Frontal Focusing of a Flooding Rainstorm

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ABSTRACT

A heavy rainstorm over Kentucky, producing extensive flooding, was concentrated in a narrow band oriented nearly zonally just south of the Ohio River. Analysis of routine surface observations showed that an intense quasi-stationary surface front formed during the 24-h period of heaviest rainfall. This front was parallel to the rainband and was some distance to the south of it. Horizontal temperature gradients reached more than 20°F over 110 km. Analysis of sea level pressure showed that geostrophic deformation was present in a small region ahead of each of two small centers of low pressure that migrated eastward along the front. Vertical cross sections normal to the front showed that conditional upright and symmetric stabilities were small or negative in the frontal updraft. It was inferred from this that the frontal updraft was unusually intense and narrow, qualitatively consistent with the intensity of the rainband.

1. Introduction

Although reliance on traditional fronts as a means of explaining what is happening in the weather may be grossly overdone in both research and operational forecasting, there are occasions when a front really plays a central role. Such a case may be the rainstorm of 12–16 February 1989, which was responsible for major flooding in parts of west and central Kentucky, causing loss of life and property. Incidentally, the flood was responsible for failure of the 0000 UTC 14 February sounding from Paducah. About half the 4-day total fell in the 24-h period ending at 1200 UTC 14 February. As can be seen in Fig. 1, the precipitation was concentrated in a band in Kentucky just south of the Ohio River, with a half-width of about 100 km. We will focus on this area and on this period.

The situation has been discussed in detail by Kirkpatrick (1992; hereafter denoted K92). In that study it was pointed out that there was very little evidence of forcing in the middle or upper troposphere, as the 500-mb charts showed little perturbation of strong west-southwesterly flow. Instead, emphasis was placed on processes in the lower troposphere. Analyses were presented of equivalent potential temperature and its advection by the flow at 850 mb, and of flow and mixing ratio on dry isentropic surfaces in the layer between 950 and 650 mb. Considerable instability was inferred from surface temperatures and dewpoints that were high for the region and time of year, but no explicit information on cumulus convection was presented. Deep convection was implied, however, by the cold tops of satellite-observed cloudiness and by radar summaries.

A frontal analysis was shown and was generally considered to provide a lifting mechanism for the moist air. No analysis of the surface temperature field, however, was presented. K92's analysis at 1200 UTC on 13 February showed a cold front from Oklahoma to Iowa with a warm front running eastward from the southeastern corner of Kansas to eastern Tennessee. Twelve hours later the cold front was shown from central Texas to eastern Illinois, while the warm front was analyzed slightly north of the earlier position. At 1200 UTC on 14 February, the cold front was shown from southeastern Texas to eastern Ohio, with the warm front southeastward from eastern Kentucky. The frontal analysis on the maps of the National Weather Service received on fascimile during this period (not shown) was similar, with a front moving from Iowa to Pennsylvania and a second separate front a short distance to its southeast, moving relatively little.

The purpose of this paper is to consider the frontal structure in the light of the evolution of the surface temperature field. We also assess the possibility that small moist symmetric stability may have been responsible for an implied strong and concentrated sheet of ascending air at the place and time of most intense rain.

2. Surface analysis

The surface observations in K92 were replotted for the three times discussed above, and sea level isobars

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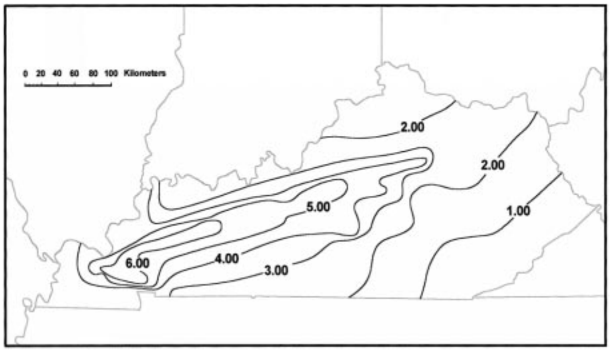

FIG. 1. Precipitation (in.) for the 24-h period ending 2300 UTC 14 Feb 1989. Same as Kirkpatrick (1992, Fig. 5).

