Convective Processes Resolved by a Mesoscale Rawinsonde Network

J. C. FANKHAUSER

National Center for Atmospheric Research, Boulder, Colo.

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ABSTRACT

Guidelines followed in designing and operating a special mesoscale rawinsonde network are discussed. Objective data reduction and analysis techniques are developed and applied to the wind, temperature and moisture data measured during a selected thunderstorm case. The goal is to appraise the sounding system's limitations and reliability for resolving the mesoscale circulations associated with convective processes. A consistent four-dimensional synoptic portrayal of the variables is achieved by accounting for balloon drift, differing station-to-station and sounding-to-sounding ascent rates, and departures from scheduled release time.

Temporal variations in the spatial distributions of computed divergence and kinematic vertical motion are in good qualitative agreement with the location and intensity of thunderstorm radar echoes, after further objective adjustments are applied to compensate for the assumed character of wind measurement and analysis errors. For the purpose of assessing data and analytical credibility, independent estimates of the energy related to diabatic heating are obtained by evaluating thermodynamic energy and moisture continuity equations. Comparative results indicate that the network upper-air data and the methods adopted for its synthesis are internally consistent. On the other hand, the distribution and magnitude of the resolved kinematic and dynamic features demonstrate that the results are clearly a function of observational spacing and do not necessarily pertain to individual convective processes.

1. Introduction

Essential to the understanding of how thunderstorms respond to and, conversely, modify their environment, is the adequate description of atmospheric variables at an appropriate scale. From the standpoint of forecasting, House (1960) has demonstrated that the conventional synoptic upper-air network is inadequate for detecting the processes relevant to such gross convective phenomenon as squall lines. The observational deficiencies become even more pronounced in the thunderstorm research area (Fujita and Brown, 1960).

In order to provide the necessary framework in which to view refined thunderstorm observations obtained from the surface networks, instrumented aircraft and radar, the ESSA National Severe Storms Laboratory (NSSL) initiated a program of mesoscale serial upper-air soundings in Oklahoma in 1966. This paper discusses the guidelines followed in establishing and operating the rawinsonde network, development of data reduction and analysis techniques, and the results of a diagnostic study directed toward defining the threshold of confidence in mesoscale upper-air observations obtained in the presence of thunderstorms.

2. On determining observational requirements

In designing an observational network the optimum spacing of observing sites should be governed by the average dimensions of the phenomena requiring detection. According to Gleeson (1959), the probability that at least one station will detect the local effects of an atmospheric feature covering an area $S$ is given by

$$P = 1 - [1 - (S/R)]^N, \quad (1)$$

where $R$ is the area of the region containing $S$, and $N$ is the number of observing stations distributed randomly within $R$.

With regard to thunderstorms, the scale of significant features varies from that of the nearly macroscale squall line down to individual convective circulations covering a few square miles. In the study of six convective situations, Newton and Fankhauser (1964) found the typical diameter of radar echoes from over 200 Oklahoma thunderstorms to be on the order of 8–12 n mi. If the disturbing effects of an individual convective circulation prevail over, say, twice the diameter of a typical storm echo (see, e.g., Shmeler, 1968), then a minimal value for $S$ in Eq. (1) can be taken as about 300 n mi$^2$. Fig. 1, based on solutions of (1) with $R$ set at 8000 n mi$^2$ (the area covered by the NSSL mesonetwork of surface stations), indicates that to achieve an 80% probability of detecting a storm of this size, approximately 40 stations would be required. In this light, it...
would appear that the NSSL surface network, which has grown to nearly 60 observing sites during the 1960’s, is reasonably well suited to observe local events associated with a typical Oklahoma thunderstorm.

Upon considering the migratory nature of thunderstorms, and by regarding serial rawinsonde ascents to be quasi-continuous, a broader interpretation can be given to S. If a “disturbance” of 20 n mi diameter moves through R along a path 50 n mi in length, then S’, the effective area swept out, will be 1000 n mi². The number of observing points required to maintain an 80% probability of detection in this case is reduced to 12. Viewed in another way, there is an 80% chance of observing features within 20 n mi of the center of a thunderstorm which translates along a 50 n mi path within an 8000 n mi² area while 12 randomly spaced sites are taking sequential soundings.

This cursory review of observational probabilities is intended as a background for assessing the feasibility and potential of a network of rawinsonde stations established expressly for thunderstorm research. Throughout the discussion only probabilities associated with detection are considered, and it will be the purpose in the remainder of the paper to approach the “definability” problem. Description, interpretation and definition are, first of all, dependent on detection, but other factors enter. Not the least of these are the accuracy of the observations and the adequacy of methods adopted for their reduction, as will be seen in later sections.

