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THE FORMATION OF TORNADIC STORMS IN NORTHWESTERN OKLAHOMA ON 2 MAY 1979

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1. INTRODUCTION

During the afternoon of 2 May 1979 severe thunderstorms formed in northwestern Oklahoma, moved east-southeastward, and produced tornadoes (some having multiple vortices), hail greater than 7 cm in diameter, and damaging straight-line winds (Fig. 1). The storms were responsible for two deaths, 25 injuries, and millions of dollars worth of property damage (U.S. Dept. of Commerce, 1979).

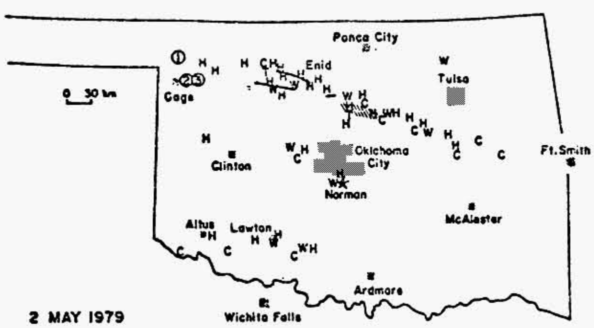

Fig. 1: Severe-weather events in Oklahoma
C - funnel cloud
H - hail
T - tornado (tracks indicated by solid line)
W - damaging wind
hatched area - swath of extensive wind damage
1,2,3 - locations of "first" echoes of storms with mesocyclones

Storm 1 produced a large tornado in Lahoma, OK.
Storm 2 produced a multiple-vortex tornado near Orienta, OK.
Storm 3 produced golfball-size hail at Isabella, OK and wind damage at Fairview, OK.

The purposes of this paper are to describe the sequence of events leading to the formation of the tornadic storms, and to suggest a mechanism by which the storms may have been triggered. Data gathered during Project SESAME-79 (Severe Environmental Storms and Mesoscale Experiment) (Alberty, et al., 1979) were used in our study. In particular, data from the service "A" teletype, the National Severe Storms Laboratory (NSSL) special rawinsonde network and WSR-57 (conventional) radar, high-resolution visible and infrared satellite photographs, and ground-based intercept observations by NSSL and OU (University of Oklahoma) crews were used. Multiple-Doppler analyses of the storms, photogrammetric analysis of the tornado movie taken by OU, and detailed damage surveys will be described elsewhere.

2. THE ENVIRONMENT OF THE STORM

At 0600 CST (all times are in CST) a long-wave trough at 500 mb was situated over southern California and western Arizona. A well defined short-wave trough in the southwesterly current (and associated cumulus convection downstream from the trough axis) extended across Central Oklahoma (Fig.2a). Another short-wave trough was located in New Mexico and Mexico, about 1000 km upstream from the stronger trough. The wave in New Mexico was characterized by strong cyclonic shear and 12-hr height falls of 40-50 m, while the wave in Oklahoma was characterized by both strong cyclonic shear and curvature and height falls of 10-20 m. A region of height rises of 10 m or less, anticyclonic shear and curvature, and a weak thermal ridge was located in between the two waves. The next direct observations of the upper-air wind and temperature fields were not available until well after the storms had begun.

The following conditions which are favorable for the formation of severe thunderstorms were also present at 0600 (Miller, 1972): a moist, low-level jet; a low-level temperature ridge positioned west of a moisture ridge; and a strong, diffluent upper-level jet.

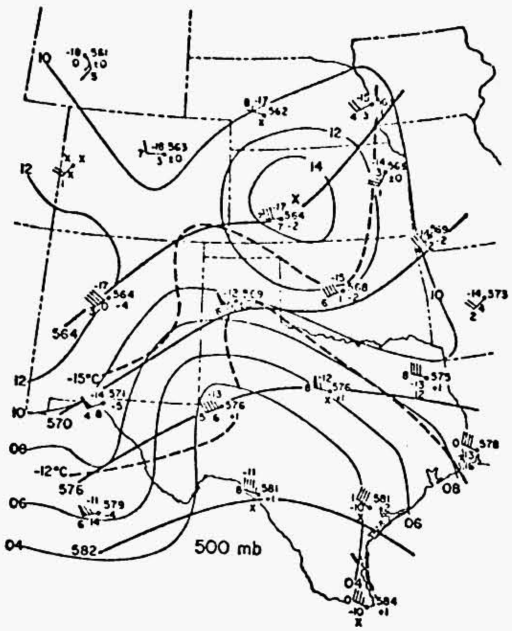

Fig. 2a: 500-mb analysis for 0600 CST 2 May 1979
thick solid lines-heights (dm)
dashed lines-isotherms
thin solid lines-absolute vorticity ($10^{-5}s^{-1}$).
(from LFM initial analysis)

