The Distinction between Large-Scale and Mesoscale Contribution to Severe Convection: A Case Study Example

CHARLES A. DOSWELL III*

NOAA, Environmental Research Laboratories—Weather Research Program, Boulder, CO 80303

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ABSTRACT

Using a case study of a relatively modest severe weather event as an example, a framework for understanding the large-scale–mesoscale interaction is developed and discussed. Large-scale processes are limited, by definition, to those which are quasi-geostrophic. Mesoscale processes are defined to be those which are linked in essence to processes occurring on both larger and smaller scales. It is proposed that convective systems depend primarily on large-scale processes for developing a suitable thermodynamic structure, while mesoscale processes act mainly to initiate convection. The case study is presented not as a "typical" event in its particulars, but rather to suggest the complex ways in which large-scale and mesoscale processes can interact. Implications for forecasting are an essential part of the discussion, since mesoscale systems are so difficult to predict with the present knowledge and technology available in operations.

1. Introduction

Late in the day of 6 May 1983 a tornadic severe thunderstorm struck Topeka, Kansas. This storm resulted in one fatality, 25 injuries, and considerable property damage. Rated an F3 tornado (see Fujita and Pearson, 1973, for a definition of the F-scale), its track paralleled that of the infamous 8 June 1966 storm (Galway, 1966), according to the *Storm Data* report, but was considerably smaller and less intense.

Although this storm had substantial impact on Topeka, this storm day would not be characterized as one with notably widespread and intense convection. Those rare days with widespread, extremely violent severe storms are often chosen for detailed study (e.g., Miller, 1972), while the more frequent but less dramatic days usually remain unexamined. I believe that this leads to a distorted picture of not only the large-scale setting conducive to severe convection (see Maddox and Doswell, 1982a,b), but also of the significance of mesoscale processes prior to (and during) severe weather.

A conceptual framework for distinguishing between large-scale processes and those operating on the mesoscale has not been established. The usual method of defining scales is to use order of magnitude arguments (see e.g., Orlanski, 1975, or Fujita, 1981), but this approach can be rather arbitrary with regard to the location of the interfaces between scale categories (e.g., where "large scale" ends and "mesoscale" begins). Emanuel (1980) proposed some physically motivated arguments based on dominant force balances. Unfortunately, mesoscale processes are difficult to characterize in such terms, because it is not clear what, if any, force balances dominate on the mesoscale.

This paper examines the 6 May 1983 case in order to understand what processes were important in determining if, when and where severe thunderstorms would develop. As discussed in Doswell (1984), a plausible scenario can be proposed for the events on this day which involves mesoscale processes that are inherently difficult, if not impossible, to predict. With this example in mind, some ideas are offered in order to begin defining and understanding the scale interaction problem—at least relating to forecasting severe convection.

This case study is not being presented as rigorous proof for a testable hypothesis. In the sense described in Hooke (1963), one never truly proves hypotheses, nor does a sample of one provide convincing evidence in support of a hypothesis. Rather, the events of 6 May 1983 serve to illustrate how important (and how subtle) the scale interaction issue can be in any given situation, as well as how challenging it can be to attempt to understand a given event, even after the fact.

2. The case of 6 May 1983

a. Large-scale setting

This case has been discussed in Doswell (1984, 1987) hereafter referred to as D84 and D87, respectively, so only the essential elements will be considered here. The situation is characterized by large-scale cyclogenesis, as indicated by the 850 and 500 mb charts on the morning of 6 May, shown in Figs. 1 and 2. These reveal a negatively tilted, short wave trough entering the central plains states, with an associated surface low in northwestern Kansas (Fig. 3). Considerable difluence

* Present affiliation: National Severe Storms Laboratory, Norman, OK 73019.

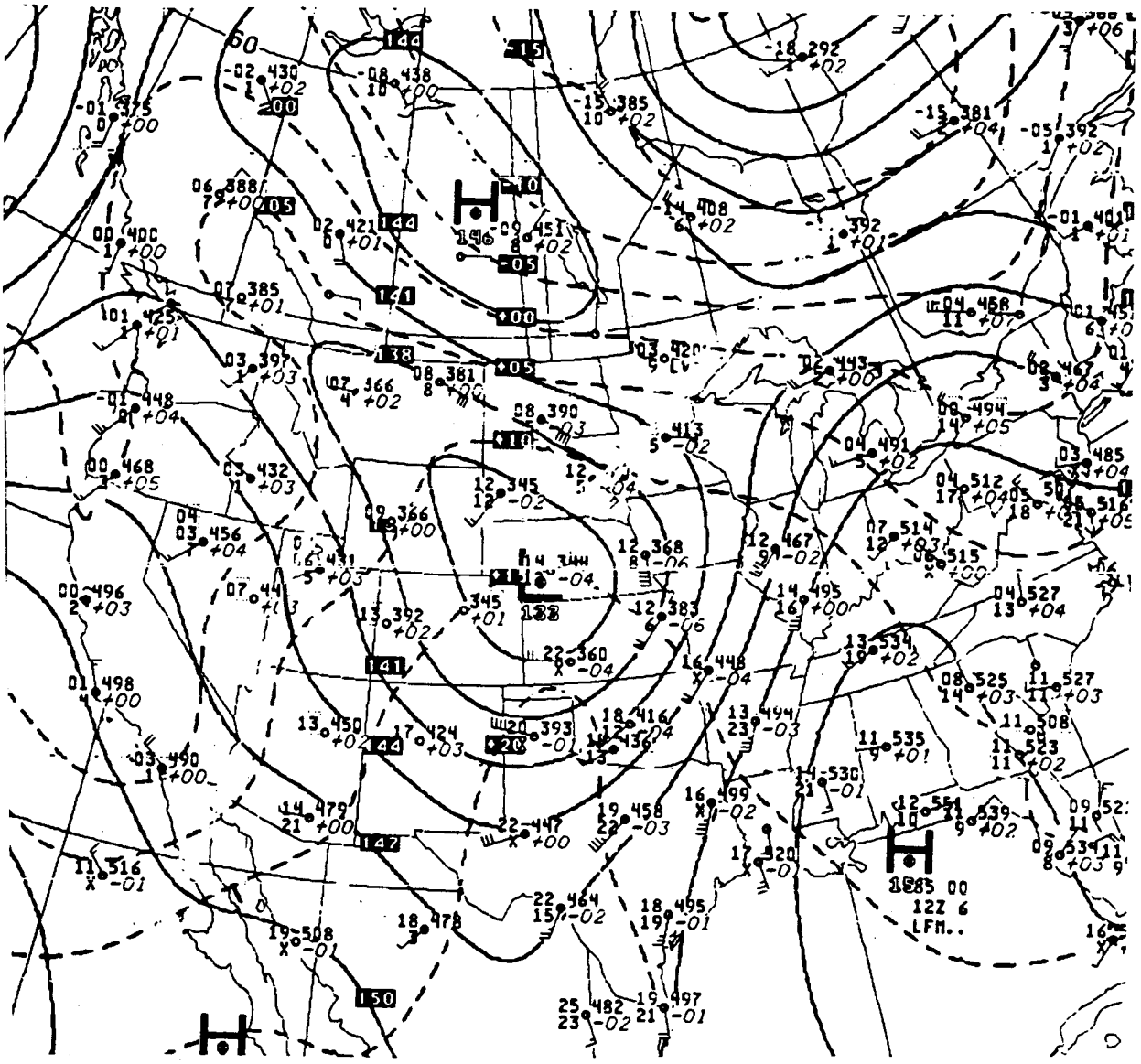

FIG. 1. National Meteorological Center (NMC) 850 mb analysis at 1200 UTC on 6 May 1983.

(at, say, 300 mb; not shown) is associated with the exit region of the upper level jet stream accompanying this approaching short wave trough, and a substantial low-level jet stream is present. (See Uccellini, 1980, for a discussion of the connections between upper and lower jet streams.) A complex pattern of surface features is evident, with a dryline in central Kansas extending across northwestern Oklahoma and into western Texas. Also of considerable interest is the developing cold front in western Kansas. In this situation a forecaster is likely to anticipate (subjectively) the strengthening of this cyclone.

While the lapse rate field (Doswell et al., 1985) reveals low values of dry static stability in eastern Kansas and southwestern Missouri (Fig. 4), there are two major obstacles to convective development in the Topeka area: a strong capping inversion (as seen in Fig. 5) and relatively modest low-level moisture (i.e., dewpoints $\leq 50°F$). A key issue for forecasters is whether or not these negative factors can be overcome as the day evolves. A forecaster might be concerned that the strong low-level jet stream flow would import more low-level moisture, and/or that the vertical motion associated with the cyclone, combined with daytime heating, could weaken the capping inversion enough to permit deep convection.

