A Short-Lived Cold Front in the Southwestern United States

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ABSTRACT

A cold front developed in Arizona in a region of initially small temperature gradient, developing to great intensity, accompanied by damaging winds over New Mexico, and then losing strength over the high plains of Oklahoma and Texas. The entire development of frontogenesis and frontolysis occurred in no more than 24 h.

The initial growth of temperature contrast was attributable mainly to horizontal variation of surface heat flux during the morning, with little heating in a region of dense cloud cover and scattered showers in the west and with intense heating in a region of only thin high clouds to the east. The accompanying ageostrophic circulation then resulted in a collapse toward discontinuity. The frontal zone maintained an approximately steady state for a few hours in early afternoon. At this time the westerly component of surface wind just ahead of the zone was not as strong as the eastward motion of the zone. The passage of the zone was accompanied by a veering and strengthening of the surface wind so that westerly components were briefly larger than the frontal motion. The tendency of the convergent wind field to produce a frontogenesis was evidently balanced by small-scale mixing.

Subsequently the pressure trough and surface wind shift propagated eastward more rapidly than the frontal temperature contrast. The contrast quickly weakened as the mixing then was unopposed. Severe convection developed during the evening as the convergent surface wind shift came into contact with humid unstably stratified air.

1. Introduction

During the 24-h period beginning 1200 UTC 26 March 1991, a strong cold front developed in a region of little initial surface temperature gradient and grew to damaging intensity, with peak associated surface wind gusts ranging from 22 to 31 m s^{-1}. It raced eastward across Arizona and New Mexico and then dissipated over the southern plains, leaving a residual of severe convection. The purpose of this paper is to document the development and dissipation of this front and to propose a physical mechanism for the events observed.

The situation displayed considerable baroclinicity over a broad area. At 850 mb, for example (not shown), the temperature ranged from slightly above 20°C over Texas to as low as 1°C in southern California. This contrast, however, occurred over a distance of about 1100 km and weakened slightly over the time period studied. Nowhere in the analyses did the gradient exceed 5°C per 100 km. Our attention, on the other hand, was directed at a comparable contrast of as much as 18°C over a distance as small as 100 km, which developed and dissipated abruptly, producing damaging winds during its time of maximum strength.

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2. Synoptic situation

The frontogenesis was the surface manifestation of a mobile tropospheric trough of moderate intensity that moved northeastward from southern California to the central plains during the period of study (Fig. 1), as another mobile trough dropped southward into the location initially occupied by the first. The events in the surface boundary layer were studied by an examination of the National Weather Service (NWS) surface maps for the area (Fig. 2), and by a series of zonal sections of 3-h changes of surface temperature and pressure along 35°N lat (and along 33°N lat toward the end of the period of interest), where the major centers of change lay. The range of longitudes shown in each section moved east with the prominent changes in the fields of station pressure and surface temperature. Beyond the limits of each section the changes and their gradient were relatively modest. More details of the surface development were given by the routine hourly and special observations from selected stations in the region of interest.

The surface maps in Fig. 2 show a weak trough along the California–Arizona border at 1200 UTC 26 March. Twelve hours later, as the 500-mb trough moved into the Rocky Mountain region, a surface cyclone formed on the lee slopes in Colorado with a more intense trough extending south. Finally, at 1200 UTC of the 27th, the surface cyclone lay along the Kansas–Nebraska border

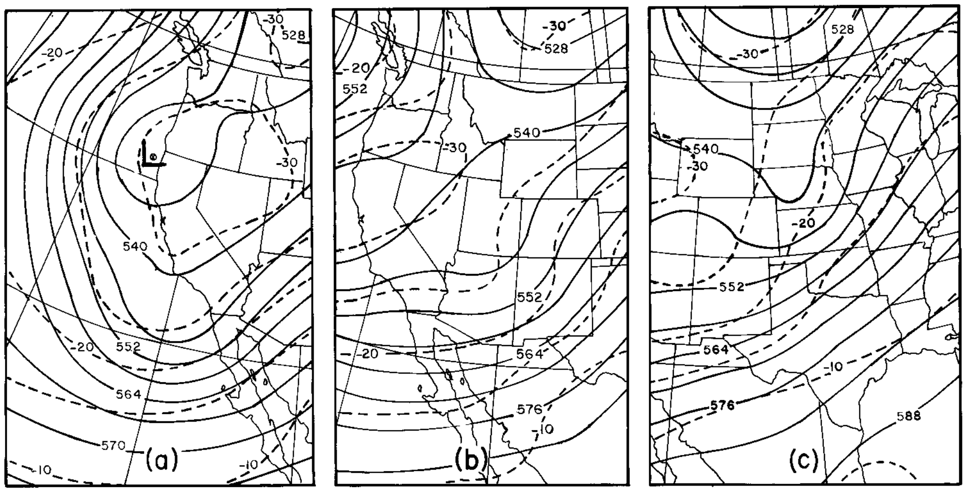


FIG. 1. Excerpts of NWS 500-mb analyses for (a) 1200 UTC 26 Mar 1991, (b) 0000 UTC 27 Mar 1991, and (c) 1200 UTC 27 Mar 1991. Solid lines are contours of geopotential height at intervals of 6 dam. Dashed lines are isotherm at intervals of 5°C.