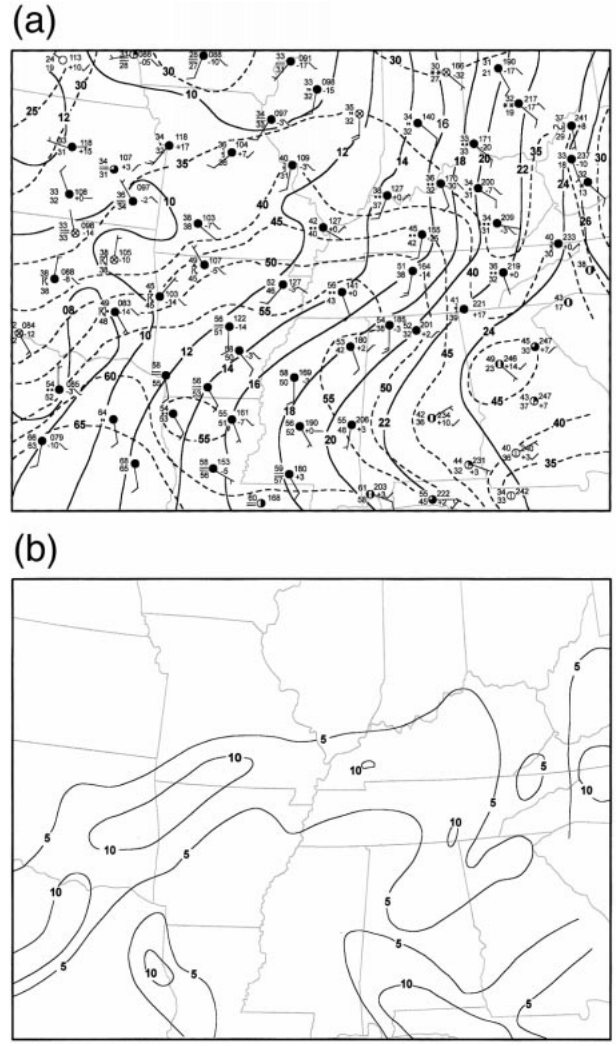

FIG. 2. (a) Sea level isobars at intervals of 4 mb (solid) and surface isotherms at intervals of 5°F (dashed) for 1200 UTC 13 Feb. The observations are transcribed from K92, and the station plotting model is standard. (b) Horizontal temperature gradient in °F per 110 km.

and surface isotherms were drawn, as shown in Figs. 2–4. In the temperature field, there was a single significant baroclinic zone. At the initial time, Fig. 2a, the zone extended from south-central Kentucky to southeastern Oklahoma. The temperature gradient was measured by setting the ends of a pair of dividers at 1° of latitude (110 km) and noting when the spacing of two or more isotherms (at intervals of 5°F) was greater or smaller than this interval. The gradient, as seen in Fig. 2b, was between 5° and 10°F per 110 km, equivalent to moderate intensity, according to the criterion proposed by Sanders (1999a), along most of its length. By 0000 UTC on 14 February (Fig. 3) it had reached more than 15°F per 110 km, equivalent to strong intensity by this criterion along its entire length. It remained strong to the end of the period of study. Figure 4a shows the zone moving southeast about 350 km during the 24-h period at the western edge of the analysis area but moving little at its eastern extremity. In Fig. 4b the gradient strengthened slightly to a value of about 20°F per 110 km from southern Kentucky to northwestern Arkansas. There was no evidence of a significant gradient accompanying an imputed cold front extending northward from this region.

In the field of sea level pressure at the start of the period of study (Fig. 2a) there was a weak trough extending eastward from a low center in eastern Oklahoma. At the end, Fig. 4a shows lows in northeastern Kentucky (north of the baroclinic zone) and in northern Mississippi at the warm edge of the zone. A substantial large-scale pressure rise had occurred over the entire region, although there had been prominent frontogenesis.

