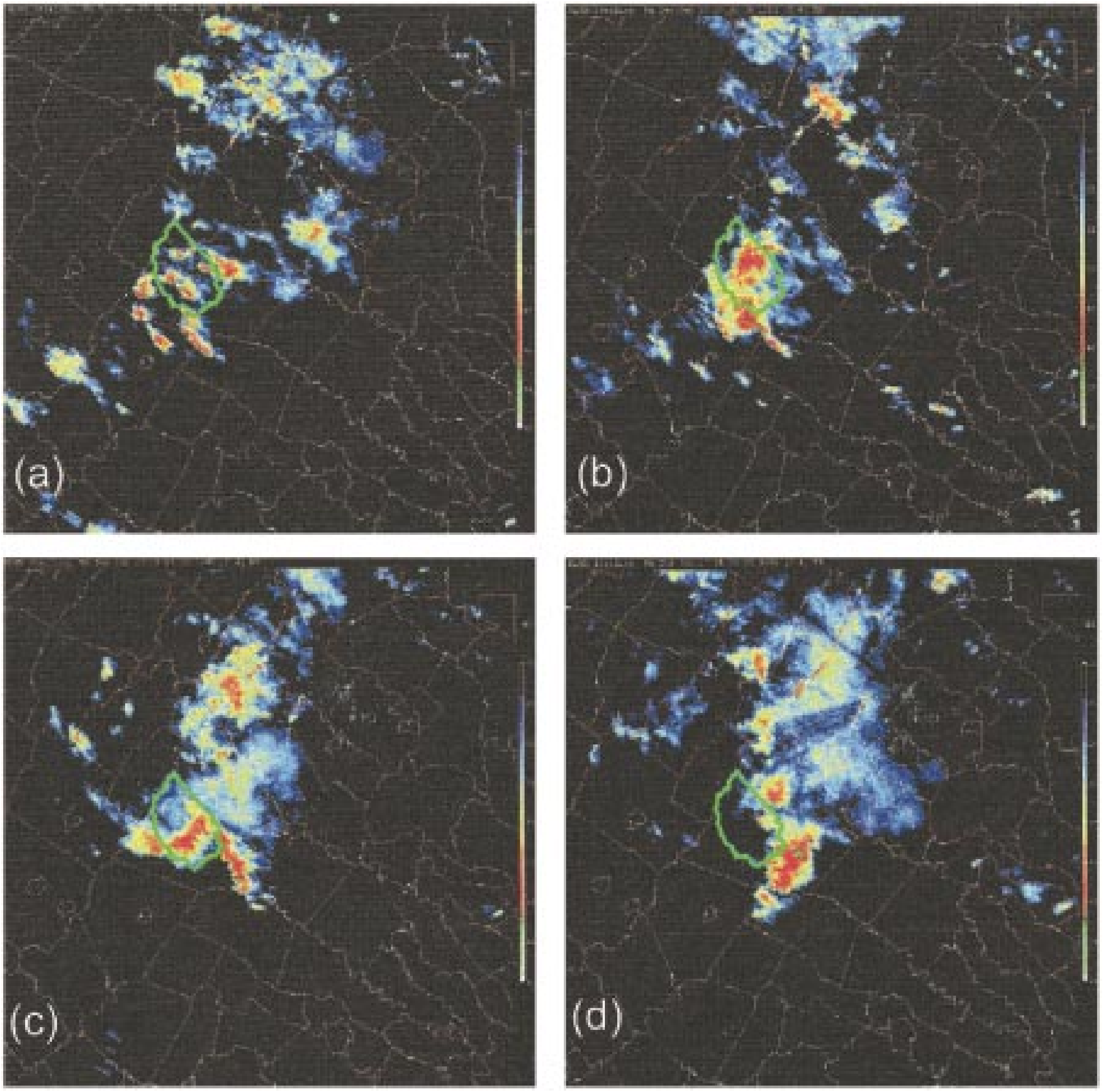
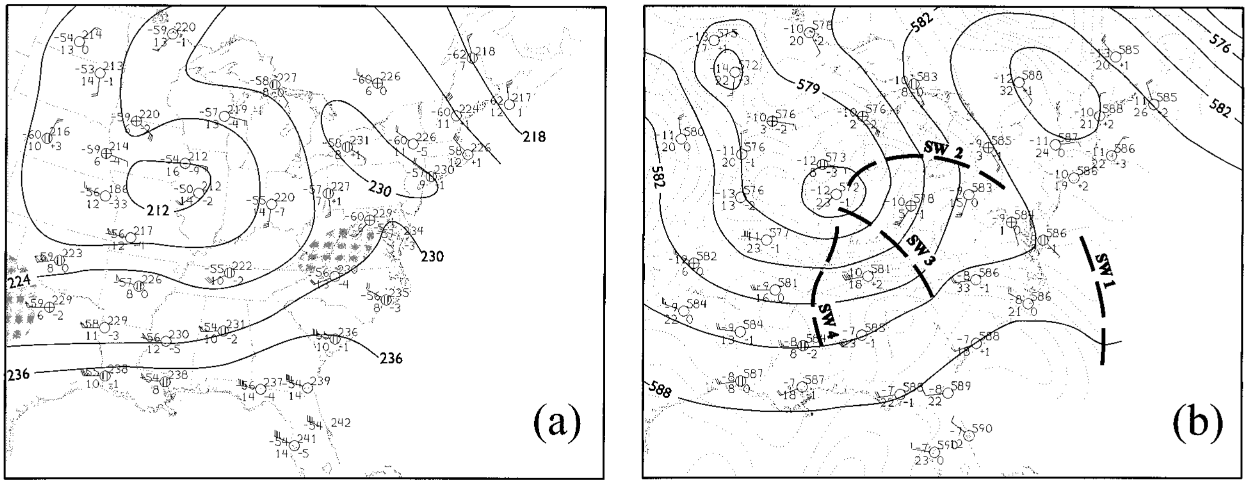


FIG. 8. Radar reflectivity images from Sterling, VA, WSR-88D on 27 Jun 1995: (a) 0600, (b) 0700, (c) 0900, and (d) 1100 UTC. Images are taken from lowest elevation radar scan (0.5°) and the Madison County border is depicted in green.

As the morning progressed, shallow echoes continued to develop in the postfrontal air over eastern Virginia. Some of these echoes briefly reached 45 dB*Z* but quickly dissipated (Fig. 14). The Piedmont system continued to move southeastward with the front (see Figs. 7, 14, 16, and 17). Although Sterling's radar indicated that echo tops prior to 1200 UTC (not shown) ranged between 6 and 10 km, tops in the Piedmont system increased to 12–15 km with reflectivities of 30 dB*Z* reaching 12 km above the surface by 1800 UTC (Fig. 15a). Meanwhile, the Madison system continued to intensify as it moved very slowly southwestward across Madison County (Figs. 14a,b). Echo tops were lower in the Madison system than in the Piedmont system, only reaching between 10 and 12 km above ground level with 30-dB*Z* echoes between 7 and 10 km above ground level (Fig. 15b), suggesting that the two systems were feeding on air with different thermodynamic properties. Closer examination of the storm region (Fig. 16) supports this notion. Large patches of low-level stratocumulus and a broad shield of high-level cirrus outflow from the Madison and Piedmont systems covered much of northern Virginia. Light rain was being reported at several stations beneath the anvil outflows. As a result, surface temperatures over northern Virginia rose very little from overnight values with readings still in the low 20°Cs by



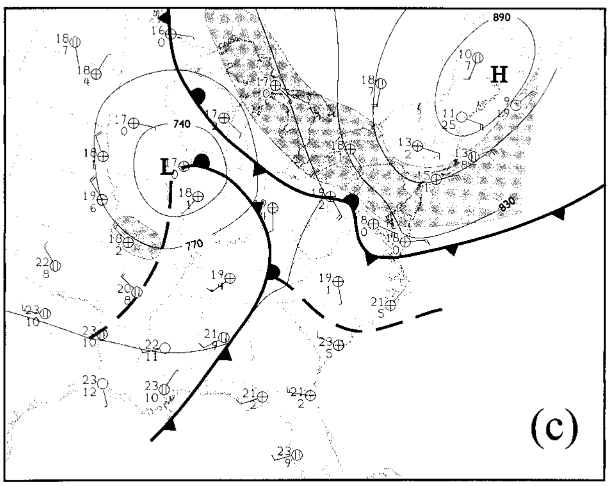


FIG. 9. Synoptic upper-air analyses for 1200 UTC 27 Jun 1995. Conventions as in Fig. 3

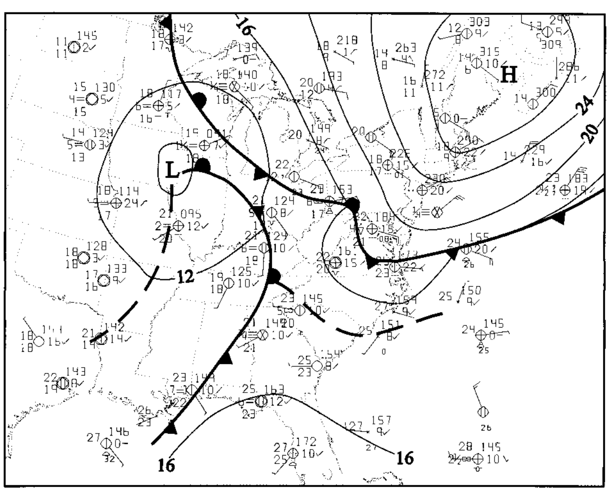


FIG. 10. Synoptic surface analysis for 1200 UTC 27 Jun 1995. Contour interval is 4 mb.

late morning. Surface data (Figs. 17a,b) indicated that this cool damp air was flowing southwestward across Madison County. Farther south and east, however, cloud cover was scattered to broken and appeared predominantly in the form of rolls of low-level cumulus (Fig. 16). Temperatures over southeastern Virginia had climbed quickly into the mid 20°Cs by late morning and into the upper 20°Cs/low 30°Cs by afternoon (Figs. 17a,b). The Piedmont system was better positioned to intercept this sun-warmed, higher-θ_e air over southeastern Virginia.