The predictions by the National Meteorological Center's Limited-Area, Fine-Mesh Model (LFM) using the morning data (Fig. 6) are consistent with the anticipated large-scale evolution of the cyclone, based on

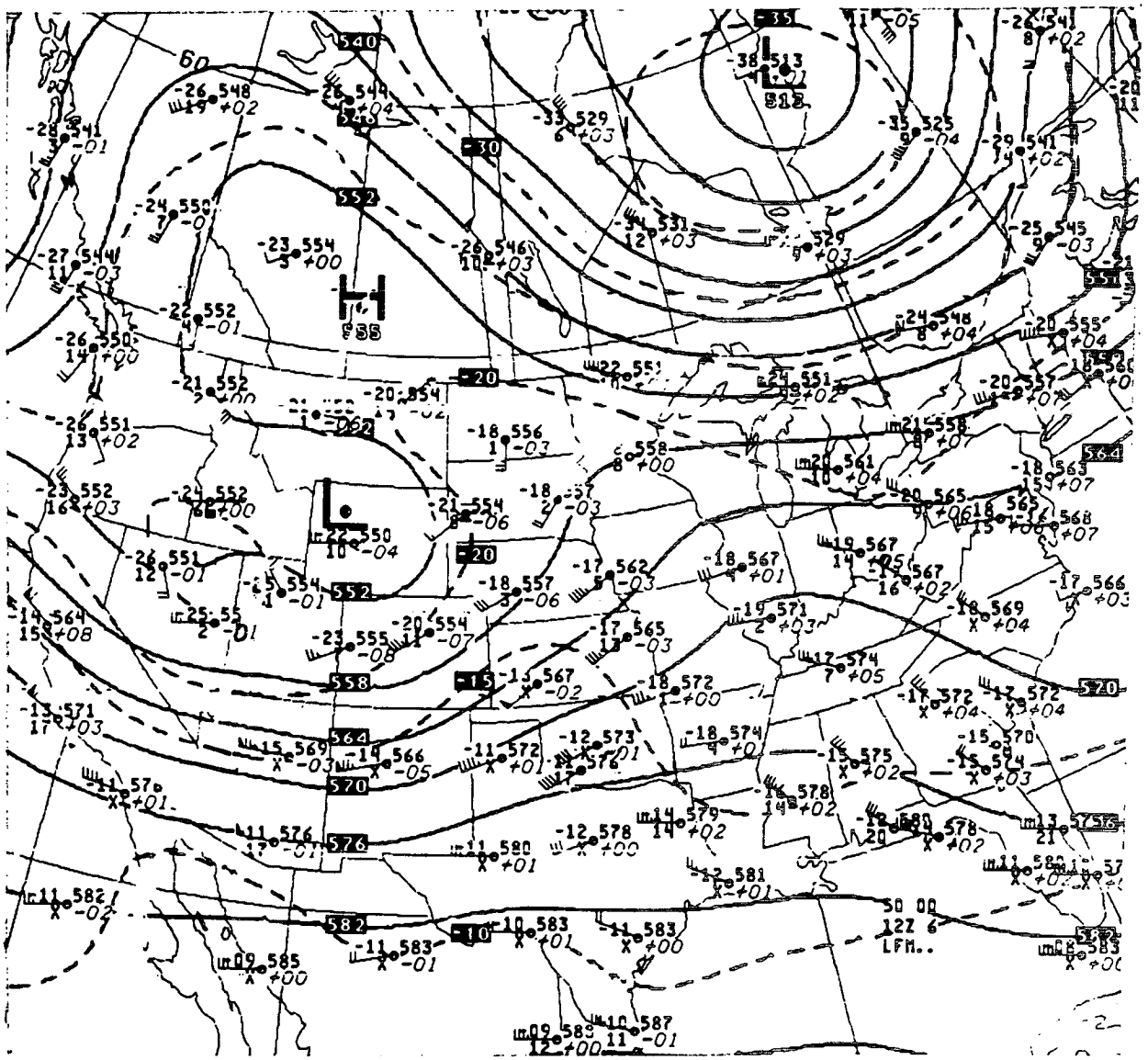

FIG. 2. As in Fig. 1 except 500 mb analysis.

this diagnosis of the morning data. This forecast by the LFM certainly fits the overall structure in the verifying analyses rather well, as exemplified in Fig. 7. However, some significant details are not well treated by the LFM, specifically the strength and movement of the cold front. While the failure of the LFM to treat this sort of detail is not altogether unexpected, the origins and movement of this front are important issues to be discussed herein.

b. Subsynoptic features

By 2130 UTC, there is strong convection (producing severe weather) over much of the eastern two-thirds of Nebraska and some isolated thunderstorms in western Kansas, but little more than towering cumulus in east central Kansas (Fig. 8). In spite of an intruding dryline (Fig. 9) and a modest increase of low-level moisture, it appears that the capping inversion has restrained the release of the convective potential revealed in Fig. 5. As noted in D84 and D87, by 2100 UTC the front (originally in western Kansas at 1200 UTC) has become quite intense and is approaching the dryline from the west. Figure 10 shows that a region of large surface pressure rises has developed behind this boundary and these appear to be driving the front rapidly southeastward. By 0000 UTC, this front has intersected the dryline and an intense squall line has developed rapidly on its leading edge (Fig. 11). This squall line includes the tornadic storm at Topeka.

While it is not possible to know unambiguously the

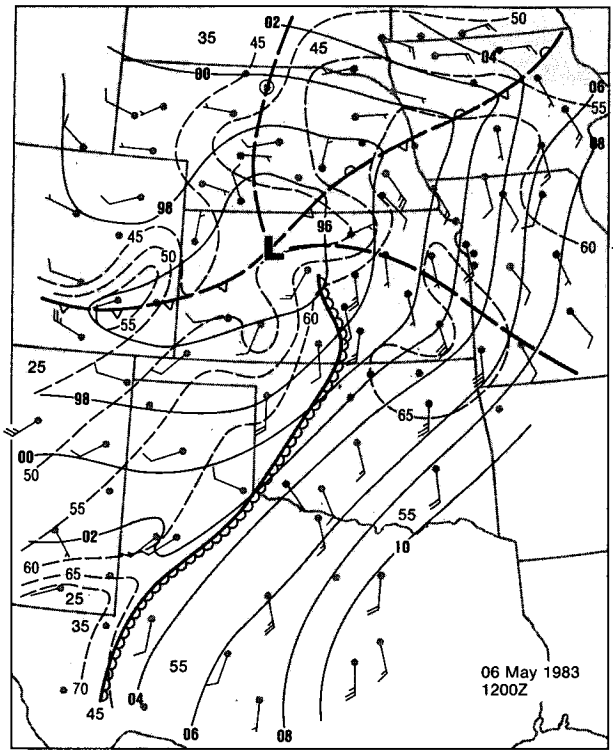

FIG. 3. Subjective surface analysis at 1200 UTC 6 May 1983. Isobars (solid lines) are at 2 mb intervals, isotherms (dashed lines) and isodrosotherms (outlines of stippled areas) are at 5°F intervals. Frontal symbols and wind barbs are conventional.

importance of this front in the initiation of the Kansas convection, it appears as something other than coincidental that deep convection began with its arrival. Of course, one cannot go back and see what would have happened had the front not intensified, so it is possible that thunderstorms would have developed along the dryline without it. I am inclined to believe that this front was indeed essential to this situation. Thus, the origins of this front may be quite important, and we shall see that the intensification of this front cannot be explained on the basis of large-scale processes, in the sense of the term "large scale" as defined in the next section of this paper.

A forecaster monitoring the hourly surface data may have noted the intensification and southeastward advance of the front behind the dryline. However, several important questions are left unanswered concerning this situation. How intense was the lifting associated with the front? Would this lift be sufficient to initiate the convection which seemed to be inhibited by the capping inversion? Would the early indications (e.g., at 1800 UTC—not shown) of important changes in this front be sustained as the front advanced? It is not clear that quantitative evaluation of, e.g., moisture convergence at the surface is sufficient to address these questions, since meaningful answers require knowledge of the vertical structure throughout the area in question (e.g., the strength and time tendency of the capping inversion, as well as the depth of the convergence along