It was difficult to be confident of the continuity of the lows during the period, because there were a number of oscillations of the baroclinic zone, as shown in Fig. 5. In fact, there was a suggestion of yet another weak low at the end of the period in eastern Texas, as seen in Fig. 4a. Thus the baroclinic zone was similar to the ‘‘steering line’’ described by Bjerknes (1919), later termed a warm front, since it extended in the path of one or more weak lows. In this case, however, the warmer air made little or no advance and rather was replaced by cold air in the western portion of the analysis area, as noted above.

A comparison of Figs. 5 and 1 shows that the strip of maximum rainfall lay parallel to and slightly north of the center of the baroclinic zone. This offset may have been due to a northward tilt of the presumed region of frontal ascent. Further investigation of the production of rain in this case is beyond the scope of this study.

3. Frontogenesis

We wish, however, to study the process of frontogenesis that produced this intense baroclinic zone. First, we examine the geostrophic flow in the region where the intensification occurred. From the simple expression presented by Miller (1948), when diabatic effects and

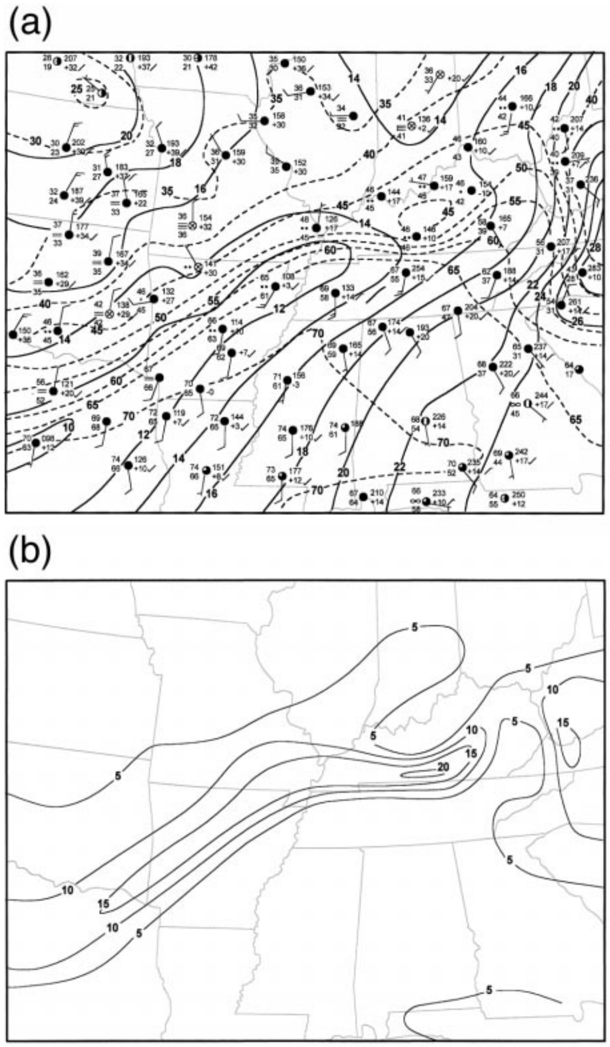

FIG. 3. Same as Fig. 2 but for 0000 UTC 14 Feb.