with a broad trough to the south and a more pronounced one to the northeast. Analyzed frontal structure included two systems. One was oriented northeast–southwest and moved eastward with the cyclone. The other was zonally oriented and moved southward, approaching the cyclone center. The analysis also included a dryline and a number of troughs with limited continuity. The only extensive area of precipitation was associated with the latter, zonally oriented, frontal system and the cyclone as it propagated toward the Great Lakes. This analysis will deal with only the former front.

3. Analysis of the front

The development of the first frontal system is illustrated by the zonal sections of 3-h change of surface temperature and station pressure (Fig. 3). The changes were used to minimize the effects of highly variable elevation in the western United States. Initially (Fig. 3a), the changes of pressure and temperature were small with weak extrema near 120°W long, but 6 h later (Fig. 3b) there had been substantial growth, with prominent centers of pressure rise and temperature fall (opposing the overall diurnal rise) near 111°W. Growth continued until 0000 UTC 27 March (Fig. 3c) when a 3-h pressure rise of more than 6 mb and temperature fall of 20°F (11°C) was observed near 107°W. Six hours later (Fig. 3d) the amplitude of the changes had weakened slightly, and the maximum pressure rise was seen to occur distinctly (approximately 200 km) farther east than the maximum fall of temperature. This behavior appears to be an example of the separation of the eastward-propagating pressure trough from the strong frontal temperature contrast described by Sanders (1999). Shortly thereafter (Fig. 3e) there was a return to low-amplitude oscillations of temperature and pressure.

Thus the entire phenomenon had developed and dissipated in no more than 24 h. This evolution evidently differs significantly from that described by Bjerknes and Solberg (1922), in which a frontal system persists from before the initiation of a wave cyclone through its life history to occlusion, a process that would occur over a period of several days. We suggest that this Norwegian legacy persists in present practice, with a ''significant'' front being regarded as a structure that persists much longer than 24 h. Yet the short-lived system described here produced significant wind damage in Arizona and New Mexico (NOAA 1991) starting at 1600 UTC and continuing through the rest of the day. Severe thunderstorms developed in Oklahoma and Texas around 2300 UTC 26 March and persisted nocturnally until about 1000 UTC of the 27th while the front itself dissipated.

It is seen in Figs. 2 and 3 that the NWS analysis showed a single front persisting through the 24-h period, with a variety of additional ''trofs.'' One of these was evidently associated with the development described above, but it was never carried as a separate front in the NWS analysis.

We have tacitly assumed so far that a rapid change in time means a strong horizontal gradient in space. To check this assumption, we prepared maps of surface potential temperature (as an attempt to compensate for the effects of elevation) and wind. These appear as Fig. 4, covering approximately the western two-thirds of the United States. Initially, at 1200 UTC on the 26th (Fig. 4a) the horizontal gradient of temperature was indeed small over the southwestern United States. A substantial region of moderate and strong gradient, as proposed by Sanders (1999), lay over the Sierra Madre Occidental in Mexico, and numerous other effects of topography can be seen. By 1800 UTC (Fig. 4b) a strong baroclinic zone had formed in southeastern Arizona, apparently a northward extension of the prior Mexican zone. This