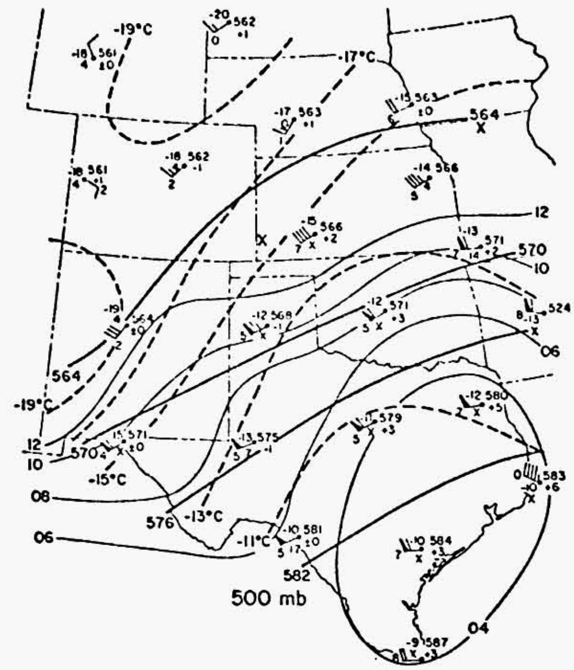

Fig. 2b: as in Fig. 2a, but for 1800 CST

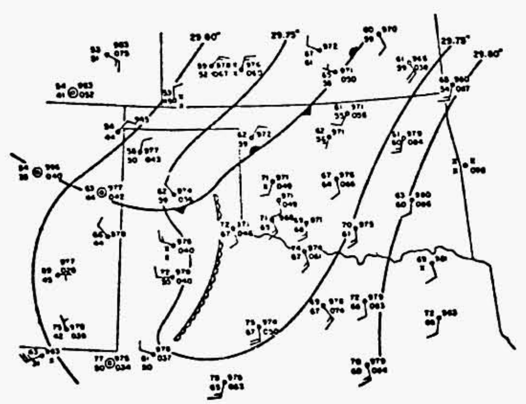

Fig. 3a: surface analysis for 1000 CST 2 May 1979
solid lines-altimeter settings
scalloped line-dryline

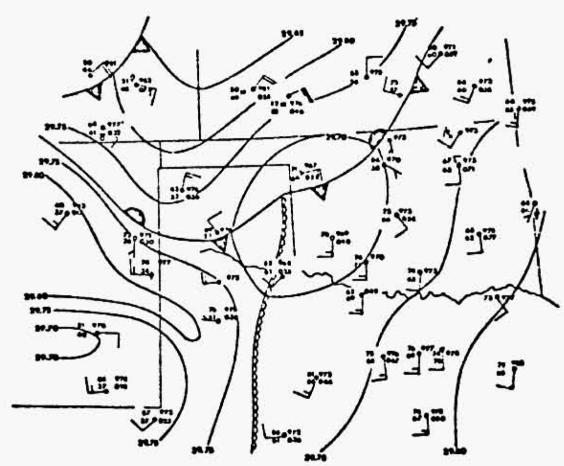

Fig. 3b: as in Fig. 3a, but for 1300 CST

In the morning a surface cold front was entering northwestern Oklahoma and the Texas Panhandle. A weak dryline (Rhea, 1966) in western Texas intersected the cold front just east of Amarillo (Fig. 3a). By early afternoon the front at the surface had become stationary from the Texas Panhandle northeastward through northwestern Oklahoma into south-central Kansas; the weak dryline, oriented in a north-northeast to south-southwest fashion, intersected the front somewhere in the eastern Texas Panhandle or extreme western Oklahoma (Fig.3b). A weak cyclonic circulation was present about the dryline/front intersection. A volume of air, evaporatively cooled during earlier thundershower activity and shielded from insolation by low clouds, covered roughly the northeastern half of Oklahoma and represented a fourth distinct air mass. (The other three were: very cool air north of the front; hot, dry air west of the dryline; and warm, moist air east of the dryline).