An estimate of the inflow environment of the convective systems at midday (1800 UTC) was determined by averaging the 1200 UTC 27 June and 0000 UTC 28 June soundings taken at three sites (Sterling, Virginia; Wallop's Island, Virginia; Greensboro, North Carolina) surrounding the storm location (see Fig. 6). These soundings, especially Sterling, did not reflect well the solar warming of the boundary layer that was occurring over eastern and southern Virginia. Therefore, observed surface temperatures and dewpoints across east-central Virginia were used to construct a surface boundary layer

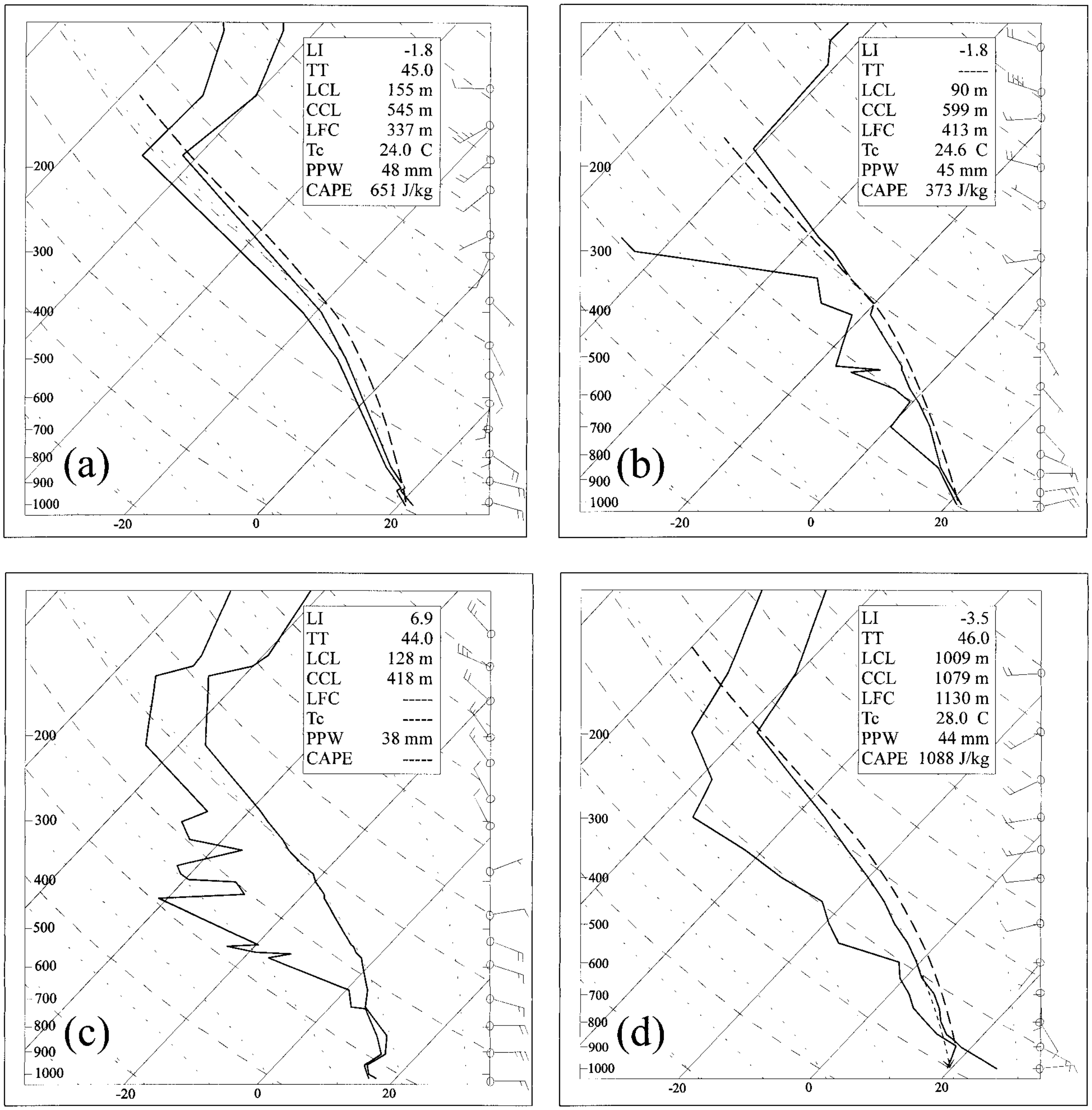


FIG. 11. Soundings for 1200 UTC 27 Jun 1995: (a) Sterling VA; (b) Wallops Island, VA; (c) Brookhaven, NY; and (d) estimated sounding for conditions over east-central Virginia at 1800 UTC. Insert and parcel paths as in Fig. 5. Moist downdraft path [short-dashed line in (d)] determined from the procedure of Foster (1958).

profile more representative of the inflow to the Piedmont system. Specifically, it was assumed that the lapse rate was dry adiabatic and that the mixing ratio was constant from the surface to the intersection of the dry-adiabatic lapse rate with the mean sounding (Fig. 11d). The modification of the surface boundary layer to reflect the solar warming increased cloud base from 150 m above ground to 1000 m, and CAPE nearly doubled from about 600 J kg^{-1} to over 1000 J kg^{-1}. Therefore, it would be expected that the Piedmont system would generally have higher cloud tops than the Madison system, especially later in the day. For both storms, however, low cloud bases and high freezing levels combined to produce a deep ($\approx$4 km) layer in which warm cloud coalescence processes could occur. Braham et al. (1957) noted that warm maritime cumulus clouds are colloidally unstable and can produce intense rainfall even when they are only a few kilometers thick. The heavy rains falling from the scattered shallow echoes over the Piedmont suggested the presence of a warm rain process. In the

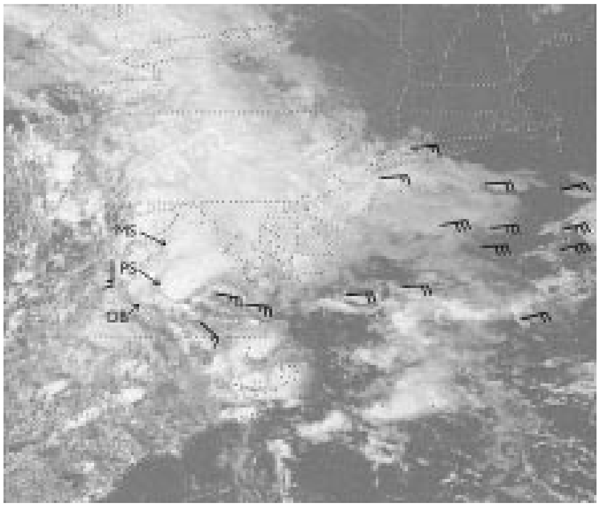


FIG. 12. Visible satellite image for 1716 UTC. Cloud-tracked winds (full barb = 5 m s^{-1}) are shown for selected low-level cloud elements; MS, PS, and OB indicate the location of the Madison storm, Piedmont storm, and outflow boundary, respectively.

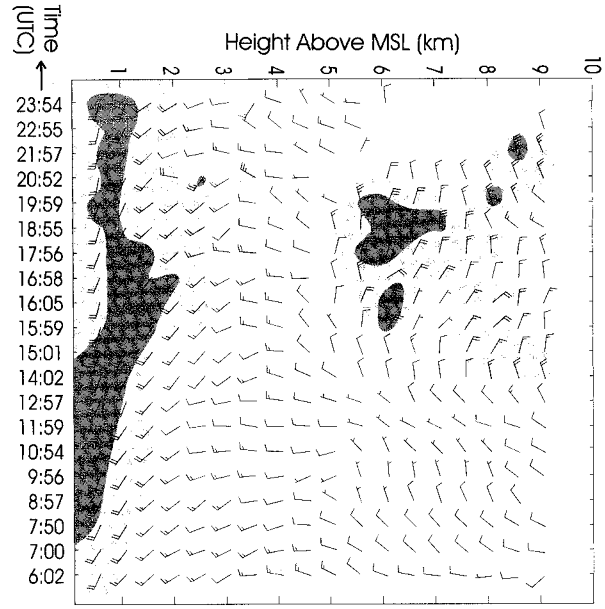


FIG. 13. VAD wind profiles from Sterling, VA, WSR-88D for 0600–2300 UTC 27 Jun 1995. Full barb = 5 m s^{-1}, light shading = 7.5 to 12.5 m s^{-1}, dark shading > 12.5 m s^{-1}.

Madison system, radar revealed a low-echo centroid with the great majority of the high reflectivity concentrated below the freezing level. As shown by Smith et al. (1996), rain rates in Madison County commonly exceeded 100 mm h^{-1} and at times briefly exceeded 300 mm h^{-1}, suggesting that the low-based, low-echo-centroid system must have developed an efficient warm cloud process.

c. *Storm propagation*

The differences in boundary layer temperature and cloud-base height between the environments of the two systems may also help to explain the difference in the systems' movements. In particular, the Madison system moved very slowly ($\approx$1.2 m s^{-1}) southwestward along the Blue Ridge while the Piedmont system advanced considerably faster ($\approx$3.3 m s^{-1}) but to the south-southeast. Both systems moved in directions with components opposite to the mean cloud-layer flow (Figs. 8, 11, 13, and 14). The fact that the systems moved against the mean cloud layer flow indicates that propagation rather than advection dominated their movement, a point illustrated by Smith et al. (1996). They demonstrated that individual cells initiated along the southern flanks of the convective systems, intensified, drifted northward between 2 and 5 m s^{-1}, and eventually dissipated. The process whereby the successive cells formed slightly farther upstream is essentially the same as the upstream propagation or ''backbuilding'' described by Chappell (1986) and Shi and Scofield (1987).