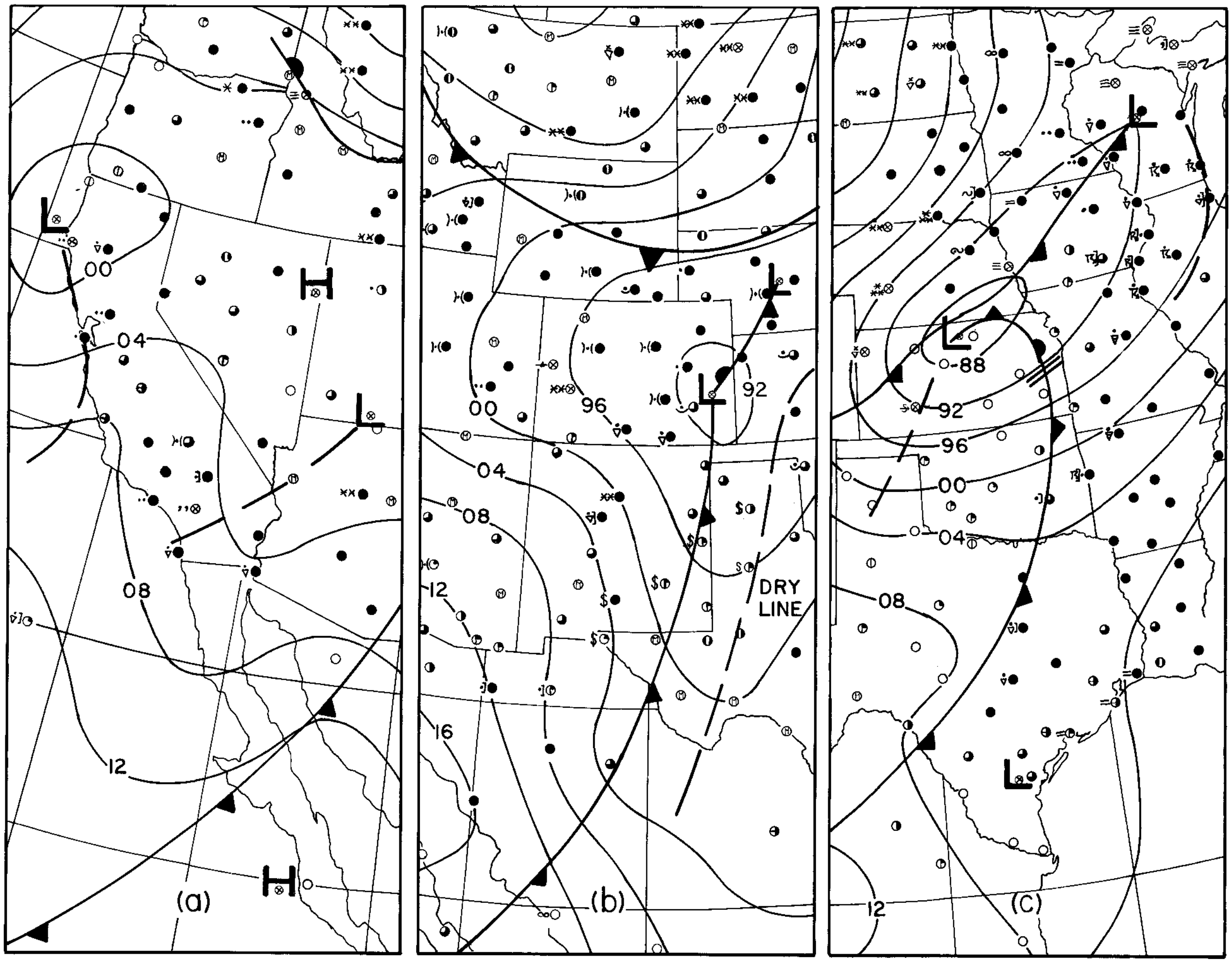


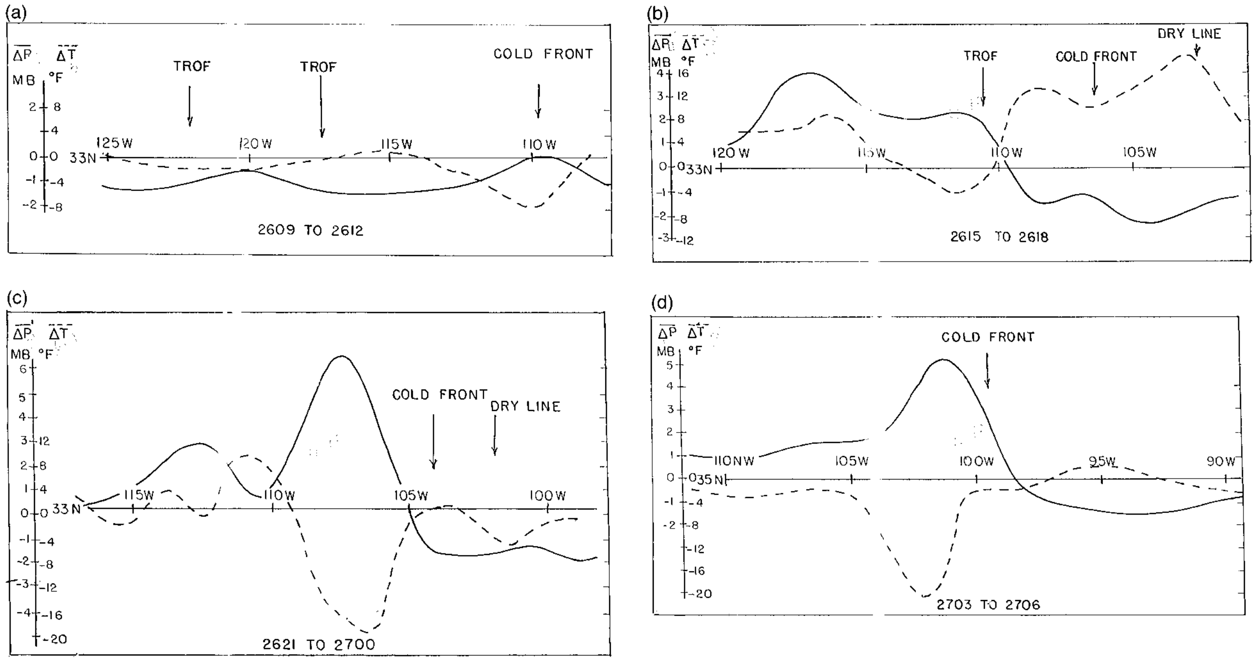
FIG. 2. Excerpts of NWS surface analyses for (a) 1200 UTC 26 Mar 1991, (b) 0000 UTC 27 Mar 1991, and (c) 1200 UTC 27 Mar 1991. Solid lines are isobars of sea level pressure at intervals of 4 mb. Heavy solid lines are analyzed fronts. Heavy dashed lines are analyzed trofs. Letters "L" indicate the position of centers of low pressure. Underlined number adjacent to low is estimated minimum pressure value in mb.