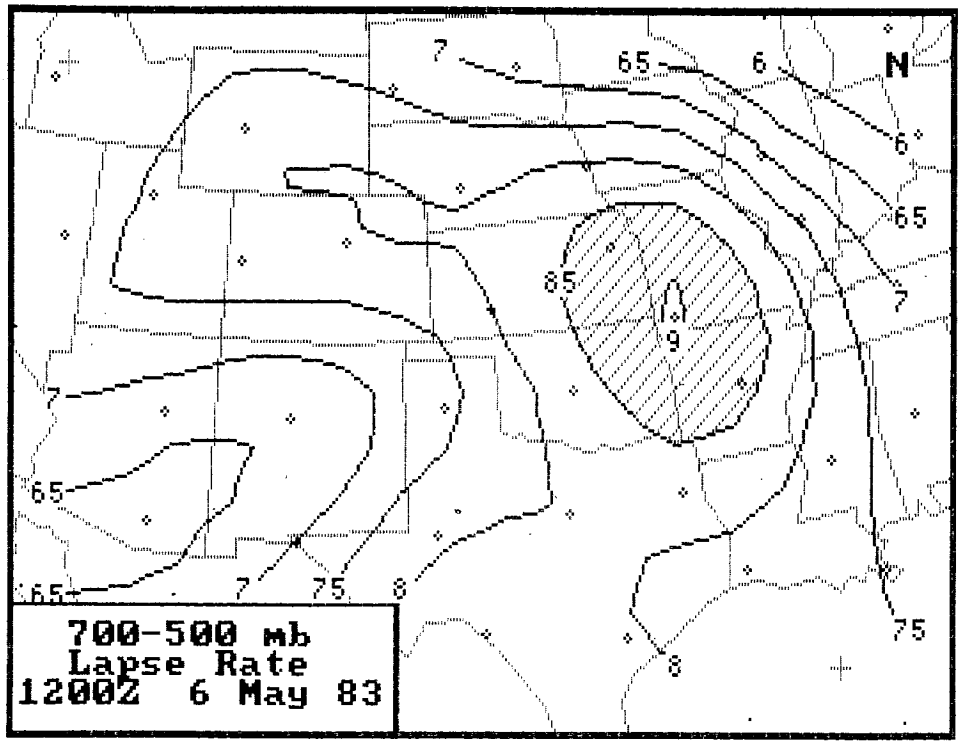

FIG. 4. 700–500 mb lapse rate, contoured at intervals of 0.5°C km^{-1}, at 1200 UTC on 6 May 1983, with the decimal point suppressed. The region with values exceeding 8.5°C km^{-1} is hatched.

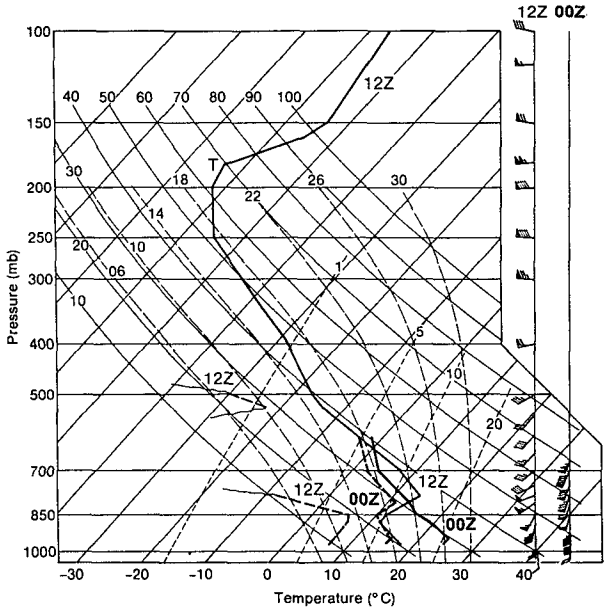

FIG. 5. Topeka, Kansas soundings plotted on a skew T–log p diagram. Plotting is conventional, with thin lines showing the sounding at 1200 UTC 6 May 1983, while the thick lines are for the sounding at 0000 UTC 7 May 1983, which was released just ahead of the approaching squall line and terminated prematurely as a result.

the frontal zone). The data available in operations simply do not permit acquisition of such knowledge. Further, a forecaster might not recognize the significance of the developments associated with the front until convection was underway.

3. Defining the roles of large-scale and mesoscale processes

Reducing the problem of forecasting deep, moist convection to its essence (e.g., McNulty, 1978), requires a triad of ingredients: moisture, conditional instability (e.g., as measured by lapse rates), and a source of lift. Remove any one of these and there well may be some important weather phenomena, but the process is no longer deep, moist convection. Other factors, such as the vertical wind profile, may influence the *type* of convection which forms. However, the presence or absence of deep, moist convection of any sort depends only on these three factors. The requirements for moisture and conditionally unstable lapse rates can be understood in terms of convective available potential energy, or CAPE (see Moncrieff and Miller, 1976), but the question of lift deserves special attention.

Lift is needed to initiate the convection, because the atmosphere is never sufficiently unstable that convective overturning can begin spontaneously (i.e., superadiabatic lapse rates), except in extremely shallow layers, usually near the surface. Substantial negative buoyancy must often be overcome before a rising parcel reaches its level of free convection (LFC). As noted in Doswell et al. (1985), the origins of this required lift are not likely to be found in large-scale ascent. The magnitude of large-scale vertical motion (on the order of a few centimeters per second) is simply too small to accomplish the needed lift in a reasonable time. Further, if large-scale lifting were the main initiating mechanism, deep convection would be expected to begin as relatively extensive cloudiness before breaking down into individual convective elements. This is contrary to observations in which convective storms normally begin as isolated elements. Perhaps one exception to this rule might be an "overrunning" situation, in which convection develops in a region of extensive cloudiness far into the cold air behind a front. The large-scale lift in such a case may well play some role in initiation, as well as modifying the local thermodynamic environment. Disregarding this rather special possible counterexample, one can conclude that the lift needed to start deep convection is generally a product of mesoscale processes.

The question then arises as how to distinguish physically the mesoscale from the large-scale process. My proposal is consistent with Emanuel (1980) in restricting large-scale processes to those which are adiabatic, hydrostatic, and in which mass continuity is satisfied, advection is dominated by the geostrophic part of the wind, and the variation of Coriolis parameter (f) is unimportant. In short, I view midlatitude, large-scale processes to be those which are *quasi-geostrophic* in a textbook sense. This distinction also excludes the special dynamics of long waves and the general circulation, which are dominated by physical mechanisms that are not well characterized by quasi-geostrophy. By assuming an equivalence between the terms "large scale" and "quasi-geostrophic," I do not imply that only quasi-geostrophic processes operate within a given order of magnitude range of length (and/or time) scales. To do so would not only be meteorologically incorrect but would leave us with the same dilemma of arbitrary scale categories we had before.

Instead, the objective of such a definition is to capture the essence of the processes which govern midlatitude extratropical weather systems. Many textbook explanations of quasi-geostrophic dynamics are available (see Holton, 1979 and Dutton, 1976), giving far more details than are offered here. What is truly intriguing about quasi-geostrophic theory is that something so simple contains the essence of the dynamics of extratropical weather systems. Using the textbook assumptions, one derives the standard quasi-geostrophic forecasting system:

$$\sigma\nabla^2\omega + f_0^2\frac{\partial^2\omega}{\partial p^2} = f_0\frac{\partial}{\partial p}[\mathbf{V}_g\cdot\nabla(\zeta_g+f)] + \nabla^2\left[-\mathbf{V}_g\cdot\nabla\left(\frac{\partial\Phi}{\partial p}\right)\right], \quad (1a)$$