3. Network design and operational procedures

Based on the foregoing statistical considerations and with due regard for logistical and economic limitations, a network of rawinsonde stations was established to coincide with the location of the existing NSSL mesonetwork of surface sites in south central Oklahoma. In Fig. 2, the positions of nine stations which participated in the upper-air program in 1966 and 1967 are shown as bold dots. During the 1966 thunderstorm season, an additional site was in operation at Childress, Tex., which is located approximately 90 km WSW of Altus, Okla., designated as LTS in Fig. 2. Table 1 lists individual site specifications and should be used in identifying station call letters referred to later in this paper. Average separation between sites is 46 n mi (85 km). Rawinsonde observations at a comparable scale and frequency have been obtained by the Japanese Heavy Snow Storm Project (Matsumoto, 1967) and, in New England, by the AFCLR Project Stormy Spring (Kreitzberg, 1968).

Following the recommendations of Fujita and Brown (1960), the center of the upper-air network is skewed slightly toward the climatological upwind (southwest) direction, with respect to the surface network, to compensate for balloon displacement during flight. In a typical convective situation, where mean winds might average 240° at 30 kt, a balloon would be displaced northeastward approximately 30 n mi by the time it reached the tropopause. Since all rawinsonde receiving equipment operates within a common assigned frequency range, a lower limit in station spacing must be maintained to insure against reception of the signals from more than one balloon at a single station. The possibility of confused signals was further obviated by distributing frequencies so that adjacent sites operated at opposite ends of the allotted range.

An operational period extending from 15 April to 15 June was selected to cover Oklahoma’s climatic peak in thunderstorm activity. Midday observations were scheduled daily at each station and allowed to continue to natural termination. On days when significant convection seemed imminent, a series of simultaneous soundings was initiated. These were purposely terminated at 100 mb in order to maintain a 90-min release interval.

Since the data produced by the network were to be used primarily for research purposes, some departures from conventional data acquisition procedures were desirable. These included close adherence to uniform balloon inflation and the recording of azimuth and elevation angles at the maximum possible frequency (10 records min⁻¹) immediately following release, and when elevation angles were <12°. In addition, scheduled release times were made flexible enough to avoid balloon ascents into centers of active thunderstorms whenever possible, because of the high likelihood of system failure in such a case. Storm avoidance was further deemed advisable, since the resolvable wavelengths permitted by the sounding frequency and spacing would not allow scales of motion measured

![Fig. 1. Nomogram giving detection probability P as a function of phenomenon size S, with diameter D and the number of observing points N, within a prescribed region of area R (adapted from Gleason, 1959).](image)
Fig. 2. NSSL mesonetwork, showing hourly frontal (heavy solid) and mesosystem (heavy dashed) boundaries, radar echo position and intensity (contour interval, 10 dB), and surface winds, for 28 May 1967. Vectors originate at surface recording stations and bold dots designate rawinsonde sites. Lower range limit in radar surveillance is indicated by dashed circle.
Table 1. NSSL rawinsonde site specifications, 1966-1967.