new cold front had reached maximum intensity 6 h later across central New Mexico (Fig. 4c). At 0600 UTC on the 27th (Fig. 4d), the front had lost its temperature gradient as rapidly as it had acquired it, except for a small patch in the Texas Panhandle. Further, the wind shift from southerly to westerly was clearly east of the ridge of maximum temperature, consistent with Fig. 3d. Finally, at 1200 UTC on the 27th (Fig. 4e) temperature gradients were modest throughout the southern plains area, except for the patch in southwest Texas.

The NWS analysis showed a single cold front near the ridge of maximum potential temperature throughout the period. The strong baroclinic zone described above was shown as a trof that finally merged with the cold front to its east. Throughout the period there was another strong baroclinic zone stretching eastward from southern Montana to the eastern edge of the analysis. This zone was appropriately marked by a cold front in the NWS analysis as it moved slowly southward.

4. Mesoscale detail

Further detail was obtained by plotting time series of hourly and special observations from a number of stations in New Mexico, Texas, and Oklahoma. Some of these were assembled into longitude–time sections and are presented in Fig. 5. The longitude span is from about 109°W (GUP, Gallup, New Mexico) to about 98°W (OKC, Oklahoma City, Oklahoma). The time period studied is from 1800 UTC 26 March to 1100 UTC 27 March. The analysis of altimeter setting in this section (Fig. 5a) shows a pressure trough moving eastward rapidly. Between 2000 and 0100 UTC its displacement yielded a speed of about 17.2 m s^{-1}, somewhat stronger than the average westerly component of surface winds over the area. Therefore the trough was propagating eastward through the air in which it was embedded. The average vertical shear of the geostrophic wind in the region indicated a meridional temperature gradient of



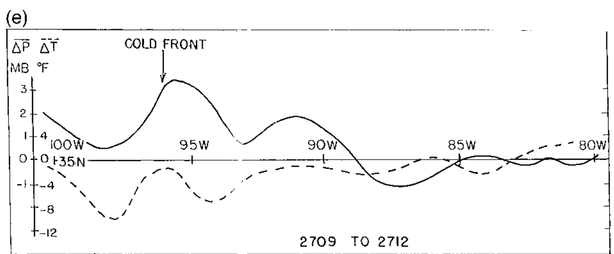


FIG. 3. Sections of change of surface temperature (dashed in °F) and sea level pressure (solid in mb) along 35°N lat, for 3-h periods ending at (a) 1200 UTC and (b) 1800 UTC 26 Mar 1991, and (c) 0000, (d) 0600, and (e) 1200 UTC 27 March 1991. For each section, longitude is indicated every 5° along the abscissa, and scales for changes of temperature and pressure are given along the ordinate. Positions of features of the NWS analysis at the end of the 3-h period are indicated on each section.

about 0.75°C (100 km)$^{-1}$. [According to Sanders (1999), quasigeostrophic theory indicates a somewhat larger propagation speed than that observed, for this meridional gradient and a wavelength of 1500 km.] Values of altimeter setting from AMA (Amarillo, Texas) are somewhat lower than might be obtained by interpolation between its neighbors. They define a weak, stationary, separate trough between 1800 and 2300 UTC, in advance of the mobile feature being discussed. Although the station pressure might contain a slight error, possibly due to the reduction to sea level, it is felt more likely that the observations represent a quasi-stationary lee trough, since the station is in a region of considerable terrain slope on the eastern side of the Rocky Mountains.

The analysis of surface potential temperature in the section (Fig. 5b) shows a strong baroclinic zone, as defined by Sanders (1999), fully formed at the starting time. This zone dissipated beginning after 0100 UTC, becoming a weak remnant after 0700 UTC. While strong, its displacement from 2100 to 0100 UTC represented a speed of 11.8 m s^{-1}, very close to the average speed of the westerly wind component in the area. Thus the band of strong temperature gradient did not propagate but was transported by the wind in its environment, and its separation from the wind shift and pressure trough was due to the propagation of the former. This explanation was proposed by Sanders (1999). The weak lee trough discussed above was distinct from this separation process and was not responsible for it.