The sounding taken at Oklahoma City (within 100 km of the thunderstorm activity) at 1800 was not directly affected by the convective activity to the north, and is therefore probably representative of the storm's environment (Fig. 4). The profiles of temperature, moisture, and wind are characteristic of the "Type I" tornado sounding (Fawbush and Miller, 1954): the troposphere was convectively unstable, with a low-level moist layer capped by an inversion and relatively dry air aloft. Assuming a mean moist-layer mixing ratio of 13.5 g kg^{-1}, we found that the lifted index (Galway, 1956) was -7. The speed of the wind increased and the direction of the wind veered substantially with height. This wind profile has been shown to be associated with long-lived simulated storms having counterrotating vortices (e.g., Klemp and Wilhelmson, 1978).

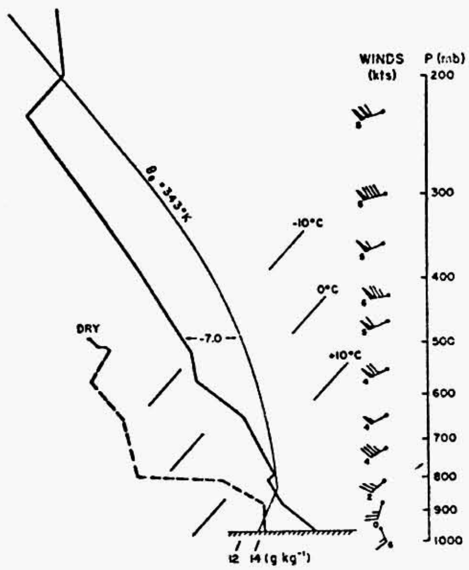

Fig. 4: sounding at OKC 1800 CST, 2 May 1979
θ_e = 343°K is estimate at time and location of thunderstorm activity

The motion of the tornadic storms was to the right of the environmental winds at all levels in the troposphere. Observations indicate that deviate motion can result from propagation along frontal zones or regions of storm-induced boundary-layer convergence (Weaver, 1979). In this case, the storms may have propagated in part by storm-scale induced convergence along inflow-outflow boundaries and in part along the northwest-southeast oriented surface boundary produced early in the morning.

3. THE FORMATION OF THE STORMS

The OU intercept team positioned at Elk City, OK first observed large cumulus towers in the distance to the north at 1336 (Bluestein, 1979). The first radar echo of the tornadic-storm complex appeared at 1342 northeast of Gage (GAG), OK. The 1400 "visible" SMS-2-E satellite photograph shows the storms forming northeast of the dryline/front intersection (Fig. 3b), coincident with an area of relatively low "surface" pressure. Some have noted that this is a preferred area for severe convection (Tegtmeier, 1974). ("Altimeter-setting" observations of pressure were analyzed because they were more plentiful than conventional surface-pressure observations. However, "altimeter settings" are pressures reduced to sea level in a standard atmosphere, and therefore horizontal variations in temperature are not accounted for.) The first severe-weather event (golfball-size hail) was reported about 11 km north of Woodward at 1415.

Throughout the afternoon and evening, severe thunderstorm activity remained very localized. Only three cells in a 100 km long line produced severe weather (Fig. 1). This type of short-lived "local" outbreak occurring along a narrow path (Galway, 1977) is called the "corridor" mode (Moller, 1979), and accounts for roughly 50-60% of all outbreaks in the Southern Plains.

The following have been shown to be severe-storm triggering mechanisms in a convectively unstable atmosphere having strong vertical shear:

1. Synoptic-scale lift downstream from an upper level short-wave vorticity maximum (Fawbush, Miller, and Starrett, 1951; Miller, 1972).

2. Mesoscale lift along a dryline (Schaefer, 1975) or "dryline bulge" (Tegtmeier, 1974);

3. Lift along thunderstorm outflow boundaries (Purdom, 1973);

4. Lift induced by gravity waves (Uccellini, 1975);

5. Lift due to localized regions of buoyancy induced by strong surface heating (Bluestein and Sohl, 1979); and

6. Mesoscale lift associated with frontogenesis (Fawbush, Miller, and Starrett, 1951; Kessinger and Bluestein, 1979).

The importance of each of these sible triggering mechanisms was investigated for the case of 2 May.

The short wave vorticity maximum at 500 mb over New Mexico at 0600 was barely discernible downstream at 1800 CST (Fig. 2b). Since the basic current was about 20-25 m s^{-1} from the southwest, one would expect the wave to have moved northeastward at about the same or slightly less speed. An arc-like configuration of clouds, probably associated with rising motion downstream from the vorticity maximum, moved northeastward from New Mexico into the Texas Panhandle at a speed of about 15-20 m s^{-1}. By 1200 the cloud feature was no longer discernible. Furthermore, in eastern New Mexico and western Texas, the surface wind speed did not increase and the wind direction did not veer. According to quasigeostrophic theory, then, in the absence of strong diabatic heating and temperature advection, the wave must have been very weak. We conclude that the wave must have weakened considerably during the morning, and was probably not strong enough by itself to trigger the severe thunderstorms. However, the extrapolated location of the cloud feature was in the vicinity of the storm at the time it formed, and estimates of the actual vertical-motion field have not been made yet.