It is well known that system propagation is intimately linked to moist downdrafts (e.g., Rotunno et al. 1988) and therefore differences in the moist downdraft potential of the two systems, or differences in the subcloud-layer environment, could contribute to significant differences in propagation. The primary cooling mechanisms for producing the negative buoyancy that drives moist downdrafts are evaporation and melting. Large liquid water contents can also generate downdrafts through precipitation drag (Srivastava 1987). It is evident from the small dewpoint depressions shown in Figs. 7, 11a,b, and 17a,b that there was very little potential for evaporatively driven, moist downdraft cooling in western Madison County. The maritime air flowing westward from the Atlantic toward Madison County had dewpoint depressions of only a few degrees. Although it is possible that the Madison system's inflow experienced some solar heating, rainfall from the high-level outflow of the Piedmont system further cooled and moistened the inflow to the Madison system, thereby lowering the cloud bases and decreasing the potential for subcloud-layer evaporative cooling in the Madison system. Cloud bases were within about 100 m of the ground and, according to the sounding at Sterling, midlevels were nearly saturated. Moreover, it was unlikely that melting-induced cooling could have contributed greatly to production of downdrafts as it did in events investigated by Szeto et al. (1988a,b) and Marwitz and Toth (1993). Reflectivity cross sections showed that the Madison system had a low-echo centroid, similar to that observed with the Big Thompson system (Caracena et al. 1979) and the Cheyenne, Wyoming, flood (Chappell and Rodgers 1988), with most of the condensate forming and falling from levels below the freezing level. On the other hand, with the extremely large rain rates that occurred with the Madison system, it *is* likely that pre-

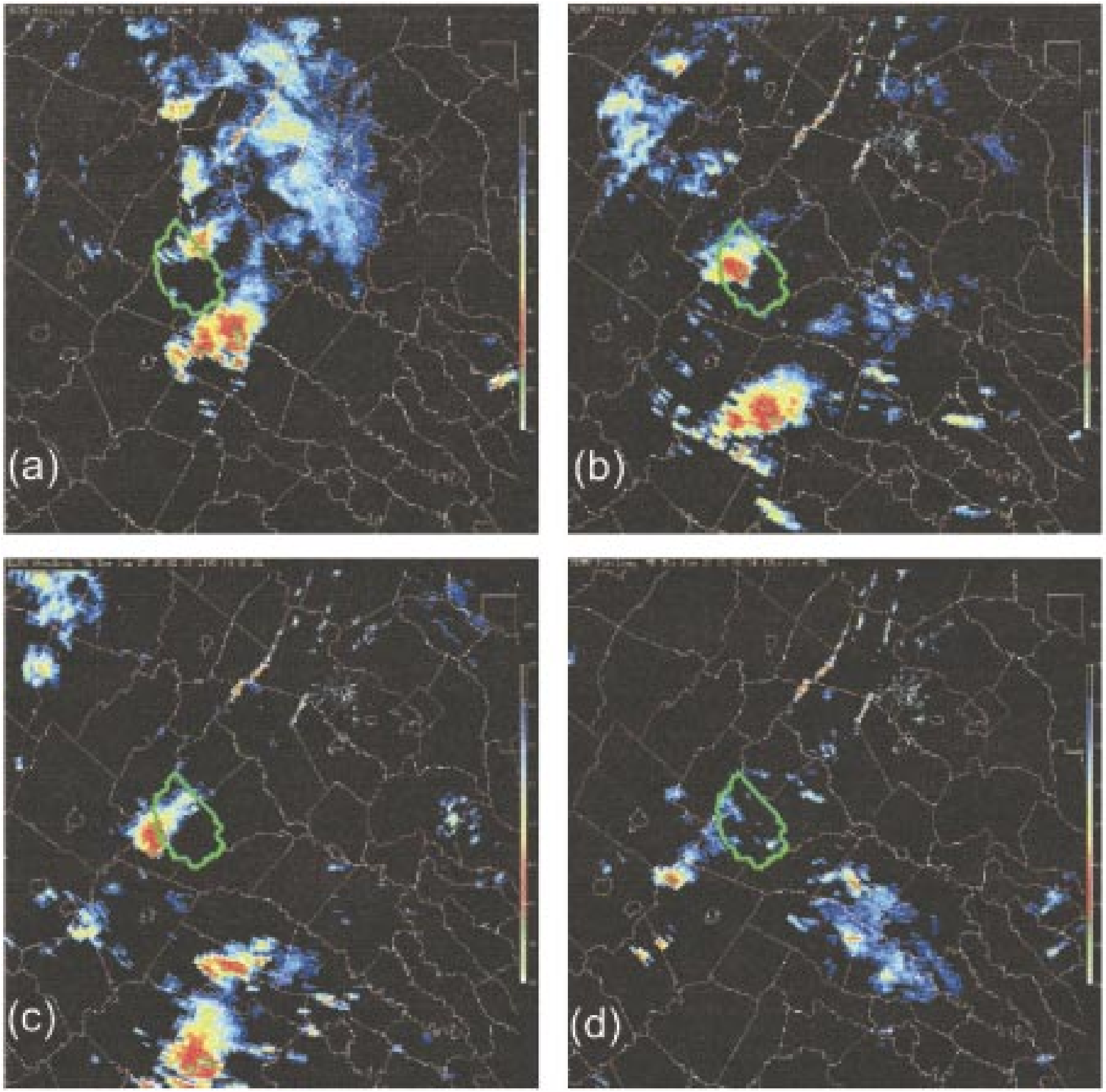


FIG. 14. Radar reflectivity images from Sterling, VA, WSR-88D on 27 Jun 1995: (a) 1200, (b) 1500, (c) 1800, and (d) 2100 UTC. Images taken from lowest elevation radar scan (0.5°).

cipitation drag played a significant role in generating moist downdrafts. Unfortunately, the beam centerline of the lowest radar scan (0.5°) was over 1.5 km above the surface in the region of interest, too high to determine conclusively if significant downdraft outflows were present and if they advanced southeastward against the incoming low-level flow to initiate new cells.

Surface observations (Fig. 7) suggest the presence of cool outflow in the vicinity of both systems. However, the higher cloud base, greater dewpoint depression, and higher temperature (therefore density) difference between the downdraft air and the inflowing environmental air for the Piedmont system (which had experienced more solar heating) offered greater potential for moist downdraft cooling, outflow, and lifting of the ambient inflowing air than for the Madison system. This is readily evident from the moist downdraft potential energy (DCAPE), which for the Madison storm environment was under 90 J kg^{-1}. Based upon the modified sounding, the DCAPE for the inflow environment of the Piedmont system was 250 J kg^{-1} (Fig. 11d). Visible satellite imagery supports the presence of outflow from the Piedmont system (Fig. 16) but, unfortunately, the lower troposphere around the Madison system was obscured by high-level clouds and no clear indication of a comparable outflow is available. Nevertheless, based upon the