<table>
<thead>
<tr>
<th>Location</th>
<th>Station identification</th>
<th>Station no.</th>
<th>Latitude N</th>
<th>Longitude W</th>
<th>Station height (m)</th>
<th>Equipment type</th>
<th>Support agency</th>
<th>Assigned operating frequency (kcs)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Chickasha, Okla. (Municipal Airport)</td>
<td>CHK</td>
<td>1</td>
<td>35°06'</td>
<td>97°58'</td>
<td>349</td>
<td>GMD-1</td>
<td>USAF(1966)</td>
<td>ESSA(1967)</td>
</tr>
<tr>
<td>Wichita Falls, Tex. (Civilian Air Terminal)</td>
<td>SPS</td>
<td>2</td>
<td>33°59'</td>
<td>98°30'</td>
<td>309</td>
<td>GMD-1</td>
<td>US Army</td>
<td>1675</td>
</tr>
<tr>
<td>Fort Sill, Okla. (USAAMC Meteorological Station)</td>
<td>FSI</td>
<td>3</td>
<td>34°39'</td>
<td>98°24'</td>
<td>362</td>
<td>GMD-1</td>
<td>US Army</td>
<td>1680</td>
</tr>
<tr>
<td>Altus, Okla. (Municipal Airport)</td>
<td>LTS</td>
<td>4</td>
<td>34°42'</td>
<td>99°20'</td>
<td>435</td>
<td>GMD-1</td>
<td>US Army</td>
<td>1685</td>
</tr>
<tr>
<td>Tinker AFB, Okla. (South side of base)</td>
<td>TIK</td>
<td>5</td>
<td>35°25'</td>
<td>97°23'</td>
<td>387</td>
<td>GMD-2</td>
<td>USAF</td>
<td>1690</td>
</tr>
<tr>
<td>Cordell, Okla. (Municipal Airport)</td>
<td>COR</td>
<td>6</td>
<td>35°18'</td>
<td>98°58'</td>
<td>482</td>
<td>GMD-1</td>
<td>USAF</td>
<td>1670</td>
</tr>
<tr>
<td>Pahs Valley, Okla. (Municipal Airport)</td>
<td>PVY</td>
<td>7</td>
<td>34°42'</td>
<td>97°13'</td>
<td>295</td>
<td>GMD-1</td>
<td>USAF</td>
<td>1670</td>
</tr>
<tr>
<td>Ringling, Okla. (Public school grounds)</td>
<td>RIN</td>
<td>8</td>
<td>34°10'</td>
<td>97°35'</td>
<td>275</td>
<td>GMD-1</td>
<td>USAF</td>
<td>1685</td>
</tr>
<tr>
<td>Watonga, Okla. (Municipal Airport)</td>
<td>WAT</td>
<td>9</td>
<td>35°51'</td>
<td>98°25'</td>
<td>472</td>
<td>GMD-1</td>
<td>USAF</td>
<td>1685</td>
</tr>
<tr>
<td>Childress, Tex. (Municipal Airport)</td>
<td>CDS</td>
<td>10</td>
<td>34°26'</td>
<td>100°18'</td>
<td>595</td>
<td>GMD-1</td>
<td>USAF</td>
<td>1680</td>
</tr>
</tbody>
</table>

* 1966 only.

within the turbulent convective cells to be dealt with effectively, even if the equipment survived the ascent.

4. Data reduction and processing

Automatic computer techniques were considered essential for reducing the network soundings, not only to provide a consistent and efficient treatment of the enormous volume of data, but also to preserve the basic accuracy of the sounding system. Except for a few minor revisions, the computer program adopted for treating raw data from single soundings is the same as that developed by Kreitzberg and Brockman (1966). Basic input is taken from the original recorder charts and control recorder printouts produced by the GMD-1 sounding system, and consists of calibrated ordinates tabulated manually from the temperature and humidity traces. These data are entered with pressure and elapsed time after launch at each whole pressure contact. In this format, intervals of input for the thermodynamic and pressure data amount to approximately 30 sec or 150 m in height.

Winds and balloon positions with respect to launch site are computed from azimuth and elevation angles entered at 30-sec intervals and interpolated to produce data concurrent with pressure and thermodynamic parameters at each pressure contact. Initial wind computations are subjected to a (1, 2, 1) smoothing process that results in a mean vector over about 600 m vertical thickness or approximately 2 min of balloon flight.

Störve-type diagrams (T vs ρ*) in Fig. 3 show plots of detailed computer output from network soundings obtained close together in time and space. Wind vectors on the right of the individual diagrams indicate the basic data density. Although the soundings display numerous common features, it is clear that point values on pressure surfaces are not apt to preserve reliable synoptic continuity. Inclusion of singularities in the analysis is avoided by deriving 50-mb means of all pertinent meteorological variables, incorporating all input data points in the averaging. As a consequence of a pressure-contact input interval, the number of points available for deriving layer means increases with the thickness of the layers.

Before the layered data can be subjected to a consistent synoptic analysis, it is necessary to account for balloon displacement horizontally from its point of release. At the mesoscale, compensation must also be made for the effects of non-simultaneous releases and differing ascent rates.

From layer means in a single station's series, balloon position, wind, temperature and mixing ratio are derived at hourly intervals for 17 layers between 950 and 100 mb by applying a linear interpolation formula of the form

$$x = x_1 + \left(1 - \frac{t-t_1}{t_2-t_1}\right)(x_2-x_1), \quad (2)$$
where $x$ represents the variable of concern and subscripts 1 and 2 refer to adjacent soundings in time $t$. Data plotted on the vertical time axes in Fig. 4 are objectively derived by the application of (2) to the conditions observed during the actual ascents represented as slanted lines. The importance of compensating for variable ascent rates and departures from scheduled release time is demonstrated by comparing ascent curves in the two series. The second scheduled release was made 15 min earlier at CHK; however, a more rapid ascent rate at FSI allows both balloons to reach the 100-mb level almost simultaneously.