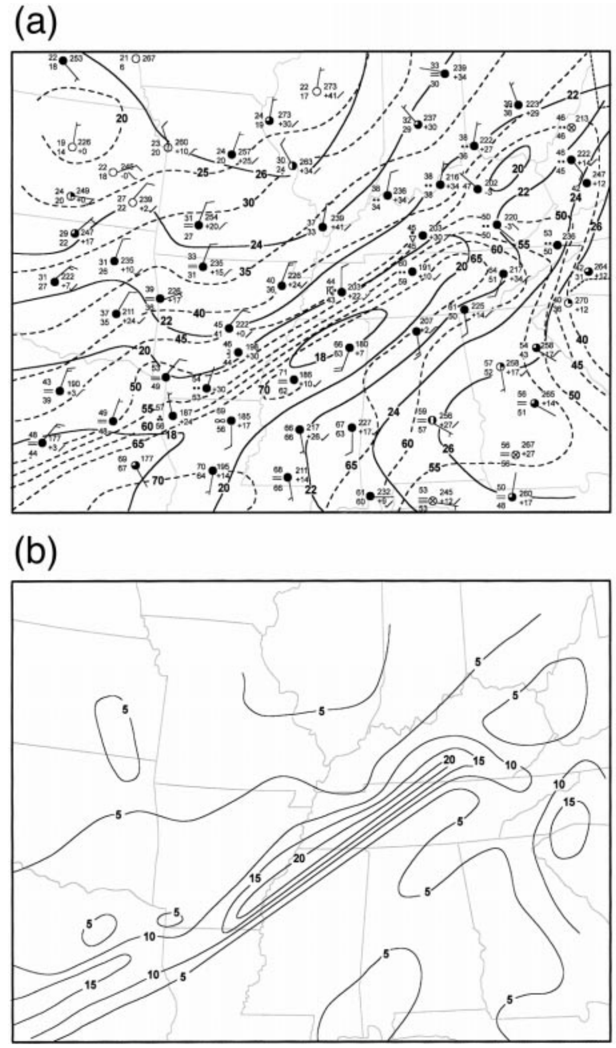

FIG. 4. Same as Fig. 2 but for 1200 UTC 14 Feb.

vertical motions are ignored, and with the *x* axis taken along the isotherms, colder air lying in the positive *y* direction, the relative rate of change of the potential temperature gradient is given by

$$d \ln(-\partial\theta/\partial y)/dt = -\partial v/\partial y \approx -\delta v/\delta y, \qquad (1)$$

where δv is measured at the cold and warm boundaries of the baroclinic zone with width δy.

Close inspection of the isobars in Figs. 2–4 indicates that the southerly geostrophic wind component on the cold edge of the intensifying baroclinic zone is weaker than the same component on the warm edge. Therefore there is geostrophic confluence. Measurements made of pressure differences over distances of about 400 km along the 40° and 50°F isotherms at 1200 UTC on 13 February gave geostrophic components normal to the isotherms of 6.1 and 7.7 m s^{-1}, respectively. This difference of components occurred over a distance of about 100 km. Twelve hours later similar components were estimated to be 6.3 and 11.1 m s^{-1} over a distance of about 50 km, while at 1200 UTC on 14 February these components were 5.7 and 8.8 m s^{-1}. Averaging these values for the first 12 h yielded a change in $\ln(-\delta\theta/\delta y)$ of 2.4, corresponding to an increase in gradient by a factor of 11.2. Similarly, for the second 12 h the implied increase was by a factor of 29.1. The interpretation of these measurements cannot be taken at face value, however, because they would require contrasts over a 110-km distance to reach between 100° and 600°F, far greater than any meteorologically reasonable value. We infer that the available contrast took place over a distance much smaller than 100 km, but we cannot say how much, given the available observations. It may be more sensible to calculate the time required for the gradient to double, given the observed geostrophic confluence. These times vary between 2.5 and 2.9 h.

It is still difficult to interpret these results, in the first place because Eq. (1) refers to a total derivative, the

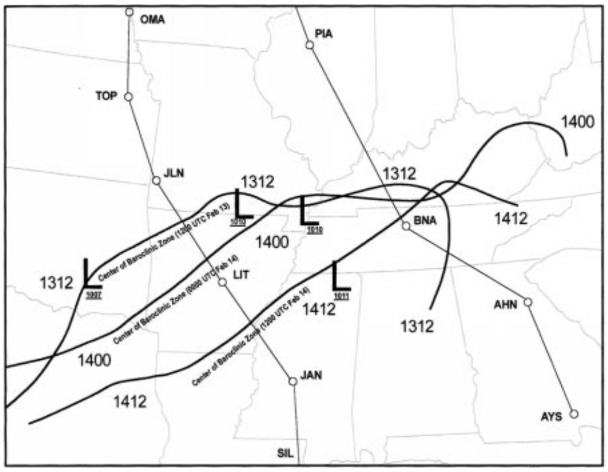

FIG. 5. Positions of the center of the frontal baroclinic zone at the three times indicated by the date and time UTC. The light solid lines show the position of the cross sections in Figs. 6 and 7, together with the identifiers of the stations from which the soundings were used.