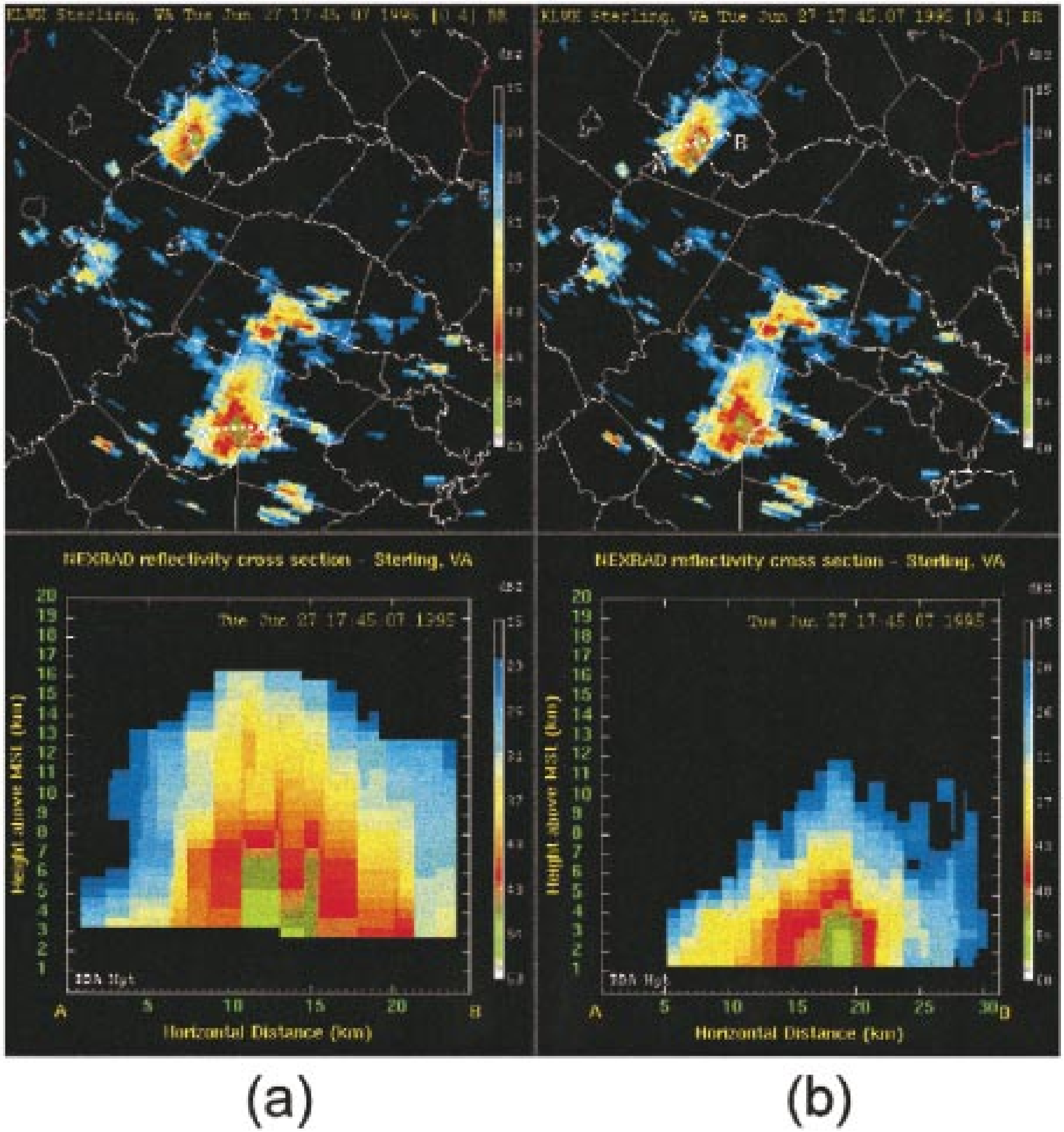


FIG. 15. Radar cross sections from Sterling, VA, WSR-88D at 1745 UTC 27 Jun 1995: (a) Piedmont storm and (b) Madison storm. Locations of cross sections (white dashed lines) are also shown.

likelihood of stronger moist downdrafts and similar environmental winds, the Piedmont system would tend to propagate faster than the Madison system, as was observed.

From a forecasting perspective, the difference in propagation rates is of great importance. Both systems produced excessive rainfall. However, the Madison system produced maximum amounts that were two to three times that from the Piedmont system. Much of this difference stems directly from the difference in propagation speed. Assuming similar rain rates for the two systems, the difference in propagation speed alone would explain most of the heavier rainfall amounts in Madison County. This agrees with Chappell (1986) and Doswell et al. (1996) who noted that slow system movement typically dominates flash-flood-producing heavy precipitation.

In order to further explore the different movements of the two systems, an empirical model (Corfidi et al. 1996) that predicts the movement of the heavy rain areas

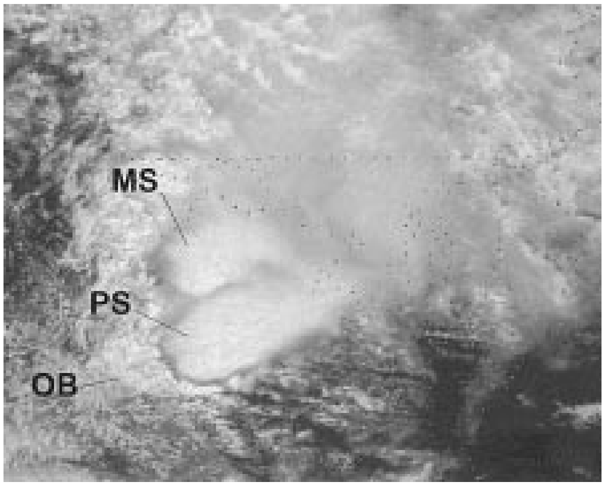


FIG. 16. Visible satellite image at 1630 UTC 27 Jun 1995; MS, PS, and OB indicate the location of the Madison storm, Piedmont storm, and outflow boundary, respectively.

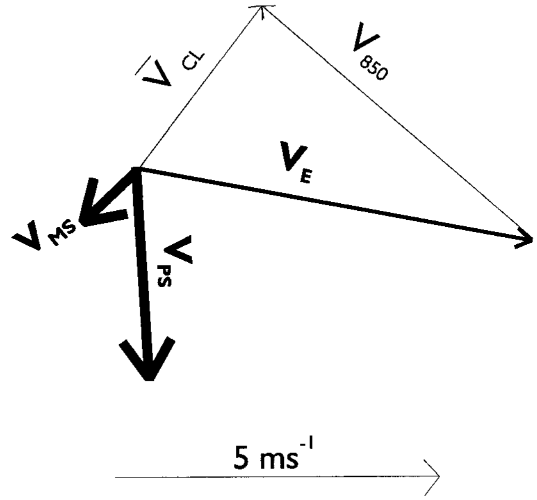


FIG. 18. Expected movement ($\mathbf{V}_E$) of a mesoscale convective system forming within the environment of the Madison and Piedmont storms. Expected movement determined following the method of Corfidi et al. (1996); V_{CL}, V_{850}, $\mathbf{V}_E$, V_{PS}, and V_{MS} represent the mean cloud layer wind, 850-mb wind, expected movement, observed movement of the Piedmont storm, and the observed movement of the Madison storm, respectively.

of mesoscale convective systems was applied to the environment of the Madison and Piedmont systems. This model relies strongly on the production of moist downdrafts to force the new convective cells that drive the propagation. Thus, the movement of the Piedmont system (with its higher cloud bases and greater moist downdraft potential) should align more closely with the movement predicted by the empirical model than the Madison system (with cloud bases virtually at the surface). Figure 18 shows that the movements of both the Madison and the Piedmont systems departed significantly from the expected movement ($\mathbf{V}_E$).

There are several possible explanations for the Piedmont system's departure from $\mathbf{V}_E$. First, the interaction of the cold front with the outflow boundary provided a preferential location for new cell growth. This intersection moved slowly southward with the front (Figs. 7 and 17) promoting a more southerly propagation for the Piedmont system. Second, the system existed in the transitional region between easterly winds and southeasterly winds. It is possible that the system was tapping into the more southerly winds just above the front. According to the Corfidi et al. model, this would result in a more southerly propagation.

While the Piedmont system's movement departed significantly from $\mathbf{V}_E$, it was closer to the expected movement than was the Madison system. The large difference between the movement of the Madison system and both

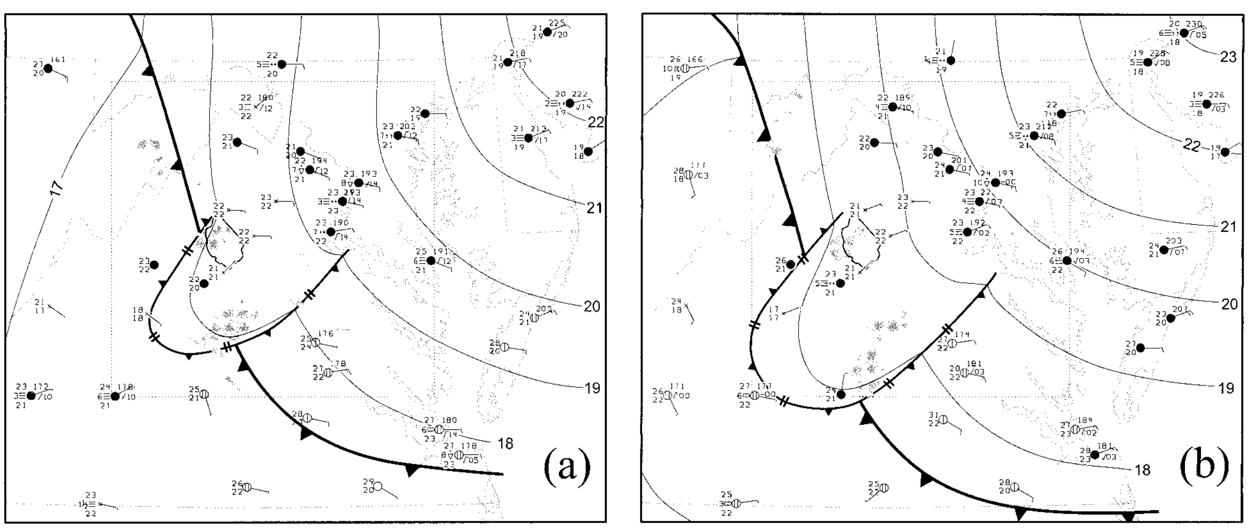


FIG. 17. Mesoscale surface analyses for 27 Jun 1995: (a) 1500 and (b) 1800 UTC. Convention as in Fig. 7.

$\mathbf{V}_E$ and the Piedmont system suggests that the physical processes controlling the movement of the Madison system were different from that of most mesoscale convective systems. This agrees with the hypothesis that the Madison system was predominantly terrain forced and, because of its low cloud bases and cooler inflow environment, its moist downdrafts were unable to force new cell growth on its eastern flank (i.e., it could not propagate away from the mountain forcing).

d. Poststorm environment

By 1800 UTC, the Madison system had reached the southwestern border of Madison County and, finally, around 2000 UTC, the areal extent of the precipitation decreased significantly and the system dissipated. New cells continued to develop farther east along the trailing outflow boundary and baroclinic zone created by the anvil shadow and precipitation from the Piedmont system; however, by 2200 UTC, this activity ceased as well.

The termination of the flood-producing convective system corresponded with the breakdown of the special confluence of conditions that favored deep slow-moving thunderstorms in central and northern Virginia. Probably the most important change was the increase in stability over northern Virginia as the New England high-pressure system pressed southward. The increase in stability is readily evident from the 0000 UTC 28 June sounding taken at Sterling (Fig. 19a). Even though the lower troposphere was still very moist, CAPE had decreased to only about 30 J kg^{-1} and the lifted index was now positive. Along with the change to a less favorable thermodynamic environment, the pronounced 200-mb divergence had shifted southward into southern Virginia and the Carolinas (Fig. 20a) and the 500-mb positive vorticity advection had weakened as shortwave trough 3 moved into Pennsylvania, signaling a switch from regional forcing to suppression (Fig. 20b). The axis of the shortwave trough passed over Madison County about 2100 UTC, which roughly corresponds to the time that the systems dissipated.