5. Frontogenesis and frontolysis

To study the frontogenetical process, we started with the frequently used expression given by Miller (1948). If the x axis is taken along the isotherm with colder air along the positive y axis, the expression is greatly simplified and becomes

$$\frac{D|\nabla\theta|}{Dt} = +\frac{\partial v}{\partial y}\frac{\partial \theta}{\partial y} + \frac{\partial \omega}{\partial y}\frac{\partial \theta}{\partial p} - \frac{\partial Q}{\partial y}.$$

Hoffman (1995) used a variation of this expression, given by Petterssen (1956), in a detailed study of this case. Hoffman's resulting patterns (not shown) are on a relatively small scale but show positive frontogenesis along the analyzed position of the cold front, growing more intense between 1500 and 2100 UTC as the gradient intensifies. This is not surprising, since the temperature gradient appears in the first (confluent) term on

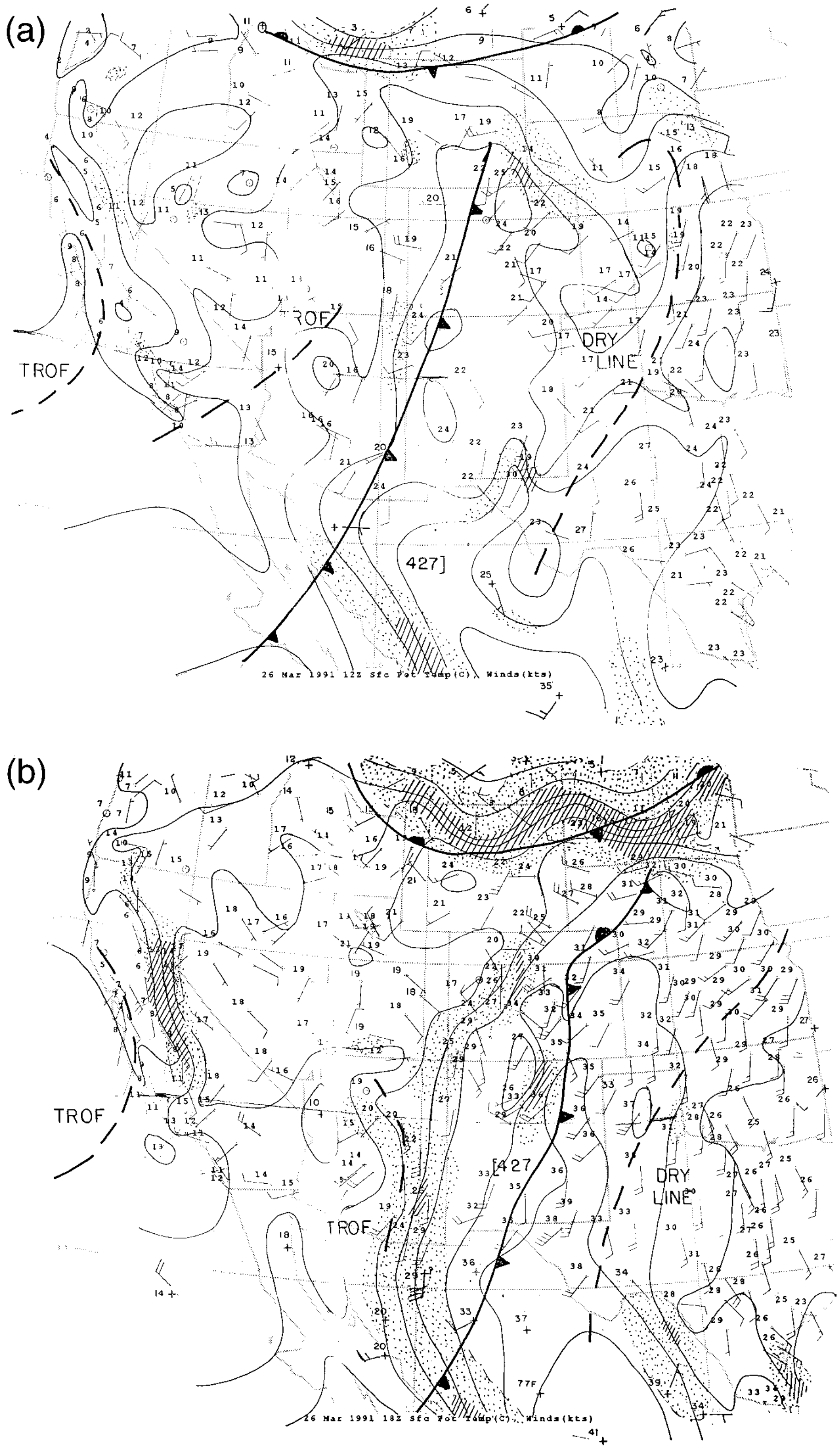


FIG. 4. Maps of surface potential temperature for the western two-thirds of the United States at (a) 1200 and (b) 1800 UTC 26 Mar 1991, and (c) 0000, (d) 0600, and (e) 1200 UTC 27 Mar 1991. The position of the corresponding longitude section in Fig. 3 is indicated by a solid line. Station models show surface wind in conventional form and potential temperature to the upper left of the origin