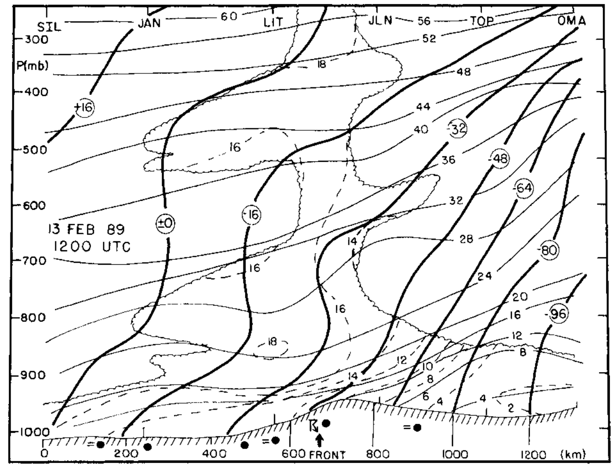

FIG. 6. Vertical cross section from Slidell, LA (SIL), to Omaha, NE (OMA), at 1200 UTC 13 Feb. Thin solid lines are potential isotherms at intervals of 4°C. The scalloped line encloses the estimated region of saturation, within which wet-bulb potential isotherms are shown as dashed lines at intervals of 2°C. Heavy solid lines are isotachs of geostrophic absolute momentum, $u - fy$, at intervals of 16 m s^{-1}, where u is the geostrophic wind component from 250° and the zero for the y axis is taken at the southern end of the section. Observations of sky cover and weather are shown along the abscissa, with the position of the front.

rate of change experienced by a moving air parcel. Interpretation requires an estimate of the relative motions of air parcels and the baroclinic zone. The latter moves little in the region of the calculation, as noted above, and the details of the airflow are uncertain in the vicinity of the zone. Geostrophically, the gradient on air parcels could easily double in less than 3 h as they entered the baroclinic zone from the south. The actual airflow, however, contained a substantial ageostrophic component after 1200 UTC on 13 February, as seen in Figs. 3 and 4, especially at the cold edge of the baroclinic zone. Thus that there may have been little actual (compared to geostrophic) flow relative to the zone during the latter 12 h of the period under study.

Further, with the available observations we can determine the gradient only over a distance of 100 km or more, as noted above, while the presumed frontogenesis probably occurred on a scale of 10 km or less. As noted above, the winds in the baroclinic zone in Figs. 3a and 4a have a decided ageostrophic component from cold to warm air, implying that a collapse toward discontinuity, as described by Hoskins and Bretherton (1972), had occurred.

So far as diabatic effects are concerned, small-scale mixing may have been diluting the gradient on the 10-km scale at almost the same rate as the larger-scale circulation was increasing it. An instance of a qualitatively similar scenario was documented by Sanders (1999b). On the other hand, on the observable 100-km scale, a diabatic effect through surface heat flux was evidently much stronger in the warm air than in the cold. That is, the temperatures in the warm air at 0000 UTC on 14 February are much warmer than could be explained by advection and must be attributable to daytime heating. Heavier cloudiness and rain in the colder air prevented such heating, so that the contrast was enhanced. This contrast was also found in the case described by Sanders (1999b), and is believed to be quite common.

4. Symmetric stability

The concentration of precipitation in the front-parallel band over northern Kentucky was sufficiently intense to prompt a consideration of the symmetric stability of the flow in the warm air aloft. Emanuel (1985) and Thorpe and Emanuel (1985) have shown that the ascent in a frontal circulation can be very narrow and intense when the moist symmetric stability is small.