Once the time-consistent output from the vertical sections is obtained, it is plotted at the appropriate interpolated balloon position on horizontal sections and analyzed to produce a three-dimensional Eulerian array on a stationary grid at specified hourly intervals. Grid-point values are read from analyses which may be regarded as quasi-objective in that, once there is assurance that soundings have not entered active con-
Fig. 4. Interpolated time sections (CST) of wind direction and speed (m sec⁻¹), plotted to right of data point; temperature (°C), to upper left; and mixing ratio (gm kg⁻¹), lower left. Slanted lines are time vs pressure altitude curves for actual soundings.
Fig. 5. Streamlines (solid), isotherms (heavy dashed) and temperature (thin dashed) analysis of interpolated 500-450 mb layer means at (a) 1200, (b) 1300, (c) 1400, (d) 1500, (e) 1600, and (f) 1700 CST, 28 May 1967. Crosses represent interpolated balloon positions and bold dots are station locations. Grid dimensions, 200 km x 200 km, as in Fig. 2.

convect, none of the data is questioned from the standpoint of physical reality. Within the constraints of the interpolated values, care is taken, however, to preserve horizontal and vertical time and space continuity, particularly in positioning discontinuities.

Examples of synoptic analysis of interpolated data at two selected levels is presented in Figs. 5 and 6. Gradients between balloons (crosses) and stations (dots) indicate that failure to account for balloon drift could result in as much as 2 m sec\(^{-1}\) and 0.5C error in wind speed and temperature evaluation in this case (see, e.g., Figs. 6e and 6f). Greater balloon-station separations in
Fig. 6 demonstrate that analytical problems associated with balloon drift become more critical at higher levels and in stronger wind regimes. Horizontal variations in the mean wind from the surface to a particular level of reference become apparent by comparing the separation between station and balloon in the northwest portion of the grid to that in the southeast (Fig. 6b).

5. Some features of mesoscale kinematics and thermodynamics in a selected thunderstorm situation

Analytical procedures outlined above were applied to network data obtained on 28 May 1967. On this day the usual ingredients favoring thunderstorm formation,
such as potential instability and pronounced vertical wind shear (see, e.g., Newton, 1963), were only marginally conducive to significant convection. Attention was purposely restricted to a situation of this type in an attempt to resolve the inherent limitations of the sounding system.

Fig. 7 shows the synoptic conditions over the central United States at the surface, 500 and 250 mb. Small scattered thunderstorms formed along the quasi-stationary front in Oklahoma shortly after noon, and by midafternoon a few storms of moderate intensity persisted within the NSSL observing facility. Surface charts in Fig. 2, corresponding to the small area delineated in Fig. 7, depict the hourly positions of frontal discontinuities, distribution and intensity of
radar precipitation echoes and the associated meso-scale winds.\textsuperscript{3}

\textsuperscript{3}Circulation features near individual thunderstorms occurring on this date have been described earlier by Booker et al. (1967), who reported on the analysis of superpressure balloon trajectories; Brown and Peace (1968), who presented the results of dual Doppler radar measurements in the subcloud layer; and Fankhauser (1968), whose analysis is based on chaff trajectories and aircraft winds at midcloud levels.

A series of five soundings, released at approximately 90-min intervals, was obtained at each of the nine sites comprising the upper-air network. The series commenced at 1100 and terminated at 1700 CST; thus, as indicated by the development of radar echoes in Fig. 2, the observational period completely encompassed the evolution of significant convective activity. Close exami-
Passage of the sharp ridge, and approach of the shortwave trough which is indicated in Fig. 7c at 250 mb, are reflected by the strong backing at the same level on the time sections. Fig. 6 demonstrates the evolution of mesoscale wind and temperature patterns as the wave system passes through the rawinsonde network. It may be noted that the speed minimum appearing at 250 mb on the time sections in Fig. 8 is located somewhat upstream from the ridge axis in Fig. 6.

A weaker synoptic scale trough at 500 mb, indicated in Fig. 7b, is accompanied by an extremely shallow but well-defined wind speed maximum which is discernible near 450 mb on all time sections. Among their empirical rules for tornado forecasting, Fawbush et al. (1951) list criteria for a narrow speed maximum in the midtroposphere. Although the thunderstorms that form in this case could not be characterized as severe, it is clear that the period and location of maximum radar echo developments shown in Fig. 2 are closely associated with the progression of the 15–16 m sec\(^{-1}\) wind speed maximum across the northwest portion of the network, shown in Fig. 5.