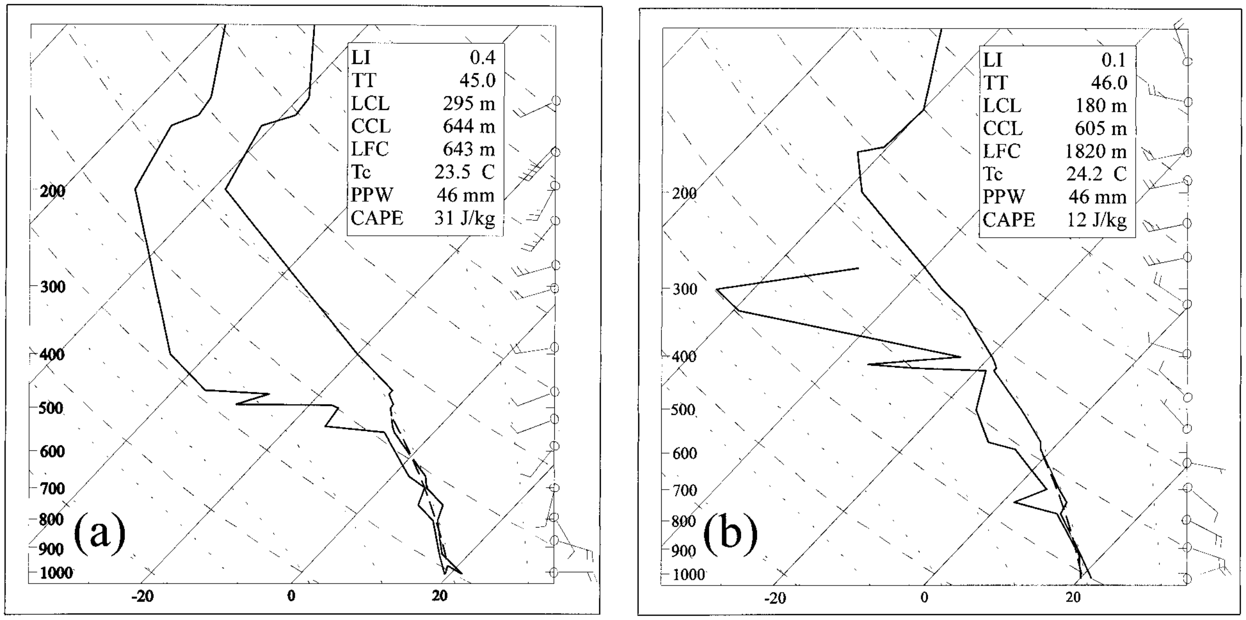
Shortly after the time that the Madison storm ended, the surface front had progressed to just south of the North Carolina–Virginia border and then arced northwestward across the Blue Ridge in southern Virginia (Fig. 21). As indicated by the Greensboro, North Carolina, sounding, air along and just south of the front was much less stable than the air over northern Virginia (cf. Figs. 19a,b). The lifted index and CAPE at Greensboro were -4.1 and 854 J kg^{-1}, respectively. Heavy showers began in the vicinity of the front that afternoon and evening as shortwave trough 4 approached from the southwest. Strong frontal convergence (not shown) coupled with the orographic lift once again produced excessive rainfall (>200 mm) by the morning of 28 June, but this time the favored zone with conditions analogous to those that resulted in the Piedmont and Madison systems was centered over the mountains and western Piedmont of southern Virginia. Remarkably, the synoptic and mesoscale patterns remained quasi-stationary over southern Virginia into the morning of 29 June, resulting yet again in similar excessive rainfall amounts over southern Virginia, northern North Carolina, and extreme southeastern West Virginia.

4. Comparison to other events

The environment, structure, and evolution of the Madison system are similar to other upslope flooding events. For example, Fig. 22 shows the 500-mb height fields prior to the Rapid City, Big Thompson, and Fort Collins floods mentioned in section 1. In all cases, the midlevels exhibited a synoptic-scale negatively tilted trough–ridge pattern that straddled a north–south mountain range. The systems initiated between the ridge and trough in an environment of weak southerly steering currents and divergence aloft. A weak shortwave trough was approaching the threat area from the southwest, contributing to the destabilization of the environment and transporting midlevel moisture into the threat area.

At the surface, the most prominent feature in all cases was a large high-pressure system, an extension of the large negatively tilted ridge aloft (Fig. 23). In each case, the surface high was building slowly southward along the eastern slopes of the mountain range. A slow-moving cold front marked the leading edge of the advancing high pressure system. A postfrontal band of strong, conditionally unstable, moisture-laden easterly winds was oriented nearly perpendicular to the mountains. The zone where these winds intercepted the mountains marked the threat area. In all cases the postfrontal conditionally unstable moist ribbon fueled the systems. Figure 24 presents soundings representative of the postfrontal environment in which the systems thrived. Note the similar characteristics of each sounding. Pronounced low-level easterly winds with a high moisture content were present in the lowest 1–2 km. This conditionally unstable air was capped by a small temperature inversion. Light southerly winds extending upward through 500 mb prevailed above the low-level easterly wind maximum. Nearly saturated, tropical-like air with a high freezing level existed above the temperature inversion in all events. Figure 25 presents a schematic of the major features common to each of the flood events.

Although the low-level inflow to the Big Thompson, Rapid City, and Fort Collins systems occurred at a much higher elevation and exhibited smaller absolute humidities than the Madison County event, their environments were more unstable. For example, the lifted index of the inflow air was around -3 to -6 in the western events, as opposed to -2 in the Madison County event. The inversion was also weaker in the Madison County event, requiring less lifting to release the available CAPE. Of course, the larger CAPE in the western systems allowed cloud tops in those systems to reach greater heights.



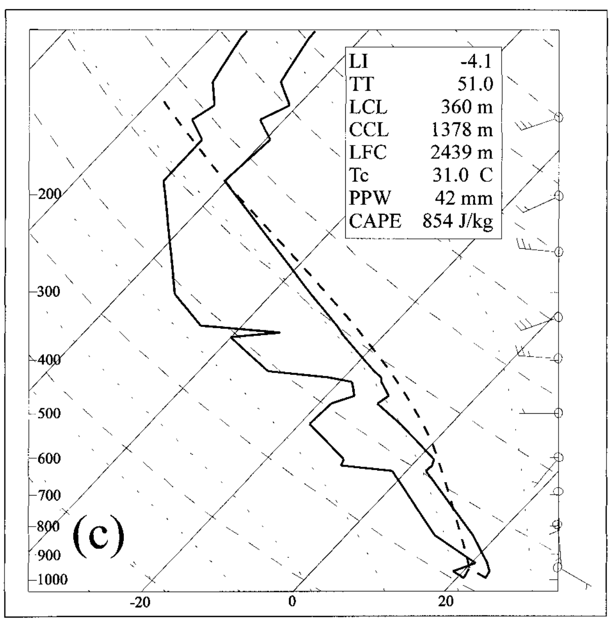
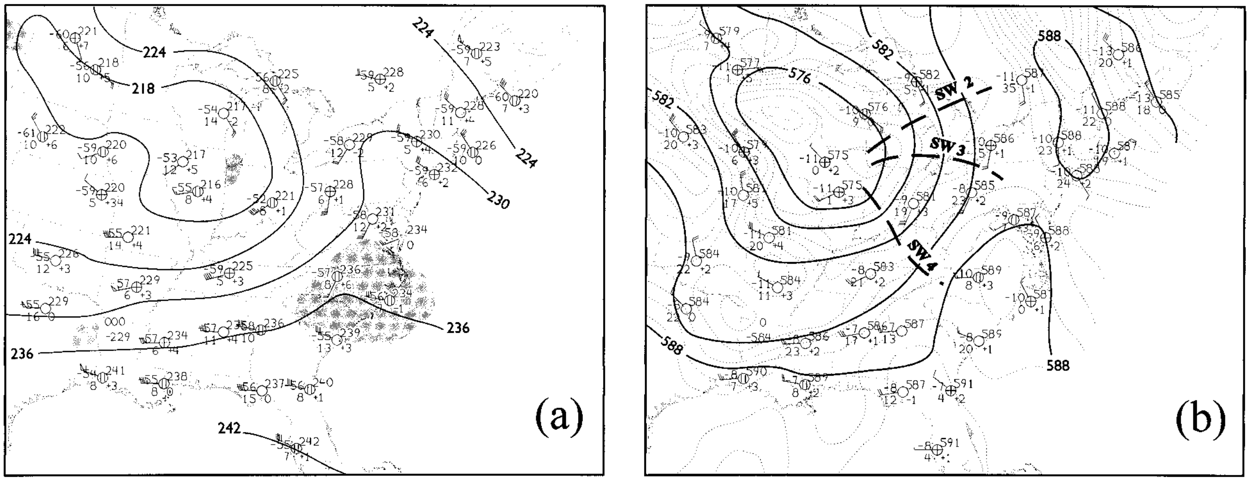


FIG. 19. Soundings for 0000 UTC 28 Jun 1995: (a) Sterling, VA; (b) Wallop's Island, VA; and (c) Greensboro, NC. Insert and parcel paths as in Fig. 5.