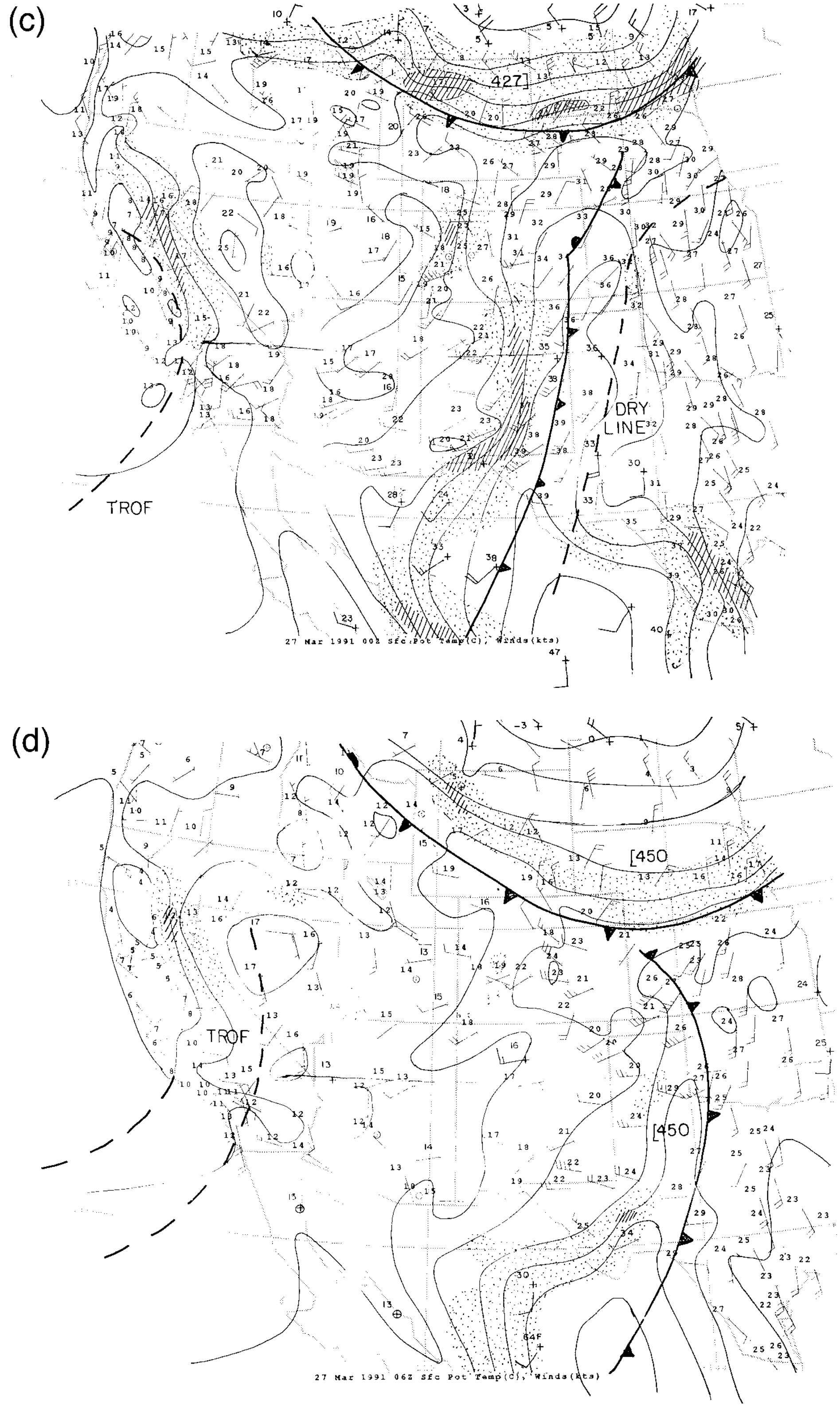


FIG. 4. *(Continued)* of the wind barb. Solid lines are potential isotherms at intervals of 4°C. Stippled areas indicate a contrast of 8°C over a distance of no more than 220 km. Hatched areas indicate the same contrast over no more than 110 km. Features from the NWS surface analyses for the same times are superposed.