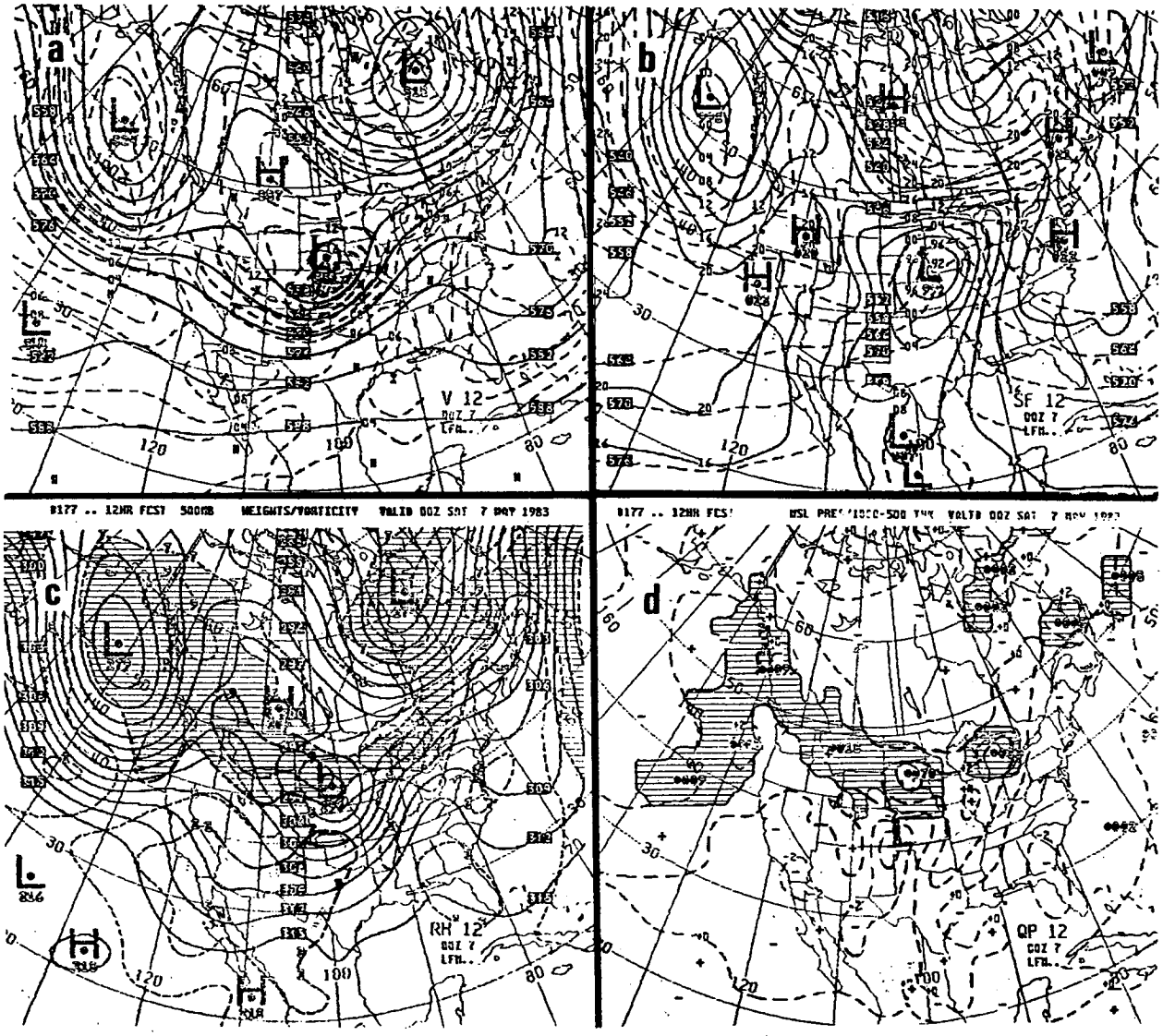

FIG. 6. Conventional four-panel LFM numerical predictions based on the 1200 UTC rawinsonde data, valid at 0000 UTC on 7 May 1983; (a) 500 mb heights and vorticity, (b) mean sea level pressure and 1000–500 mb thickness, (c) 700 mb heights and mean (surface to 500 mb) relative humidity, and (d) accumulated precipitation and 700 mb vertical velocity.

$$\sigma\nabla^2\chi + f_0^2\frac{\partial^2\chi}{\partial p^2} = -\sigma f_0 \mathbf{V}_g\cdot\nabla(\zeta_g+f) + f_0^2\frac{\partial}{\partial p}\left[-\mathbf{V}_g\cdot\nabla\left(\frac{\partial\Phi}{\partial p}\right)\right], \quad (1b)$$

where the notation is as in Holton (1979)[1]. The so-called omega equation (1a) is a diagnostic equation for the vertical motion ($\omega \equiv dp/dt$), while (1b) is a diagnostic equation for the height tendency ($\chi \equiv \partial\Phi/\partial t$).

The omega equation is typically of most concern in physical discussions of this system, since one must endeavor to understand the origins of vertical motion when the horizontal winds are essentially geostrophic. I shall not repeat the textbook discussions of this important topic, but I wish to draw attention to the approach of Hoskins et al. (1978) and Hoskins and Pedder (1980), in which the forcing term on the right-hand side of (1a) is cast in terms of the so-called **Q**-vector divergence, viz,

$$\sigma\nabla^2\omega + f_0^2\frac{\partial^2\omega}{\partial p^2} = -2\nabla\cdot\mathbf{Q} \quad (2)$$

where the **Q**-vector is defined by

[1] An important auxiliary assumption used to derive (1a, b) is that the dry static stability (σ) is at most a function of pressure p. This standard assumption has important implications, to be explored in a later publication.

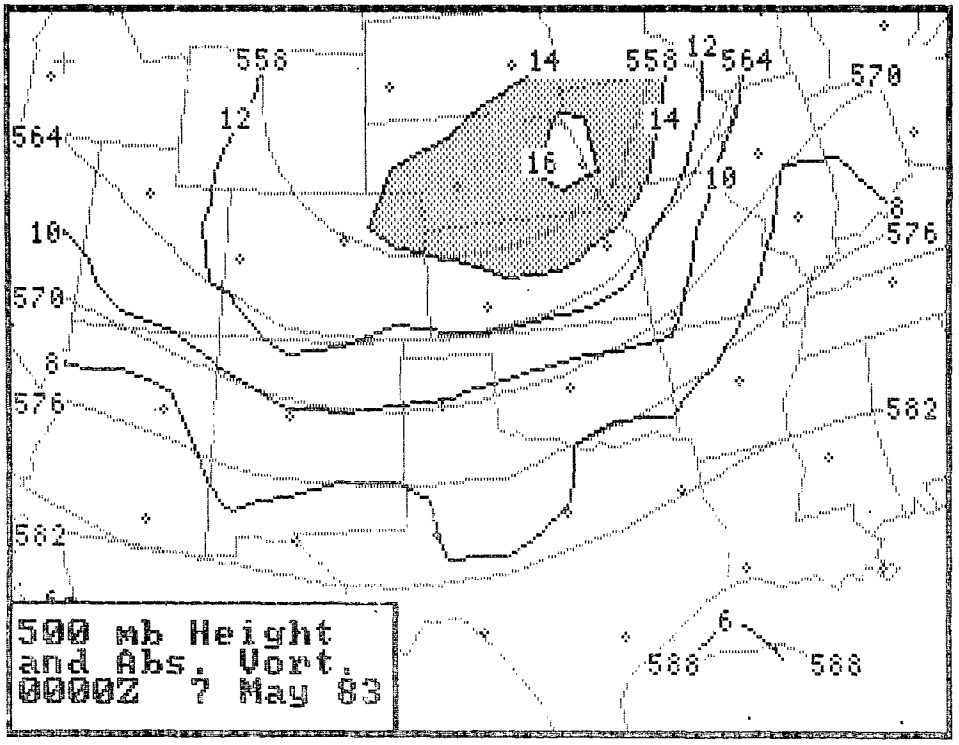

FIG. 7. The 500 mb geopotential height (light contours, at an interval of 6 dam) and geostrophic absolute vorticity (heavy contours, at an interval of 2×10^{-5} s^{-1}) at 0000 UTC 7 May 1983. This should be compared to the 500 mb height/vorticity panel of Fig. 6a.

$$\mathbf{Q} \equiv \left[\frac{\partial \mathbf{V}_g}{\partial x} \cdot \nabla\left(\frac{\partial \Phi}{\partial p}\right), \frac{\partial \mathbf{V}_g}{\partial y} \cdot \nabla\left(\frac{\partial \Phi}{\partial p}\right)\right]. \quad (3)$$

While the **Q**-vectors are less easily interpretable in physical terms than the forcing terms in the more tra-

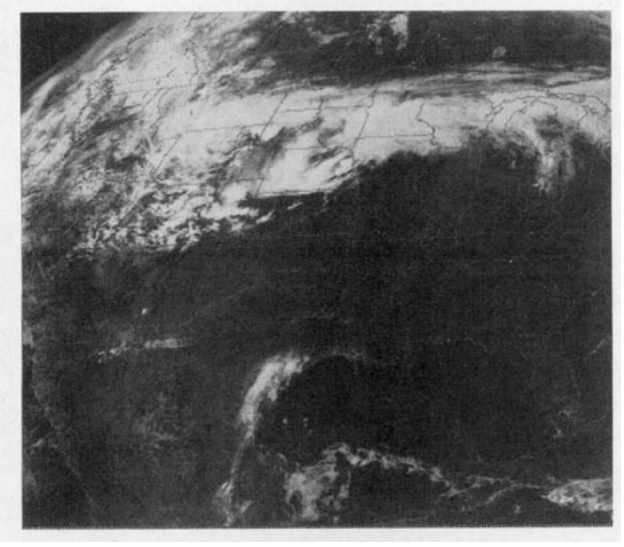

FIG. 8. Visible satellite image at the indicated date and time (UTC). Note the thunderstorms along the Nebraska–Kansas border and the cumulus congestus in northeastern Kansas.

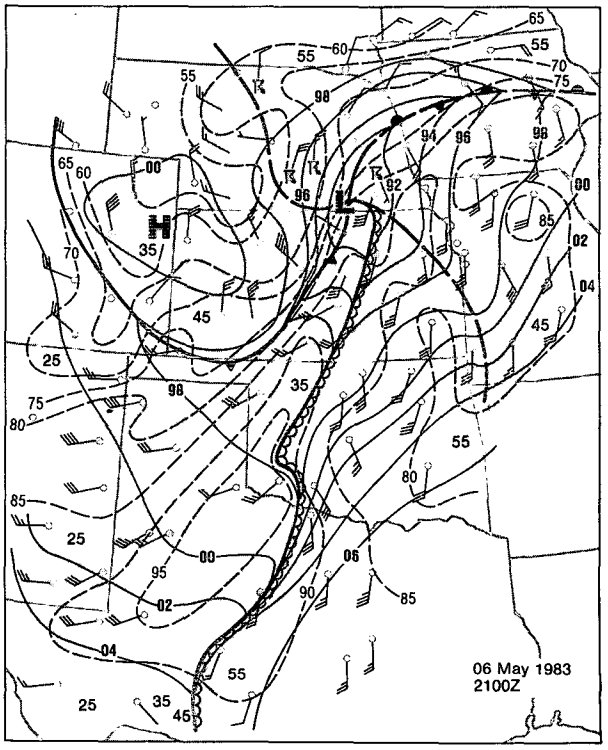

FIG. 9. As in Fig. 3 except at 2100 UTC.