Although the winds were 15-20 m s^{-1} from the west-southwest at 700 mb, the winds at the surface on the west side of the dryline remained relatively weak (2-7 m s^{-1}) and there was no evidence of a "dryline bulge" or other localized convergence along the dryline caused by downward transport of momentum west of the dryline (McGinley and Sasaki, 1975).

At 1300, prior to storm development, the SMS satellite photograph did not show any line of cumulus congestus along the surface boundary separating the thunderstorm-cooled air to the northeast from warm air to the southwest. None was observed by the intercept crews either. However, at 1400, after the storms had already formed, a faint northwest-southeast oriented line of enhanced cloudiness extending southeast of the storm complex was visible on the SMS satellite photograph. Since the line appeared as or slightly after the storm had formed, it is difficult to determine what role, if any, the boundary had played in triggering the storm. The line intersected the storm complex to the northeast of where new towers were building. The line may even have been enhanced as a result of storm formation.

Before the storms had formed, wavelike cloud features oriented approximately north-northwest to south-southeast, perhaps visual evidence of gravity-wave motion (Erickson and Whitney, 1973), were apparent on the 1300 SMS-satellite photograph to the east of where the storms later formed. After the storms formed, the wavelike features became very pronounced and assumed a northwest-southeast orientation. It is not known if these clouds were indicative of gravity waves travelling along the low-level inversion perpendicular to the flow. The surface-pressure traces from Vance AFB (near Enid, OK) shows no evidence of well-defined gravity wave fluctuations. If there were gravity waves aloft, it is unclear whether or not they were strong enough to trigger the storms. It is also possible that the storms themselves somehow enhanced or excited gravity-wave motion along the low level inversion.

It is unlikely that the storm formed only as the result of local heating because heating was strong in other places where storms did not form.

Analyses of the frontogentical function (Petterssen, 1956) based on subjective streamline, isotach, and temperature analyses are shown in Figs. 5 and 6 for 1000 and 1300, i.e., roughly 3½ and ½ hours before the storms formed. The two dimensional frontogenetical function is a measure of the individual rate of change of the horizontal temperature gradient following horizontal motion. It is defined as

$$F = \frac{1}{2}\,|\vec{\nabla}_h\theta|\,(D\cos 2\beta - \delta),$$

where $|\vec{\nabla}_h\theta|$ is the magnitude of the horizontal gradient of potential temperature, D is the resultant horizontal deformation, β is the angle between the axis of dilatation and the isentropes, and δ is the horizontal divergence. For a given static stability, the strength of the vertical circulation is a function of F (in the absence of friction and diabatic heating) (Sawyer, 1956). As deformation increases the horizontal temperature gradient, a thermally-direct vertical ageostrophic circulation forms which acts to increase the thermal wind shear. Convergence at the surface under the rising branch of the circulation further intensifies the temperature gradient.

At 1000 there is an elongated zone of relatively high values of F (in excess of 5×10^{-10} deg K $m^{-1}s^{-1}$) east-southeast of the surface front. The parts of F due to deformation and convergence are the same order of magnitude, $2\text{-}3 \times 10^{-10}$ deg K m^{-1} s^{-1}. At 1300 there is a local, intense maximum in excess of 17×10^{-10} deg k m^{-1} s^{-1} in northwest Oklahoma and the northeast Texas Panhandle. The frontogenetical function more than doubled and became more concentrated locally in just three hours. The first echo appeared just northeast of GAG at 1342 i.e., in the vicinity of the maximum in F at 1300. An elongated zone of large values of F ($6\text{-}9 \times 10^{-10}$ deg K m^{-1} s^{-1}) due to deformation paralleled the front. A classic deformation field, whose axis of dilatation was approximately parallel to

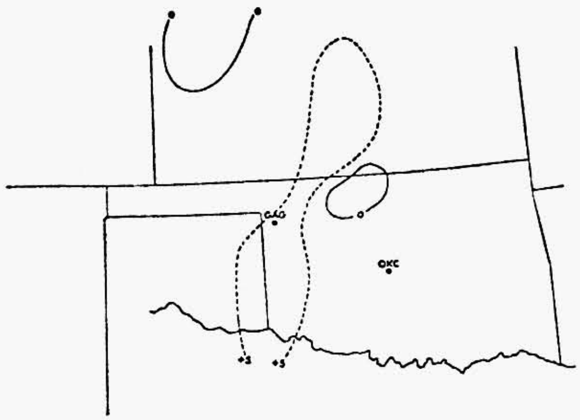

Fig. 5: frontogenetical function at surface 1000 CST, 2 May 1979 (10^{-10}°K $m^{-1}s^{-1}$)

the isotherms, was located northeast of the surface low. A localized area of large values of F(10×10^{-10} deg K m^{-1} s^{-1}) due to convergence was situated northeast of the surface low.