(e)

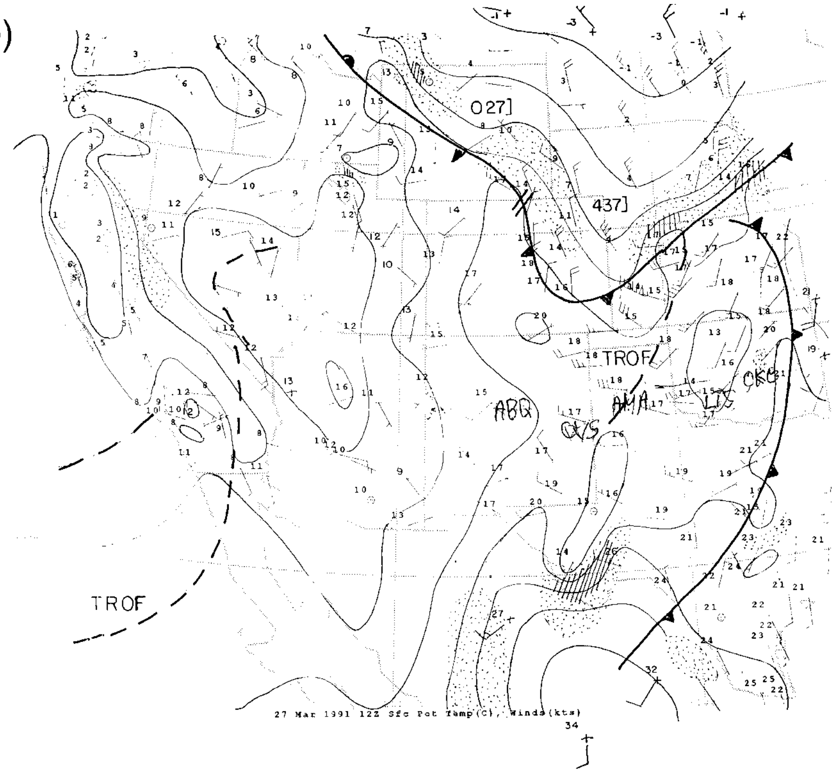


FIG. 4. *(Continued)*

the right-hand side of the above expression. It is difficult to judge what the time integral is, so as to be able to relate the results to the evolution seen on the maps. Further, the time derivative in the frontogenesis expression is a substantial one, not a local one, so that it is necessary to determine the flow of air relative to the frontal zone. This was done for the surface flow by obtaining the zonal component of the wind observations at 19 available stations in New Mexico and western Texas and Oklahoma, relative to the 11.8 m s^{-1} speed of the zone of strong gradient. Observations were used at each station for the time interval encompassing the rapid temperature fall. The zonal component was taken because the orientation of the baroclinic zone at this time was approximately meridional. Data from Guadeloupe Pass, Texas, were excluded because this station is at a high elevation in restricted terrain, reported speeds nearly twice what was observed elsewhere, and was not considered to represent conditions outside a very small region in its vicinity. The calculated components were related to the potential temperature field by interpolating at each station for the value occurring at the time of each multiple of 5°F in the temperature record at that station, from 100° to 75°F. The resulting profile, for this range of values, is shown in Fig. 5b. The relative flow was weakly but consistently toward the frontal zone at its warm edge and also, in a limited region, at its rear edge. This behavior reflects the veering of the wind and the temporary increase in speed as the zone passed. Subsequently the wind weakened as the zone moved away to the east. On average, the relative component was very nearly zero, as noted above. It is not clear how robust the averages discussed above are, because there was considerable variability from station to station. The consistency of the pattern, however, lends credibility to it.

Thus the wind field was convergent in the frontal zone, representing the ageostrophic tendency to collapse toward discontinuity, as analyzed by Hoskins and Bretherton (1972). Since the zone was in an approximately steady state during the period examined (2200–0400 UTC), the inflowing air parcels ahead and behind must have been cooled and heated diabatically, respectively, to maintain a balance. Intense lateral mixing appears to have been the mechanism for these diabatic changes. Reed et al. (1994) found a similar pattern in a mesoscale model simulation of a cold front in an intense oceanic storm. Since from Fig. 5b the estimated distance between the isotherms for 95°F and 85°F (experienced by most of the stations examined) was about 49 km, and the corresponding change in zonal surface wind increased from -1.5 m s^{-1} to $+1.3$ m s^{-1}, the frontal convergence was 5.7×10^{-5} s^{-1}, while the gradient of potential temperature was 11.4°C $(100\ km)^{-1}$, yielding an average rate of frontogenesis by an adiabatic process of 6.5×10^{-9}°C s^{-1}. This value is comparable to those found by Hoffman (1995), but refers to a frontal-scale average in distance and time. There was evidently an