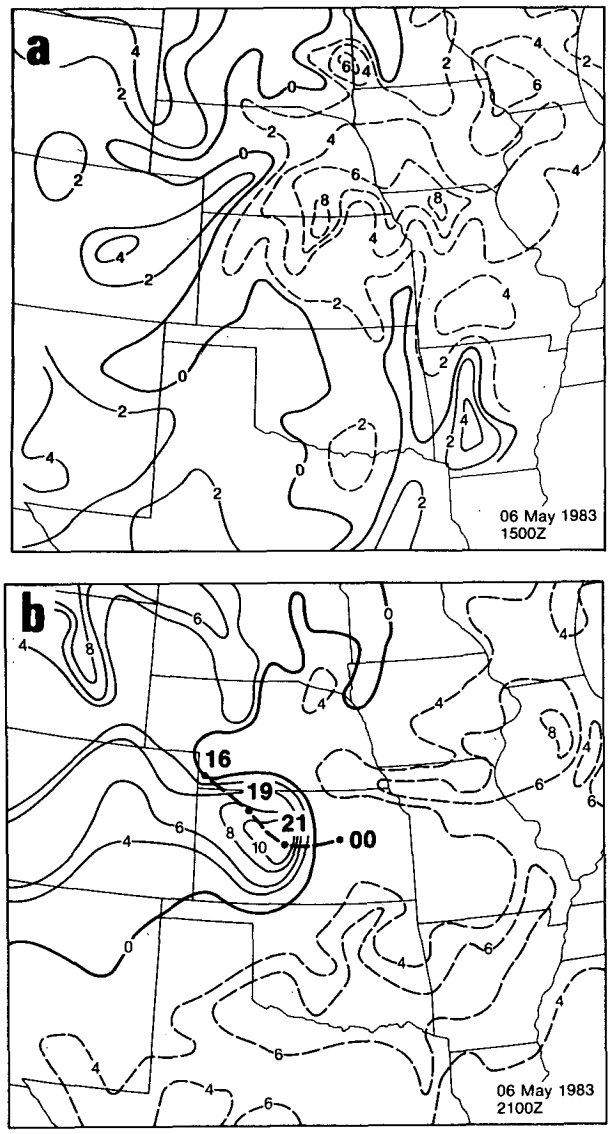

FIG. 10. Analysis of 2-h altimeter setting changes at (a) 1500 UTC and (b) 2100 UTC 6 May 1983, where the solid contours are altimeter setting rises (at intervals of 0.04 in Hg) and the dashed contours show altimeter setting falls (also at an interval of 0.04 in Hg). On (b), the dashed heavy line shows the track of the intense rise center, with selected locations at indicated times (UTC) shown by large dots.

ditional form (1a), they offer the advantage of combining the separate contributions from differential vorticity and thickness advection into a single forcing term on the right-hand side of (2).

Thus, (3) indicates that if there is *forcing* for vertical motion, there has to be a horizontal variation in the geostrophic wind (i.e., the height gradient cannot be uniform) and that variation must have a component parallel to the horizontal temperature gradient. For a purely barotropic atmosphere, there is no horizontal temperature gradient, so there can be no quasi-geostrophic forcing. In an equivalent barotropic atmosphere the thermal advection vanishes, so all the forcing must come from differential vorticity advection. The most intense forcing would be expected to be associated with the most baroclinic large-scale systems. Strictly speaking, a weather system could have large **Q**-vectors and still be characterized by weak forcing if the **Q**-vector pattern happened to have weak divergence. It is important to distinguish between the forcing for vertical motion and the response to that forcing, viz., the vertical motion itself. Specification of the right-hand side of (2) gives the forcing, but this is not equivalent to specifying the *response,* which requires finding a solution to (2). I shall return to this point later.

In spite of the success of the quasi-geostrophic system in explaining much of the dynamics of extratropical weather, modern numerical models for operational weather forecasting employ a more primitive set of equations. A primary motivation for employing more sophisticated models (short of the primitive equations), like the geostrophic momentum approximation (Hoskins and Bretherton, 1972 and Hoskins, 1975), is to be able to incorporate smaller-scale processes (e.g., the collapse of a large-scale gradient into a front) within a reasonably simple conceptual framework. Detail of this sort is not necessary nor desirable if one wishes to restrict attention to the essentials of large-scale processes.

At 1200 UTC 6 May 1983, the quasi-geostrophic forcing for vertical motion appears in Fig. 12, calculated in the manner described in Barnes (1985, hereafter B85), for three different pressure levels. Negative values of $\nabla \cdot \mathbf{Q}$ (i.e., convergence of the **Q**-vectors) imply forcing which favors upward motion, as can be seen by inspecting (2). The forcing in Iowa and Missouri decreases with height, while the forcing over the high plains of Colorado, Wyoming, and Montana increases with height. The net result is that the quasi-geostrophic forcing tilts westward with height over Kansas, a con-

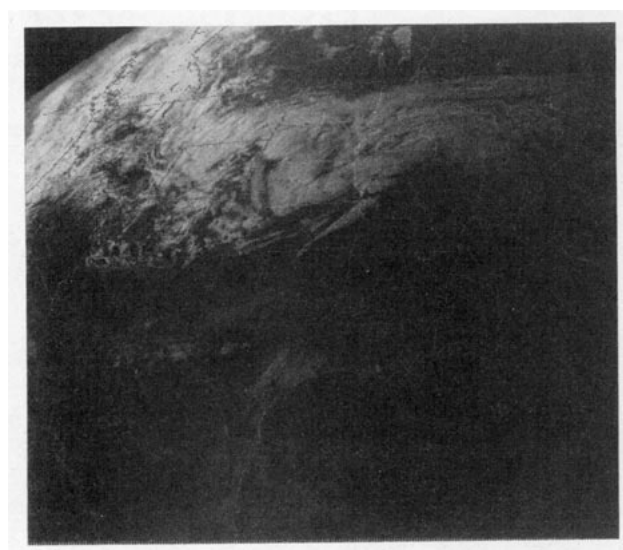

FIG. 11. As in Fig. 8 except at 0000 UTC 7 May 1983.

dition which is supportive of cyclogenesis (see Holton, 1979, §9.2.1).

At the same time, Fig. 13 shows the quasi-geostrophic frontogenesis (calculated from $2\mathbf{Q} \cdot \nabla\theta$, as in Hoskins and Pedder, 1980, and in B85), with little or no large-scale frontogenesis in the western Kansas region where the important cold front formed after 1200 UTC. Analysis of the same quantity at 0000 UTC, 7 May 1983 (not shown) reveals a similar absence of quasi-geostrophic contributions to the observed frontogenesis. While the LFM certainly suggested the presence of this front (recall Fig. 6b), the diagnostics indicate that it cannot be explained via large-scale processes. Barnes (1985) has an excellent discussion concerning the application of quasi-geostrophic frontogenesis diagnostics, the substance of which is that such calculations can be used to assess the favorability of the large-scale environment for frontogenesis, but they do not incorporate the processes by which the frontal gradients attain their observed values. Hence, this evaluation is most relevant early in the formation of a front, a condition which applies to the 1200 UTC data on 6 May 1983.

The problem of forecasting convection involves more than large-scale processes, but there does seem to be clear observational evidence for an association between large-scale systems and moist, deep convection. What are the physical origins for this association? It seems likely that the thermodynamic environment favorable for intense convection was created via large-scale processes as early as 1200 UTC (the large CAPE seen in Fig. 5). It also can be argued (as in D84) that the large-scale cyclone was responsible for the advance of the cold air behind the dryline. Although it cannot be demonstrated from the observations available, the large-scale lift also may have weakened the capping inversion in northeastern Kansas (although apparently not enough to permit the dryline-forced vertical motion, by itself, to develop strong thunderstorms). In other cases (e.g., Doswell et al., 1985), the large-scale advective processes bring about sufficient thermodynamic support to permit strong convection within initially unfavorable environments.

If we must look elsewhere for the lifting required to get deep, moist convection underway, a reasonable place to start is the so-called mesoscale. However, "mesoscale" must be defined. Following Emanuel's (1980) suggestion again (i.e., basing our scale division arguments on physical grounds) fails without a dominant force balance within the time and space ranges considered as candidates for mesoscale. Hence, all the terms in the so-called "primitive" equations are of potentially comparable magnitude in one situation or another. This problem is reflected in the rather wide range of processes considered in reviews of mesoscale meteorology (e.g., Atkinson, 1981). Included are lake-effect snows, convective mesosystems, downslope windstorms, solendoidal circulations (fronts, sea–land breezes, etc.), solitary waves, etc. While it is possible to consider each process separately in terms simpler than the complete "primitive" system of equations, there does not appear to be any obvious way to simplify that system of equations in a way which applies to *all* these mesoscale processes. Some involve moisture while others don't; some can ignore Coriolis while others can't, and so on. Is there any way to unify these disparate processes under some common physical grounds?