Differential diabatic heating also contributed to frontogenesis: North of the front (in southwestern Kansas) the low-level air was cooled by evaporation of drizzle, while south of the front (in southwestern Oklahoma) skies were clear and the air was being heated.

The vertical circulation associated with frontogenesis is therefore a likely candidate for triggering the storms.

4. SUMMARY AND CONCLUSIONS

Tornadic thunderstorms formed over a small area in northwestern Oklahoma, northeast of a weak surface low along the intersection of a weak dryline and a stationary front. Evidence was presented that the rate of change of horizontal temperature gradient due to deformation and convergence had increased substantially in a three-hour period beginning 3½ hours prior to storm development and that the frontogenetical-function field had a well defined local maximum in the vicinity of storm formation 30 minutes <u>before</u> storm formation. It is hypothesized, therefore, that the vertical circulation associated with frontogenesis triggered the storms.

It is not clear what role, if any, a weakening shortwave vorticity maximum aloft played in promoting frontogenesis. Furthermore, the role of gravity waves propagating along an inversion east of the storm over a cool, moist, stable layer has not been determined. The thunderstorm-cooled air boundary may have been important mainly in creating a localized region of low-level convergence along the front; there may not have been significant convergence along the boundary if there hadn't been a front.

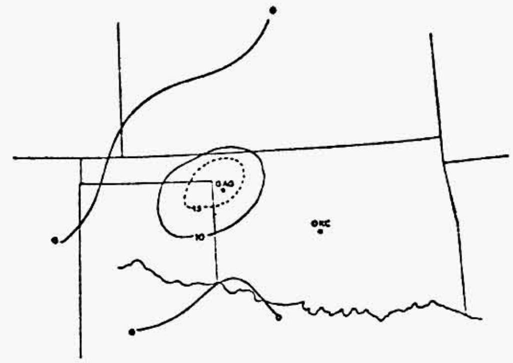

Fig. 6a: as in Fig. 5, but for 1300 CST

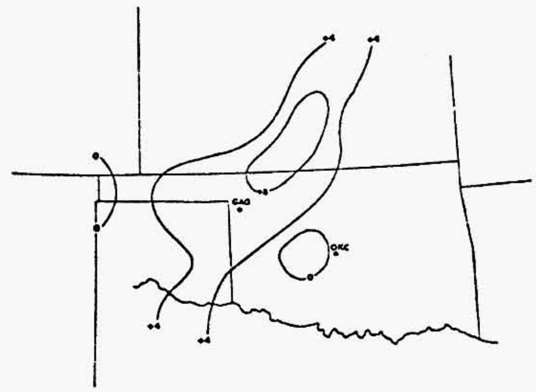

Fig. 6b: as in Fig. 6a, but for deformation only

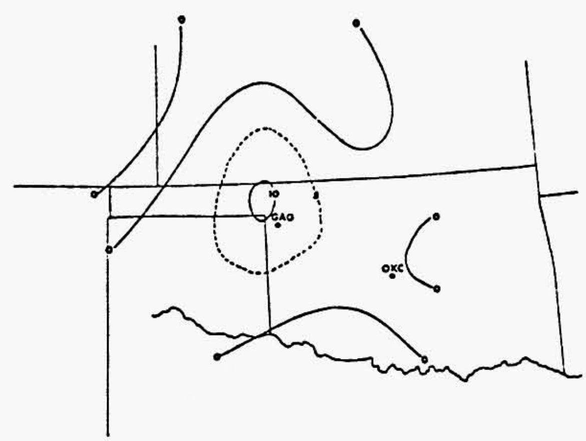

Fig. 6c: as in Fig. 6a, but for convergence only

A case study of this storm complex is continuing. Some questions which need to be answered are as follows: What is the role of the nondivergent and divergent components of frontogenesis? Why did the weak surface low form? Why is the frontogenetical function a maximum northeast of the low? Why were the tornadic storms so localized? We hope that further analysis of the SESAME data will provide answers to these questions.

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