To evaluate this stability, we prepared two vertical cross sections from south-southeast to north-northwest, through regions of heavy precipitation, approximately normal to the flow aloft (Figs. 6 and 7). The first of these, for 1200 UTC on 13 February, runs from Slidell, Louisiana, to Omaha, Nebraska, while the second, for 0000 UTC on 14 February, extends from Waycross, Georgia, to St. Cloud, Minnesota. Both show analyses of potential temperature (and where the atmosphere is inferred to be saturated, wet-bulb potential temperature) and ''absolute momentum,'' $u - fy$, where u is the geostrophic wind component from 250°, following the recommendation of Schultz and Schumacher (1999). The regions of saturation are inferred from surface observations of cloud and precipitation, from the soundings of relative humidity, and from the satellite imagery. Where the wet-bulb potential temperature decreases upward in a saturated region, the air is conditionally unstable for upright convection, if the effects of condensate loading and perturbation pressure gradient are neglect-

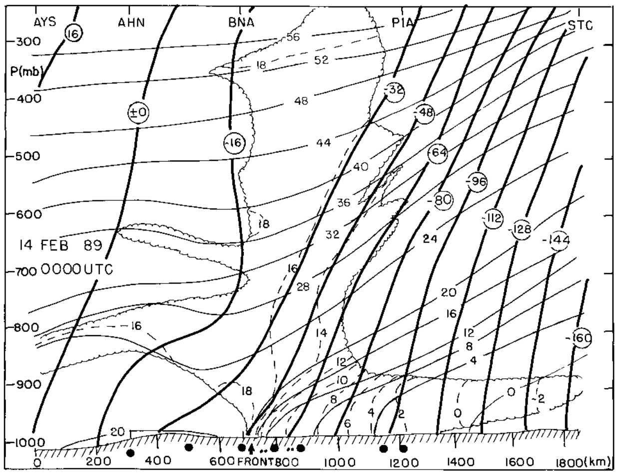

FIG. 7. Same as Fig. 6 but for a section from Waycross, GA (AYS), to St. Cloud, MN (STC), at 0000 UTC 14 Feb.

ed. Where the isopleths of the thermodynamical variable are more nearly vertical than those of the absolute momentum (i.e., when the thermodynamical variable decreases upward along an isopleth of absolute momentum), symmetric instability can be said to exist, if we assume little variation of wind direction along the flow.

From an examination of Figs. 6 and 7 it is seen that except above the surface boundary layer at 0000 UTC, the atmosphere is symmetrically stable outside the areas of saturation. Within the cloudy regions the air is symmetrically neutral or slightly unstable up to about 600 mb at 1200 UTC and to about 500 mb at 0000 UTC. Further, some portions of the saturated regions are conditionally unstable for upright convection. Since the timescale of this convection is much smaller than that of slantwise convection, the former would be expected to dominate at first. The result of this cumulus convection, however, should be to eliminate this instability by producing a vertically constant wet-bulb potential temperature. If vertical shear of the large-scale wind is still present, the atmosphere would still be symmetrically unstable. It is therefore likely that the ascent associated with the front was intense, because of the presence of both upright and slantwise convection.

5. Conclusions

A major rainstorm in the south-central United States was highly concentrated in a nearly zonal band in Kentucky just south of the Ohio River. More than 6 in. of rain fell in the band, which had a half-width of about 100 km. During the 24-h period of heaviest rain, an intense quasi-stationary front developed approximately parallel to the rainband and some distance to its south. The front lay ahead of two or more small-amplitude low pressure centers, which propagated eastward along the developing front. In a limited region ahead of these centers there was geostrophic deformation that apparently was responsible for the collapse toward discontinuity of a preexisting temperature contrast brought about mainly by the horizontal gradient of surface heat flux. The frontogenesis thus appeared to represent a two-stage process. The ascent of the warm air in the front is believed to be especially strong and concentrated because the conditional upright and symmetric stabilities of air in this region were evidently small or negative.

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