A unification is possible, but not by identifying which terms dominate the governing equations. Instead, I wish to draw attention to what makes "mesoscale" different from other scales of motion in the atmosphere. The essence of large-scale meteorology can still be captured if many things are neglected, and similarly, many factors in the governing equations (e.g., Coriolis effects, curvature of the earth, etc.) can be neglected with microscale meteorological processes, but mesoscale meteorology is different in that it is hard to neglect anything and retain any generality. Mesoscale processes are not studied in isolation from processes on scales above and below mesoscale. The very root of the term "mesoscale" indicates that it stands *between* scales. Thus, I propose that mesoscale processes be defined as those which cannot be understood without considering large scale *and* microscale processes.

When considering the front in Kansas on 6 May 1983, the quasi-geostrophic diagnostic evaluation of frontogenesis makes it apparent that the development of this intense temperature gradient was not the direct result of large-scale processes. Extensive cloudiness in southwestern Nebraska adjacent to clear, dry air in western Kansas is the most likely origin for the development of this front, as described in D84[2] (also see Segal et al., 1986). However, once such a mesoscale boundary developed, its subsequent march across Kansas, eventually to influence convective developments near Topeka, was influenced by the developing, low-level circulation associated with the large scale cyclogenesis. This front (and the convection along it) may have had some effect on that large scale system as well, although such an effect has not been identified in this study.

As suggested earlier, a single case study certainly does not permit generalization, but it is possible to try to create a coherent picture of the case studies available in the literature and through experience. I conclude that, in general, large-scale flows create the favorable thermodynamic environment while mesoscale pro-

[2] Some brief, but intense, thunderstorms that developed in the cold air behind this boundary during the afternoon could have contributed their outflows to the maintenance of the cold air mass. However, they were widely scattered and brief, so it is unlikely that they could represent a complete explanation for this extensive and intense mesohigh. Interested readers should consult D84 for a more thorough discussion.

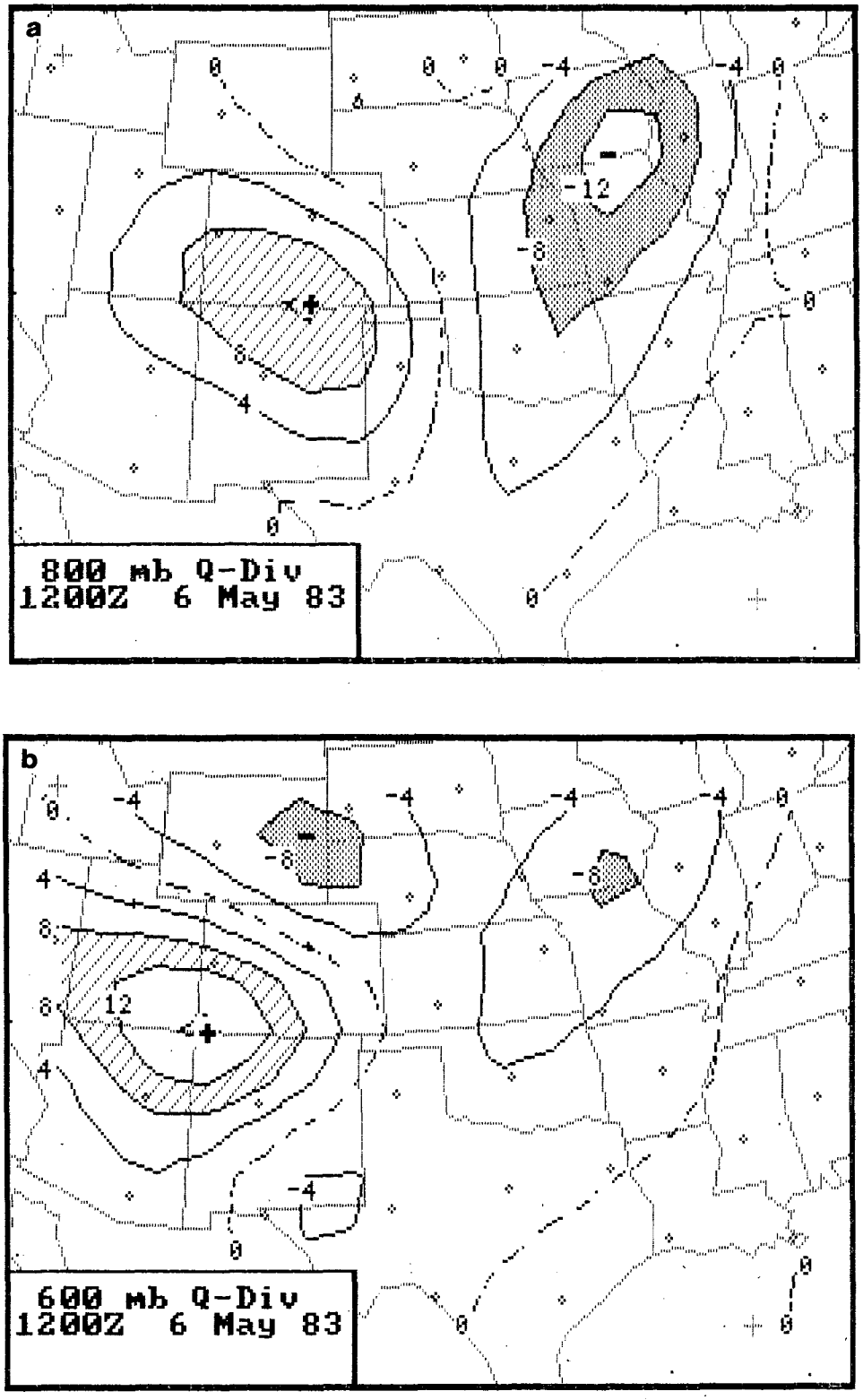

FIG. 12. The Q-vector divergence at 1200 UTC 6 May 1983 for (a) 800 mb, (b) 600 mb and (c) 400 mb. The contour interval is 4×10^{-17} s^{-3} mb^{-1}; hatching (in alternate contour bands beginning at 8×10^{-17} s^{-3} mb^{-1}) indicates quasi-geostrophic forcing leading to descent, while stippling (in alternate contour bands beginning at -8×10^{-17} s^{-3} mb^{-1}) indicates forcing leading to ascent. Rawinsonde sites are located by the small clusters of four dots.

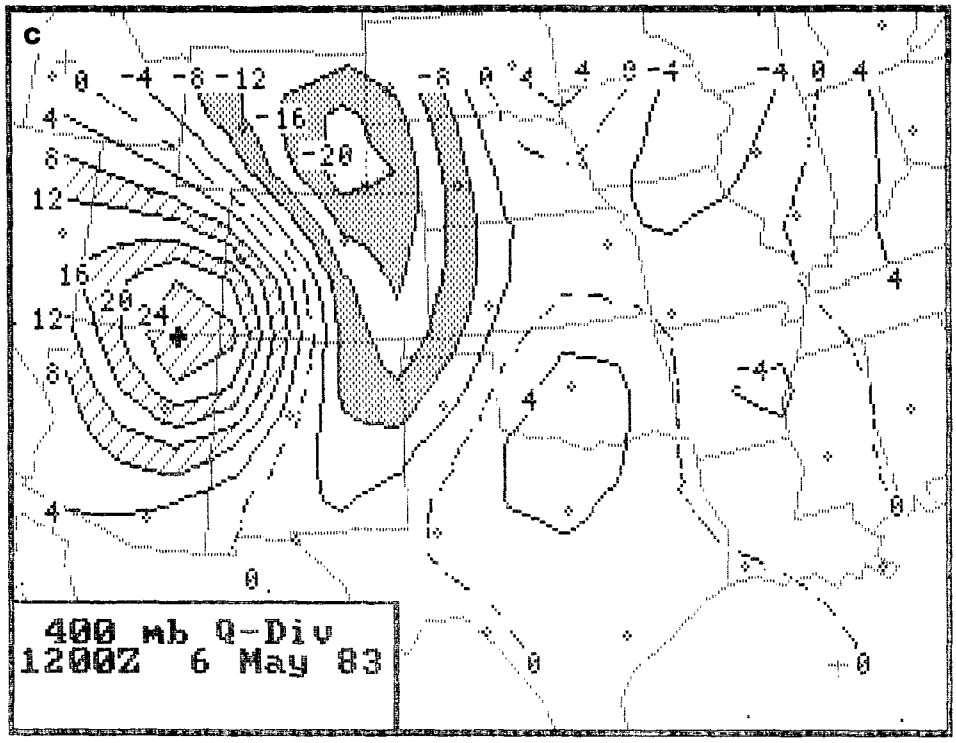

FIG. 12. (*Continued*)