The moist low-level inflows resulted in very low lifting condensation levels, which promoted the development of the stratocumulus clouds observed moving rapidly toward the mountains in the Madison County, Big Thompson, and Rapid City events. The very low lifting condensation levels also dictated that cloud bases were near or at ground level in the higher elevation regions where the heavy rains developed. As a result, the formation of subcloud-layer evaporationally driven moist downdrafts was absent or very limited. Furthermore, the moist troposphere also limited entrainment of dry air and hindered development of evaporationally driven downdrafts. Still further, these three systems exhibited a low-echo centroid. As pointed out by Caracena et al. (1979), the combination of a low-echo centroid, high freezing level, and very low cloud bases indicates that warm rain processes must have played a significant role in these events. Thus, with little cooling from melting or evaporation, the moist downdraft production of cold pools and their associated outflows would be very limited.

The degree to which cool moist downdrafts were present in these systems is very important. The presence of strong downdrafts can cause systems to propagate faster,



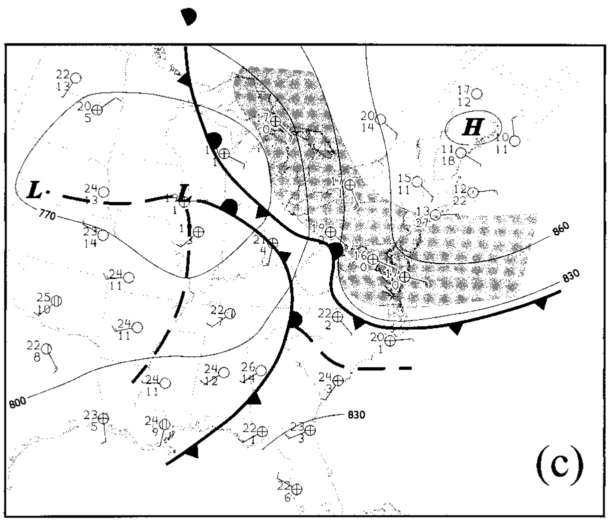


FIG. 20. Synoptic upper-air analyses for 0000 UTC 28 Jun 1995. Conventions as in Fig. 3.

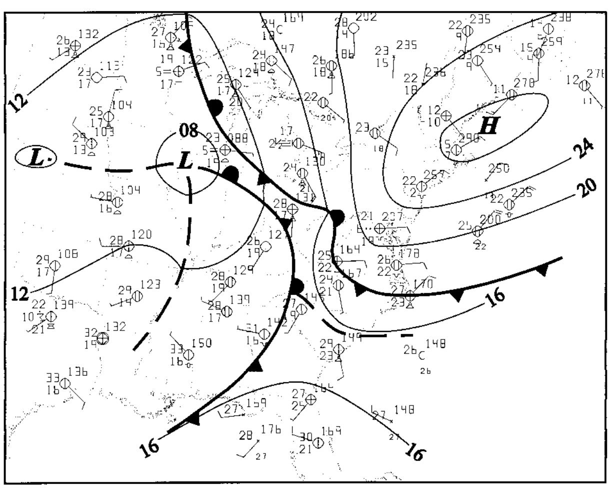


FIG. 21. Synoptic surface analysis for 0000 UTC 28 Jun 1995. Contour interval is 4 mb.

which, in the present case, would likely have displaced the systems from the higher terrain and reduced the rainfall amounts. Because the downdrafts in these events tended to be very weak, the terrain remained the primary focusing mechanism. As the ribbon of strong, moist, and conditionally unstable, low-level easterly winds approached the mountains, the capping inversion was breached or lifted and convection initiated. Because the low-level easterly jet was several hundred kilometers wide, it was *local terrain features* that determined exactly *where* along the ridges the convection would be focused.

In the Rapid City and Big Thompson floods, there was a significant reduction in temperature behind the front. Postfrontal temperatures were sufficiently low to prevent the release of the instability until the air was forced up the mountain. This was not the case in the Madison County event, where there was almost no drop in temperature immediately behind the front. Indeed, if the low-level deck of stratocumulus clouds and the high-level outflow from the Piedmont system had not been

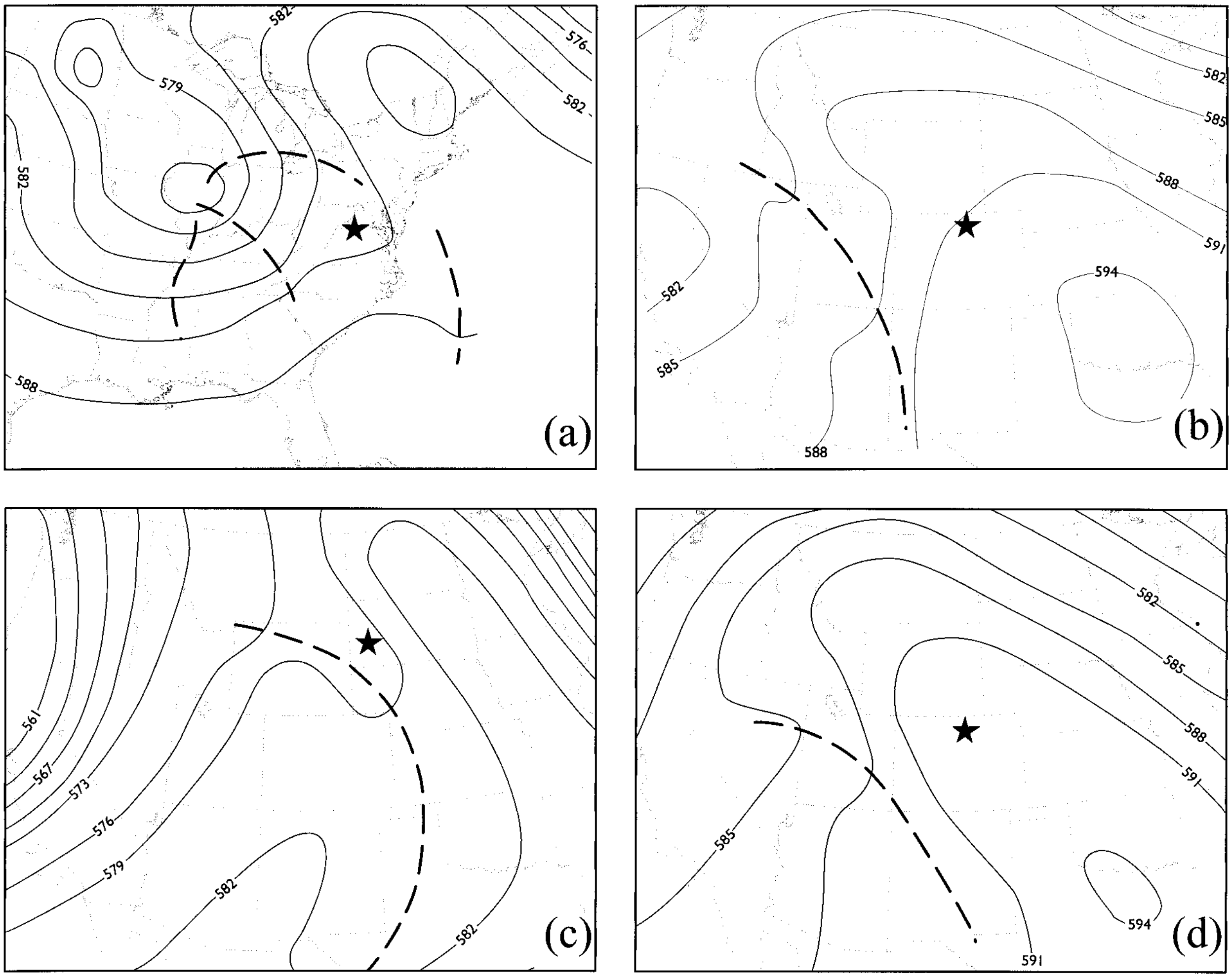


FIG. 22. The 500-mb height analyses near the onset time of severe leeside flash floods: (a) Madison County, VA (1200 UTC 27 Jun 1995); (b) Fort Collins, CO (0000 UTC 29 July 1997); (c) Rapid City, SD (0000 UTC 10 Jun 1972); and (d) Big Thompson Canyon, CO (0000 UTC 1 August 1976). Shown are storm location (star), heights (solid lines, 3-dam intervals), and shortwave trough axes (bold dashed lines). (c) and (d) From Maddox et al. (1977).

present, it is conceivable that the convective temperature would have been achieved across northern Virginia, disrupting the mechanisms that produced the quasi-stationary Madison system.

The postfrontal low-level easterly jet in the Big Thompson, Fort Collins, and Rapid City cases was also characterized by a significant moist ribbon (Fig. 23). Dewpoints in the ribbon were 3°–6°C higher than surrounding areas. In contrast, there was no significant increase in dewpoints in the postfrontal air of the Madison County event and there was no east–west gradient in moisture. Therefore, the easterly winds could not advect in air that was more moist than the air that was already in place south of the front. It is possible, however, that the maritime properties of the postfrontal flow off the Atlantic enhanced the warm cloud microphysical processes that produced the heavy rain rates. It is also possible that the stronger postfrontal easterly flow was necessary to force sufficient orographic lifting to eliminate the capping inversion and locally increase CAPE.

While the moisture ribbon provided an additional focus for the Big Thompson and Rapid City events, the Madison system had to rely on other focusing mechanisms. The zone between air that had achieved its convective temperature and air that was too stable narrowed the region where terrain-focused convection was possible. The system developed where the low-level jet of conditionally unstable air first encountered mountains high enough to break the capping inversion and release the instability, which happened to be the Blue Ridge in Madison County.