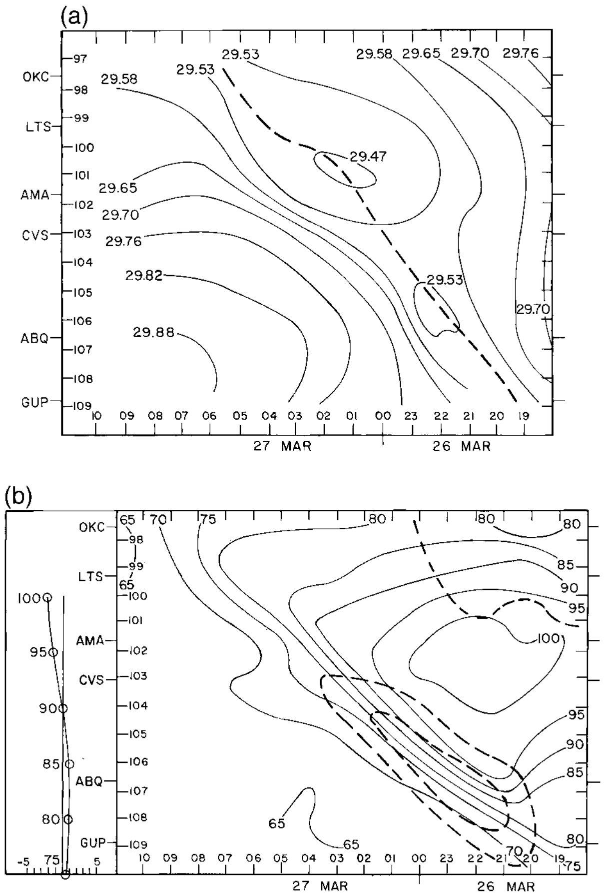


FIG. 5. Time–longitude sections of (a) altimeter setting (inches) and (b) surface potential temperature along 35°N lat for selected surface stations. Station locations are shown on Fig. 4e. Time, from 1800 UTC 26 Mar 1991 to 1100 UTC 27 Mar 1991, is shown along the abscissa. West longitude and station location appear along the ordinate. (a) Solid lines are isopleths of altimeter setting at intervals of 0.059 in. and the dashed line represents the major pressure trough. (b) Solid lines are isotherms at intervals of 5°F. Outer and inner dashed lines indicate regions where a contrast of 15°F occurs over distances of 220 km and 110 km, respectively. The profile near the left edge of (b) shows the front-relative westerly component of surface wind (m s^{-1}), averaged over selected stations at selected times. See text.

approximate balance between this adiabatic frontogenesis and diabatic frontolysis due to mixing within the frontal zone, as noted above. When the frontogenetical effect was no longer present, as the wind shift propagated away from the temperature gradient, the mixing was unopposed and diluted the frontal contrast quickly.

The question remains how the substantial temperature contrast was created initially, say, between 1200 and 1800 UTC 26 March. To study this question, we calculated trajectories from the surface winds and temperatures as mapped for these times and for 1500 UTC, for parcels terminating at stations in eastern Arizona and in southern and central New Mexico. Diabatic temperature changes ranged from -2°C in central Arizona, where there was abundant cloudiness and scattered showers, to $+16$°C in New Mexico, which was characteristically sunny. Evidently the horizontal gradient of surface heat flux was responsible for the creation of the substantial contrast, reinforced by some evaporative cooling in the scattered showers. Similar effects on the modification of cold fronts and the generation of mesoscale circulations have been studied by Segal et al. (1986, 1992, 1993).

6. Concluding summary

We are now in a position to describe the series of events that produced this short-lived cold front and limited its lifetime. During the morning of 26 March a mobile upper-level trough of moderate amplitude moved northeastward from southern California and a strong horizontal contrast of potential temperature developed, between cool air over Arizona and hot air over New Mexico. The role of the upper system was to maintain a damp, weakly ascending, inner mass of air with showery precipitation over Arizona. Thus the solar heating of the ground in this region was limited. On the eastern periphery of this system only thin high clouds were present, providing little impediment to the intense solar heating of the dry ground. The dynamical response to this thermodynamical development was the growth of ageostrophic flow superposed on the weak initial southerlies. This ageostrophic flow was from colder toward warmer air and concentrated the temperature gradient until it was balanced by the dilution due to turbulent mixing. Thus, the frontal zone remained in a nearly steady state for a few hours during the afternoon. It moved eastward at close to 12 m s^{-1}, the average westerly component of the surface winds in which it was embedded. The surface trough and wind shift, however, moved eastward at about 17 m s^{-1}, thus propagating through the flow in which it was embedded. Any cyclone center or trough in the baroclinic westerlies must be expected to behave in this manner, because of the effect of synoptic-scale convergence ahead, and divergence behind, the feature. The consequence of this propagation was the removal of the frontogenetical wind shift from the frontal zone, so that the frontolytical effect of mixing was able to act unopposed. The temperature contrast dissipated from this effect, and from the effect of strong nocturnal cooling of the warmest air, in a few hours. Unstably stratified moist air from the Gulf of Mexico reached the wind shift line in the late evening of the 26th, producing a significant outbreak of severe convection, as the frontal temperature contrast became negligible.

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