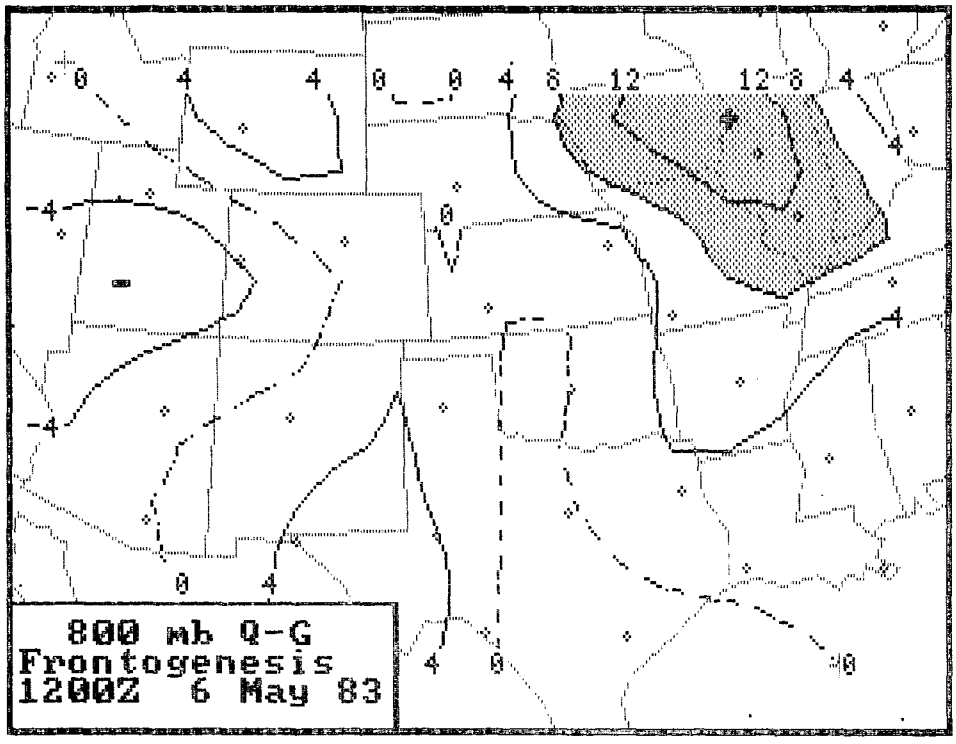

FIG. 13. Quasi-geostrophic frontogenesis at 1200 UTC 6 May 1983, contoured at intervals of 4×10^{-16} $K^2\,m^{-2}\,s^{-1}$. Positive values imply quasi-geostrophic frontogenesis (values greater than 8×10^{-16} are stippled), while negative values imply frontolysis.

cesses serve to provide the lift needed for convective initiation.

Given this generalization, mesoscale processes may at times be sufficient to initiate deep convection in environments that may be only marginally favorable for (or even inhibitive to) convection. Some examples come readily to mind: strong terrain-induced lift in situations with limited total moisture (e.g., mountain thunderstorms) and intense low-level forcing in the presence of a strong capping inversion (often the case with classic supercell thunderstorms), or when the air being lifted is only weakly unstable (as in the cases described by Carbone, 1982, and Emanuel, 1980). In such environments, the result may be short-lived thunderstorms or storms which remain isolated (although possibly intense) rather than growing into a widespread area of convection. As suggested by Chimonas and Kallos (1986), the mesoscale forcing may also be sufficient to allow the convection to continue, despite the unfavorable character of the large-scale environment, by means of local (i.e., mesoscale) environmental modification.

On the other hand, when the mesoscale initiation process operates in large-scale environments favorable to deep, moist convection, the result is most likely to be widespread, intense thunderstorms. The type of storm may be determined by ancillary factors like the vertical wind shear structure and/or the morphology of the mesoscale forcing (i.e., whether or not the mesoscale lift is linear in character). Extensive mesoscale convective systems (e.g., as in Maddox, 1980) generally are characterized by this combination of favorable large-scale conditions and mesoscale forcing (see B85, for instance). This is also true for such dramatic examples of widespread, intense severe weather as 3–4 April 1974.

Most convective events probably fall somewhere between these extremes. Like the Topeka storm on 6 May 1983, the mesoscale forcing may be in a region which is neither strongly favorable nor strongly inhibitive. Further, it is not necessarily obvious what constitutes a "favorable" environment, unless the sort of convective event under consideration has been specified. For supercell thunderstorms, a restraining inversion can be considered favorable, unless it is too strong to be broken by the processes operating on the mesoscale; then, the relative rarity of widespread, extremely violent convective weather can be understood in terms of the rarity of concatenating extremely favorable large-scale environments with the appropriate mesoscale processes to initiate (and, perhaps, enhance) the large scale potential.

4. Dynamic and thermodynamic factors

My experience in operational thunderstorm forecasting suggests that the contributions of dynamic and thermodynamic factors often are considered as separate issues. Moreover, there seems to be no universally accepted definition for either one of the terms "dynamics" and "thermodynamics" in reference to convective storms. However, I shall give my interpretation of what most forecasters mean by them, in spite of the high probability that many individuals will disagree with that interpretation.

By dynamics, most forecasters are referring to processes associated with large-scale weather systems. In midlatitudes, such systems are predominantly the baroclinic, quasi-geostrophic, extratropical cyclones, the study of which has permeated the scientific literature since the turn of the century. Our knowledge of these systems forms the basis of our understanding how to forecast the weather. In general the type, intensity, and location of observed weather is related rather closely to extratropical cyclones (at least in midlatitudes). However, as suggested by B85, the relatively weak large-scale forcing of summer poses a real problem for forecasters, especially with regard to forecasting convective phenomena, because of the apparently diminished role of large-scale weather systems in producing convective weather events.

Nevertheless, as B85 has demonstrated so clearly, the forcing associated with quasi-geostrophic processes can remain important in convective storms despite the decrease in overall intensity of the extratropical cyclones during the warm season for two essential reasons. First, the overall increase in lapse rates during the warm season makes a given amount of forcing more effective in producing large-scale vertical motion. The weaker forcing of summer may yield vertical motion roughly comparable to that of cool season systems because it is easier to lift air when the lapse rate is large. Second, as the lapse rate increases, baroclinic processes tend to yield systems which have smaller spatial and temporal scales (see Gall, 1976). Although the systems themselves are smaller and somewhat weaker, they can still be important in modulating the convective environment. Thus, I propose that the term "dynamics" as used in weather forecasting should be defined as *large-scale, quasi-geostrophic forcing* (as discussed before), which can be assessed from diagnostic models like that described in B85.

When considering thermodynamics in the context of convective storm forecasting, the standard tools of diagnosis are indices of various types—the Showalter index (Showalter, 1953), the Lifted index (Galway, 1956), the *K*-index (George, 1960), and so on. Such indices are keyed primarily to mandatory pressure levels and suffer from at least two deficiencies. First, if the thermodynamic structures they are designed to parameterize happen to change dramatically near a mandatory level, the resulting index value may be misleading. Second, by combining moisture and lapse rate measures into a single index, these indices provide little insight into those processes which may alter moisture and lapse rate distributions independently. While the

first deficiency can be overcome by measuring convective potential in terms of CAPE instead of indices tied to mandatory levels, this fails to address the second deficiency associated with indices. The CAPE observed in a given sounding can change as a result of changes in moisture, lapse rate, or both. Doswell et al. (1985) have pointed out that only limited CAPE may be found in soundings at a particular analysis time, but the lapse rate and/or moisture distribution may be configured so that large CAPE values will be generated subsequently.

Therefore, I propose that the term "thermodynamics" be defined as *the combination of moisture and lapse rate distributions which makes deep, moist convection possible.* This definition is intended to account not only for situations with substantial existing CAPE, but also for situations with little or no CAPE at analysis time that are evolving so as to yield large CAPE in the near future. My intention is to emphasize the changing, rather than static, thermodynamic structures important to convection.

While moisture is a more or less passive variable in the flow dynamics (short of condensation), the lapse rate can have a substantial impact on the evolution at all scales, up to and including the large-scale dynamics.[3] By the same token, large-scale dynamic processes modify moisture and lapse rate distributions. Thus, a sort of synergistic interaction between dynamics and thermodynamics can occur. Via this interaction, an environment favorable for convection develops, making it conceptually erroneous to consider dynamic and thermodynamic factors independently. The interaction also has the potential for modifying the flow field in ways which favor certain types of convection (via the local vertical wind shear, as discussed in Weisman and Klemp, 1984, for example). If the notion is accepted that the type, intensity, and distribution of convective storms is dependent on the environment in which the convection occurs, then it becomes exceedingly important to understand how that environment is created and maintained. To envision dynamic and thermodynamic factors as essentially independent is to make that understanding difficult, if not impossible.