Despite these minor differences, the end result was the same in all four events. Individual cells continually initiated along the southern flank of the system and were slowly advected northward by the weak steering flow. As the cells passed into and through the system, they

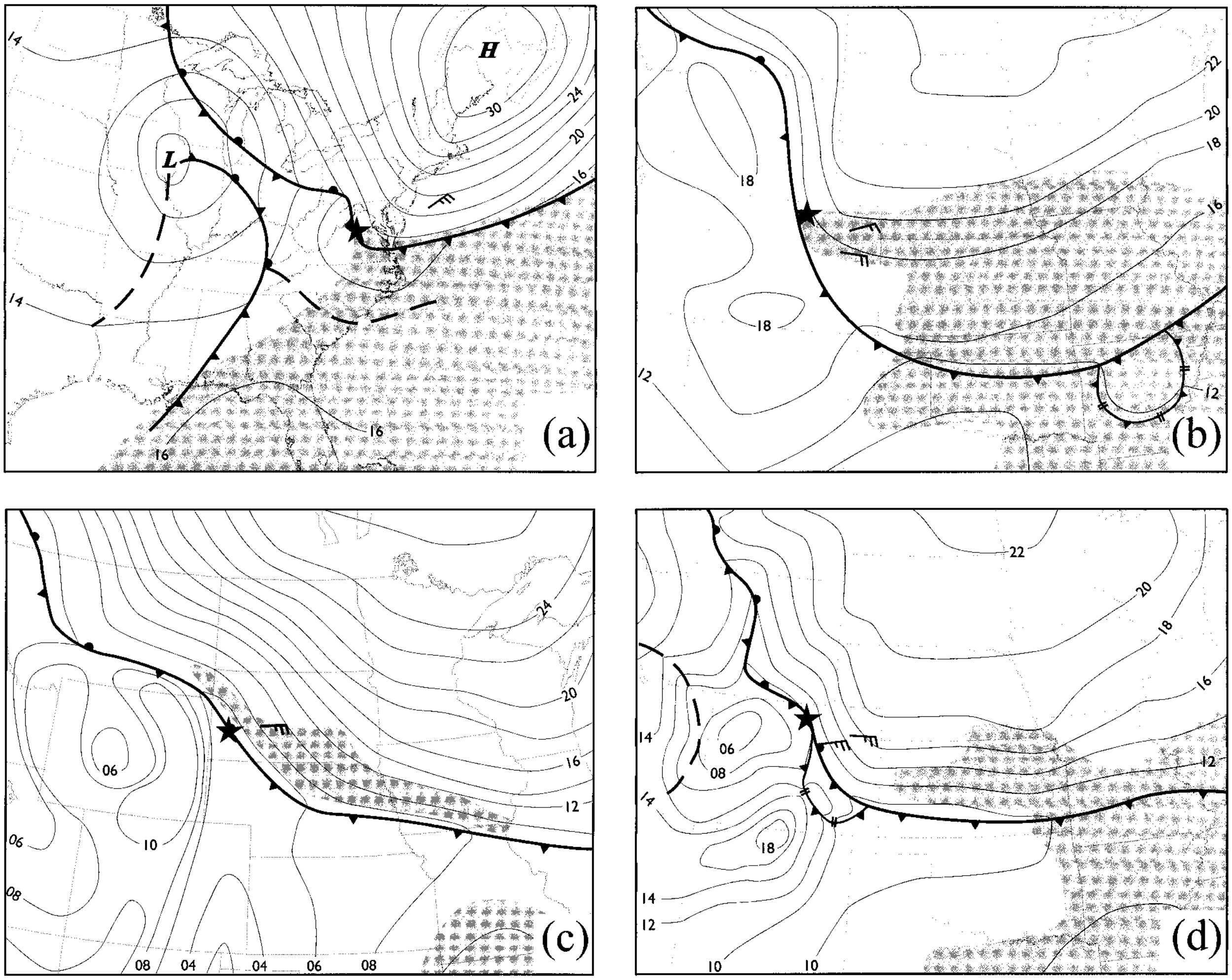


FIG. 23. Surface analyses near the onset time of severe leeside flash floods: (a) Madison County, VA (1200 UTC 27 Jun 1995); (b) Fort Collins, CO (0000 UTC 29 Jul 1997); (c) Rapid City, SD (0000 UTC 10 Jun 1972); and (d) Big Thompson Canyon, CO (0000 UTC 1 Aug 1976). Shown are surface winds near the storm (full barb = 5 m s^{-1}), dewpoint > 22.0°C [shading, (a)], 17.7°C [shading, (b)], and dewpoint > 18.3°C [shading, (c) and (d)]. (c) and (d) Adapted from Maddox et al. (1977).

first matured and then weakened and eventually dissipated, thereby maintaining an overall quasi-stationary convective entity. The combination of minimal dry air entrainment and warm rain processes made the systems efficient producers of precipitation. Focusing of convection occurred where the conditionally unstable high-θ_e moist ribbon intersected the higher terrain. The very slow movement of the axis of postfrontal moist inflow allowed convection to remain over the same area for several hours, thereby generating the extremely heavy rainfall. These conditions exhibit many of the ingredients that Doswell et al. (1996) noted are typically present with major flash floods.

5. Concluding remarks

The atmospheric conditions responsible for the Madison County flood bore a striking resemblance to other devastating upslope floods in the United States, particularly the Big Thompson and Fort Collins floods along the eastern slopes of the Colorado Rockies and the Rapid City flood along the eastern slopes of the Black Hills of South Dakota. The similarity of the large-scale environments of the *western* flash floods and the particular mesoscale features instrumental in the production of the excessive rainfall was first recognized by Maddox et. al. (1978). The Madison County event demonstrates that the same general pattern and ingredients can result in major leeside flash flooding along the lesser orographic features of the *eastern* United States. It also demonstrates that the exceedingly heavy rainfall with these events does not require large amounts of CAPE.

It is not known how often this combination of synoptic and mesoscale features does *not* result in terrain-induced quasi-stationary convective systems with heavy rainfall. Nevertheless, forecasters should be aware that this type of situation *sometimes* results in excessive rainfall. Although this paper focused on a single scenario

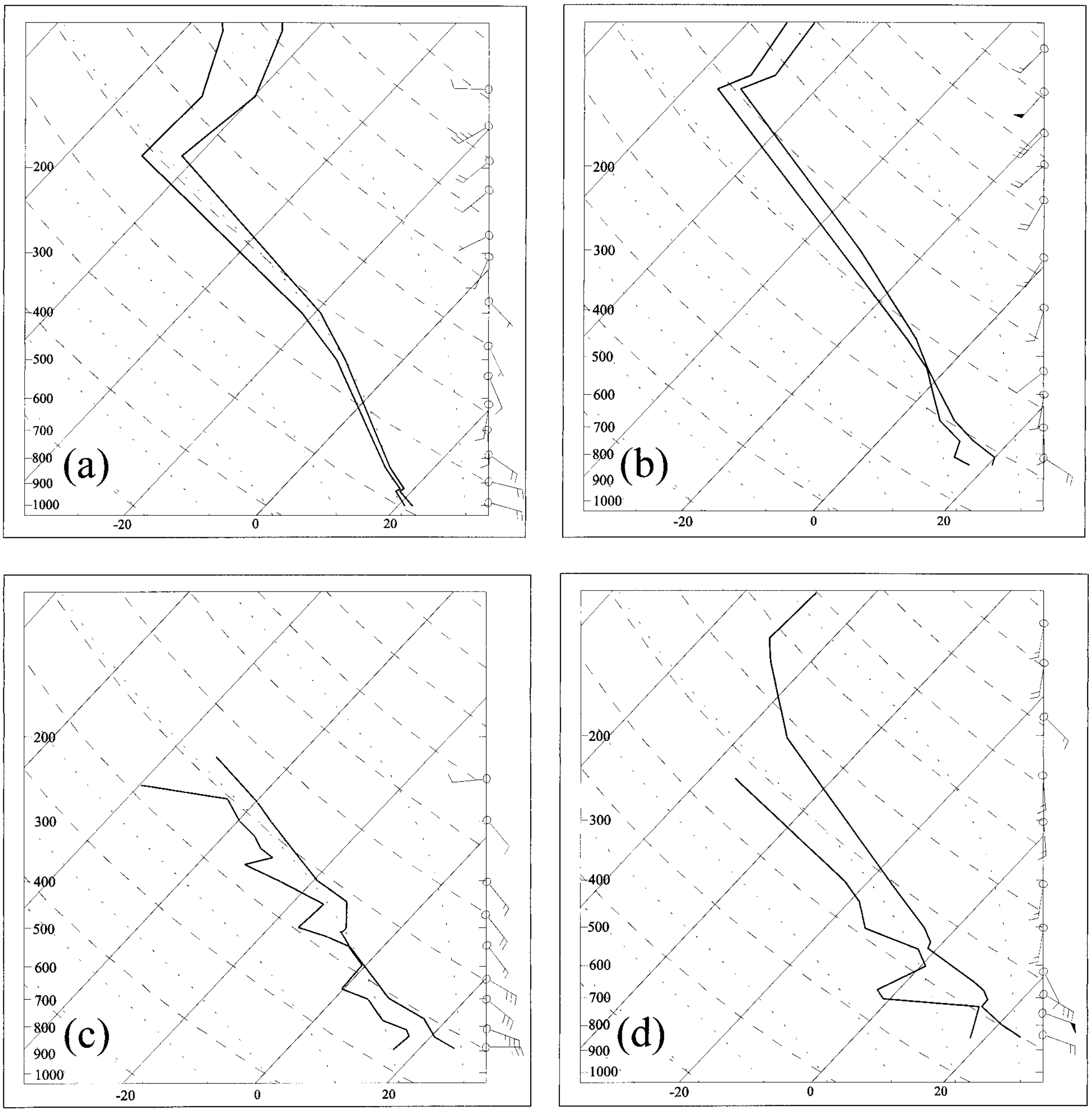


FIG. 24. Soundings near the onset time of severe leeside flash floods: (a) Madison County, VA (1200 UTC 27 Jun 1995, Sterling, VA); (b) Fort Collins, CO (0000 UTC 29 Jul 1997, Denver, CO); (c) Rapid City, SD (0000 UTC 10 Jun 1972, Rapid City, SD); and (d) Big Thompson Canyon, CO (0000 UTC 1 Aug 1976, reconstructed Loveland, CO). (c) and (d) Adapted from Maddox et al. (1977).

in which terrain-forcing coupled with cloud-scale and mesoscale processes permitted the genesis of new cells to remain quasi-stationary, Maddox et al. (1979), Chappell (1986), and Doswell et al. (1996) make it clear that there are other scenarios in which a slow-moving or quasi-stationary convective system can occur. For example, midlatitude west-slope (windward) upslope events have been documented wherein the local terrain features induce deep convective overturning of moist postfrontal Pacific air (Reynolds 1997). Similar to the east-slope events addressed in this investigation, the conditions favourable for west-slope events also evolve very slowly and can generate excessive rainfall as they remain quasi-stationary for several hours. It is therefore important that forecasters for any region with significant orographic relief recognize the meteorological ingredients that are responsible for these events (see Doswell et al. 1996). In particular, as pointed out by Maddox et

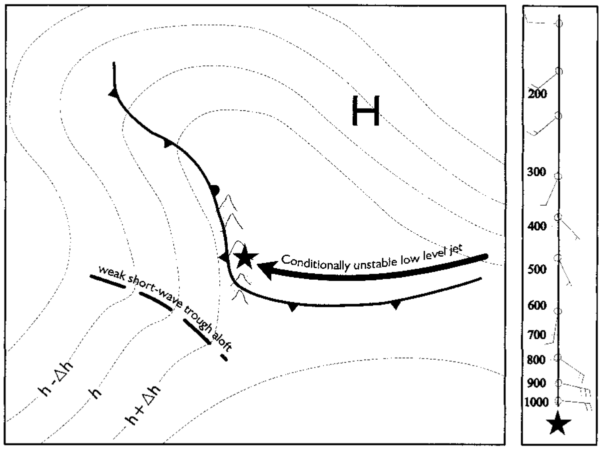


FIG. 25. Idealized schematic of the synoptic pattern associated with terrain-induced convective flooding events. Shown are the threat region (star), 500-mb height pattern (dotted lines, Δh = 3dam), mountains (shaded), and typical wind profile (full barb = 5 m s^{-1}) above threat region.

al. (1979) and Schwartz et al. (1990), forecasters should be wary any time a moist, conditionally unstable low-level jet is forced to rise over a stationary/slow-moving front, boundary, or orographic feature, especially if the low-level jet is expected to intersect the feature at the same location for an extended period of time.

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REFERENCES

Anthes, R. A., 1977: A cumulus parameterization scheme utilizing a one-dimensional cloud model. *Mon. Wea. Rev.,* **105,** 270–286.

Barnes, S. L., 1964: A technique for maximizing details in numerical weather map analysis. *J. Appl. Meteor.,* **3,** 396–409.

Braham, R. R., Jr., L. J. Battan, and H. R. Byers, 1957: Artificial nucleation of cumulus. *Cloud and Weather Modification: A Group of Field Experiments, Meteor. Monogr.,* No. 11, Amer. Meteor. Soc., 47–85.

Cahir, J. J., J. M. Norman, and D. A. Lowry, 1981: Use of a real-time computer graphics system in analysis and forecasting. *Mon. Wea. Rev.,* **109,** 485–500.

Caracena, F., R. A. Maddox. L. R. Hoxit, and C. F. Chappell, 1979: Mesoanalysis of the Big Thompson system. *Mon. Wea. Rev.,* **107,** 1–17.

Chappell, C. F., 1986: Quasi-stationary convective events. *Mesoscale Meteorology and Forecasting,* P. S. Ray, Ed., Amer. Meteor. Soc., 289–310.

——, and D. M. Rodgers, 1988: Meteorological analysis of the Cheyenne, Wyoming flash flood and hailstorm of 1 August 1985. NOAA Tech. Rep. ERL 435-FSL 1, 51 pp.

Corfidi, S. F., J. H. Merritt, and J. M. Fritsch, 1996: Predicting the movement of mesoscale convective complexes. *Wea. Forecasting,* **11,** 41–46.

Dennis, A. S., R. A. Schleusener, J. H. Hirsch, and A. Koscielski, 1973: Meteorology of the Black Hills flood of 1972. Institute of Atmospheric Sciences, Rep. 73-4, 41 pp. [Available from Institute of Atmospheric Sciences, South Dakota School of Mines and Technology, Rapid City, SD 57701.]

Doswell, C. A., H. E. Brooks, and R. A. Maddox, 1996: Flash flood forecasting: An ingredients-based methodology. *Wea. Forecasting,* **11,** 560–581.

Forbes, G. S., R. A. Anthes, and D. W. Thomson, 1987: Synoptic and mesoscale aspects of an Appalachian ice storm associated with cold-air damming. *Mon. Wea. Rev.,* **115,** 564–591.

Foster, D. S., 1958: Thunderstorm gusts compared with computed downdraft speeds. *Mon. Wea. Rev.,* **86,** 91–94.

Fritsch, J. M., J. Kapolka, and P. A. Hirschberg, 1992: The effects of subcloud-layer diabatic processes on cold air damming. *J. Atmos. Sci.,* **49,** 49–70.

Giordano, L. A., and J. M. Fritsch, 1991: Strong tornadoes and flash-food producing rainstorms during the warm season in the mid-Atlantic region. *Wea. Forecasting,* **6,** 437–455.

Hoxit, L. R., and Coauthors, 1978: Meteorological analysis of the Johnstown, Pennsylvania flash flood, 19–20 July 1977. NOAA Tech. Rep. ERL 401-APCL 43, 71 pp.

Maddox, R. A., F. Caracena, L. R. Hoxit, and C. F. Chappell, 1977: Meteorological aspects of the Big Thompson flash flood of 31 July 1976. NOAA Tech. Rep. ERL 388-APCL 41, 83 pp.

——, L. R. Hoxit, C. F. Chappell, and F. Caracena, 1978: Comparison of meteorological aspects of the Big Thompson and Rapid City flash floods. *Mon. Wea. Rev.,* **106,** 375–389.

——, C. F. Chappell, and L. R. Hoxit, 1979: Synoptic and meso-α scale aspects of flash flood events. *Bull. Amer. Meteor. Soc.,* **60,** 115–123.

Marwitz, J., and J. Toth, 1993: The Front Range blizzard of 1990. Part I: Synoptic and mesoscale structure. *Mon. Wea. Rev.,* **121,** 402–415.

National Oceanic and Atmospheric Administration, 1972: Black Hills flood of June 9, 1972. Natural Disaster Survey Rep. 72-1, 20 pp.

——, 1995: *Storm Data.* Vol. 37, 200–215.

Pearce, M. L., and S. G. Hoffert, 1997: The SNAP Weather Information Integrated Forecast Tool (SWIIFT): Next generation software for research and operational meteorology. Preprints, *13th Int. Conf. on Interactive Information and Processing Systems (IIPS) for Meteorology, Oceanography, and Hydrology,* Long Beach, CA, Amer. Meteor. Soc., 269–272.

Peppler, R. A., and P. J. Lamb, 1989: Tropospheric static stability and central North American growing season rainfall. *Mon. Wea. Rev.,* **117,** 1156–1180.

Peterson, W. A., and Coauthors, 1999: Mesoscale and radar observations of the Fort Collins flash flood of 28 July 1997. *Bull. Amer. Meteor. Soc.,* **80,** 191–216.

Reynolds, D., 1997: Cool season mesoscale convective trains. Western Region Tech. Attachment 97-27, 5 pp.

Rotunno, R., J. B. Klemp, and M. L. Weisman, 1988: A theory for strong, long-lived squall lines. *J. Atmos. Sci.,* **45,** 463–485.

Schwartz, B. E., C. F. Chappell, W. E. Togstad, and X.-P. Zhong, 1990: The Minneapolis flash flood: Meteorological analysis and operational response. *Wea. Forecasting,* **5,** 3–21.

Shi, J., and R. A. Scofield, 1987: Satellite observed mesoscale convective system propagation characteristics and a 3–12 hour heavy precipitation forecast index. NOAA Tech. Memo. NESDIS 20, Washington, DC, 43 pp.

Smith, J. A., M. L. Baeck., M. Steiner, and A. J. Miller, 1996: Catastrophic rainfall from an upslope thunderstorm in the central Appalachians: The Rapidan Storm of June 27, 1995. *Water Resour. Res.,* **32,** 3099–3113.

Srivastava, R. C., 1987: A model of intense downdrafts driven by the melting and evaporation of precipitation. *J. Atmos. Sci.,* **44,** 1752–1773.

Szeto, K. K., C. A. Lin, and R. E. Stewart, 1988a: Mesoscale circulations forced by melting snow. Part I: Basic simulations and dynamics. *J. Atmos. Sci.,* **45,** 1629–1641.

——, R. E. Stewart, and C. A. Lin, 1988b: Mesoscale circulations forced by melting snow. Part II: Application to meteorological features. *J. Atmos. Sci.,* **45,** 1642–1650.

Thompson, H. J., 1972: The Black Hills flood. *Weatherwise,* **25,** 162–167.

Young, G. S., and J. M. Fritsch, 1989: A proposal for general conventions in analyses of mesoscale boundaries. *Bull. Amer. Meteor. Soc.,* **70,** 1412–1421.