5. Summary and discussion

The severe weather events in eastern Kansas on the evening of 6 May 1983 have been used to illustrate a conceptual framework for understanding the processes of scale interaction leading to intense convection, at least in midlatitudes. By limiting the definition of the term "large scale" to quasi-geostrophic processes, quantitative estimates of the large-scale contributions to any weather situation can be provided. Thus, this viewpoint can be applied to any midlatitude weather event, be it convective or otherwise. The large-scale contribution to the event can then be extracted, leaving a residual which is unexplained from the large-scale viewpoint. One might call this unexplained residual the "subsynoptic scale," including all nonquasi-geostrophic processes.

The next element in the scale interaction problem is to try to provide a physically meaningful way to interpret the term "mesoscale." The proposed interpretation is that mesoscale processes are those which cannot be understood (and hence cannot be forecast accurately) by considering them in isolation from processes operating on scales above and below. By assuming that the *primary* role of mesoscale processes is to provide the lifting necessary to initiate deep, moist convection, then it is possible to understand how the interaction between mesoscale and large scale governs the intensity and evolution of the resulting convection. This is not to say that the sole role for mesoscale processes vis-a-vis convection is to serve as a means to initiate moist convective ascent. Mesoscale processes may modify the local environment in ways that can change the type of convection possible (as in Wicker et al., 1983), or that even alter the potential for any convection at all. In fact, mesoscale processes may be necessary in some situations to accomplish the final concatenation of the ingredients for a particular weather event—e.g., the rainbands described by Chimonas and Kallos (1986).

Further, the mesoscale events are determined in large part by the large-scale structures. For example, upslope flow depends on the position of the topography relative to the large-scale system. In general, both internally and externally forced mesoscale instabilities depend on large-scale conditions. Were this not the case, mesoscale systems would be far more common and might appear to be more or less random.

The argument I have advanced about the vertical motion on large scales indicates that convection typically is *initiated* by mesoscale (or smaller scale, like that of the individual storm cell) events. This means forecasters should recognize that large-scale systems are not likely to be the "trigger"—a term used frequently to refer to the lifting necessary for initiation—for convective episodes.

This case underscores the importance of the mesoscale to convective events but necessarily leaves unanswered the problem of how to make mesoscale forecasts. While I have attempted to provide a conceptual framework for understanding the interaction between large-scale and mesoscale processes, this does not constitute a quantitative evaluation of the interaction. Without a detailed, quantitative analysis of scale interactions, a rigorous approach to forecasting convec-

[3] As noted previously, a strict interpretation of quasi-geostrophic theory requires that the lapse rate not vary horizontally. However, one still can consider the effect of changing the lapse rate over broad areas through changing the value of the assumed horizontally uniform σ in (1a, b)—see the discussion by Gall, (1976).

tive mesosystems remains elusive. Also, a consequence of my definition of mesoscale meteorology is that it implies some grasp of virtually all of meteorology in order to understand mesoscale processes. Thus, there can be no "mesoscale meteorologists"—only meteorologists who may have a special interest in mesoscale events. In this same sense, one cannot do mesoscale forecasting and analysis; rather, one does forecasting and analysis which may happen to require knowledge of mesoscale processes from time to time (see D87 for further discussion on this point). To be committed to mesoscale meteorological aspects of research and/or forecasting is to be forced to deal with processes ranging from microphysics to the global circulation. Limiting attention to only those phenomena of a peculiar "mesoscale" length and time range is, in my view, tantamount to failing to grasp the essence of mesoscale meteorology.

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REFERENCES

Atkinson, B. W., 1981: *Mesoscale Atmospheric Circulations.* Academic Press, 495 pp.

Barnes, S. L., 1985: Omega diagnostics as a supplement to LFM/MOS guidance in weakly forced convective situations. *Mon. Wea. Rev.*, **114,** 2121–2141.

Carbone, R. E., 1982: A severe frontal rainband. Part I: Stormwide hydrodynamic structure. *J. Atmos. Sci.*, **39,** 259–279.

Chimonas, G., and G. Kallos, 1986: Flow dynamics and stability in a severe rainband. *J. Atmos. Sci.*, **43,** 1505–1516.

Doswell, C. A. III, 1984: Mesoscale aspects of a marginal severe weather event. *Preprints Tenth Conf. Weather Forecasting and Analysis,* Clearwater Beach, Amer. Meteor. Soc., 131–137.

——, 1987: Short range forecasting. *Mesoscale Meteorology and Forecasting,* Amer. Meteor. Soc., 689–719.

——, F. Caracena and M. Magnano, 1985: Temporal evolution of 700–500 mb lapse rate as a forecasting tool—A case study. *Preprints 14th Conf. Severe Local Storms,* Indianapolis, Amer. Meteor. Soc., 398–401.

Dutton, J. A., 1976: *The Ceaseless Wind.* McGraw-Hill, 579 pp.

Emanuel, K. A., 1980: Forced and free mesoscale motions in the atmosphere. *Collection of Lecture Notes on Dynamics of Mesometeorological Disturbances,* Y. K. Sasaki, N. Monji and S. Bloom, Eds., Cooperative Institute for Mesoscale Meteorological Studies, 189–259.

Fujita, T. T., 1981: Tornadoes and downbursts in the context of generalized planetary scales. *J. Atmos. Sci.*, **38,** 1511–1534.

——, and A. D. Pearson, 1973: Results of FPP classification of 1971 and 1972 tornadoes. *Preprints Eighth Conf. Severe Local Storms,* Denver, Amer. Meteor. Soc., 142–145.

Gall, R., 1976: Structural changes of growing baroclinic waves. *J. Atmos. Sci.*, **33,** 374–390.

Galway, J. G., 1956: The lifted index as a predictor of latent instability. *Bull. Amer. Meteor. Soc.*, **37,** 528–529.

——, 1966: The Topeka tornado of 8 June 1966. *Weatherwise,* **19,** 144–149.

George, J. J., 1960: *Weather Forecasting for Aeronautics.* Academic Press, 673 pp.

Holton, J. R., 1979: *An Introduction to Dynamic Meteorology.* Second ed., Academic Press, 391 pp.

Hooke, R., 1963: *Introduction to Scientific Inference.* Holden-Day, 101 pp.

Hoskins, B. J., 1975: The geostrophic momentum approximation and the semi-geostrophic equations. *J. Atmos. Sci.*, **32,** 233–242.

——, and F. P. Bretherton, 1972: Atmospheric frontogenesis models: Mathematical formulation and solution. *J. Atmos. Sci.*, **29,** 11–37.

——, and M. A. Pedder, 1980: The diagnosis of middle latitude synoptic development. *Quart. J. Roy. Meteor. Soc.*, **106,** 707–719.

——, I. Draghici and H. C. Davies, 1978: A new look at the ω-equation. *Quart. J. Roy. Meteor. Soc.*, **104,** 31–38.

Maddox, R. A., 1980: Mesoscale convective complexes. *Bull. Amer. Meteor. Soc.*, **61,** 1374–1387.

——, and C. A. Doswell III, 1982a: An examination of jet stream configurations, 500 mb vorticity advection and low-level thermal advection patterns during extended periods of intense convection. *Mon. Wea. Rev.*, **110,** 184–197.

——, and ——, 1982b: Forecasting severe thunderstorms: A brief consideration of some accepted techniques. *Nat. Wea. Dig.*, **7,** 26–30.

McNulty, R. P., 1978: On upper tropospheric kinematics and severe weather occurrence. *Mon. Wea. Rev.*, **106,** 662–672.

Miller, R. C., 1972: Notes on analysis and severe-storm forecasting procedures of the Air Force Global Weather Central. Air Weather Service Tech. Rep. 200 (Rev.), Air Weather Service, Scott Air Force Base, 190 pp.

Moncrieff, M. W., and M. J. Miller, 1976: The dynamics and simulation of tropical cumulonimbus and squall lines. *Quart. J. Roy. Meteor. Soc.*, **102,** 373–394.

Orlanski, I., 1975: A rational subdivision of scales for atmospheric processes. *Bull. Amer. Meteor. Soc.*, **56,** 527–530.

Segal, M., J. F. W. Purdom, J. L. Song, R. A. Pielke and Y. Mahrer, 1986: Evaluation of cloud shading effects on the generation and modification of mesoscale circulations. *Mon. Wea. Rev.*, **114,** 1201–1212.

Showalter, A. K., 1953: A stability index for thunderstorm forecasting. *Bull. Amer. Meteor. Soc.*, **34,** 250–252.

Uccellini, L. W., 1980: On the role of upper tropospheric jet streaks and leeside cyclogenesis in the development of low-level jets in the Great Plains. *Mon. Wea. Rev.*, **108,** 1689–1696.

Weisman, M. L., and J. B. Klemp, 1984: The structure and classification of numerically simulated convective storms in directionally varying wind shears. *Mon. Wea. Rev.*, **112,** 2479–2498.

Wicker, L., H. Bluestein and D. Burgess, 1983: The pre-storm environment of the severe weather on 22 May 1981 in western Oklahoma. *Preprints 13th Conf. Severe Local Storms,* Tulsa, Amer. Meteor. Soc., 281–282.