Barbé (1957) found perturbations in mesoscale upper-air observations which are comparable to those demonstrated in Fig. 8. Failure of horizontal temperature gradients to support the degree of vertical shear indicated in the time sections, and by Figs. 5 and 6, substantiates the conclusion by Weinstein et al. (1966) and Kreitzberg (1968) that motions at these scales are strongly ageostrophic. From thermal wind considerations, pronounced backing with height in the 500–200 mb layer would imply that strong cold advection exists over nearly all of the western portion of the network toward the middle of the observational period. Analysis of the temperature advection in the 350–300 mb layer at 1500 CST, shown in Fig. 9, indicates that cold advection aloft is actually confined to a narrow band which is aligned along and maximized just downstream from the active convection shown in Fig. 2d. Advection patterns at other times display the same banding along similar correlations with the location of most vigorous thunderstorms, in that the axis marking transition from cold-to-warm advection at high levels lies over the area of strongest radar echoes. The contribution of differential advection toward maintaining potential instability is most pronounced over the north central portion of the grid at 1500 CST and helps to explain the intensity and persistence of the largest echoes in Fig. 2d.

Fig. 10 represents the three-dimensional distribution of the mean horizontal winds in four selected 50-mb layers centered at the designated levels. The diluent, large amplitude, short wavelength impulse near 225 mb, the layer of maximum wind speed between 500 and 400 mb, and the convergence associated with the southward moving frontal zone, all discussed in this section, are illustrated by the transitions in the vector fields.
Fig. 10. Machine plotted three-dimensional distribution of horizontal mean wind in selected 50-mb layers: grid interval, 20 km. Vectors originate at dots (8.5 m sec\(^{-1}\) = 10 km).
6. Resolved distributions of divergence and vertical velocity

On an 11x11 grid corresponding to the region shown in Fig. 2, the mean divergence \( \mathbf{D} \) of the horizontal wind in 17 layers between 950 and 100 mb is computed from

\[
\mathbf{D}_k = \left( \frac{\partial \bar{u}}{\partial x} \right)_p + \left( \frac{\partial \bar{v}}{\partial y} \right)_p,
\]

where \( \bar{u} \) and \( \bar{v} \) represent average velocity components in the layer \( k \), and subscript \( p \) denotes reference to constant pressure surfaces. In the layer between the ground and the first overlying nonintercepting constant pressure surface \( p_1 \), approximate compensation for the effects of the variable surface pressure \( p_0 \) at the lower boundary is included by applying

\[
\mathbf{D}_1 \approx \frac{1}{\delta p} \left( \frac{\partial (\bar{u} \delta p)}{\partial x} + \frac{\partial (\bar{v} \delta p)}{\partial y} \right)_p,
\]

where \( \delta p = \bar{p}_0 - p_1 \) is the mean pressure depth of the particular horizontal grid interval involved.\(^4\) In the practical evaluation of (4), \( \bar{u} \) and \( \bar{v} \) are taken from the denser surface network winds, shown in Fig. 2, and are considered representative of the mean components in the lowest layer, which varies in pressure depth from about 30 mb in the southeast portion of the grid to less than 5 mb in the northwest.

Integration of the equation of continuity,

\[
\frac{\partial \omega}{\partial p} + \mathbf{D}_k = 0,
\]

yields the vertical component of velocity, \( \omega_p \), at the top of all layers \( k \) between 950 and 100 mb as

\[
\omega_p = \omega_p + \Delta_p + \mathbf{D}_k \Delta p, \quad (\Delta p = 50 \text{ mb}).
\]

As is often the case, results obtained from (6) were physically realistic and acceptable in low and middle tropospheric levels but diminished in credibility in the upper layers. At least two factors can be cited that contribute to a decrease with height in the reliability of kinematic vertical motion estimates. First, it is well known, and has been demonstrated analytically (Duvechal, 1962; Reiter, 1963), that wind measurements from a GMD-1A system deteriorate with decreasing elevation angles, which are the combined consequence of strong winds and sounding duration (indirectly height). Second, as a result of vertically integrating (5), errors or shortcomings in the wind analyses in individual layers tend to accumulate with height.

An approach used by some authors to improve estimates of \( \omega \) obtained from (6) is to solve the second-order form of (5) resulting from differentiation with respect to pressure (see, e.g., Lateef, 1967). After choice of appropriate boundary conditions for \( \omega \) at the top and bottom of an atmospheric column, the solutions obtained are equivalent to adjusting the divergence at all levels by a constant. In view of the indicated character of wind measurement and analysis errors, application of the technique would appear to lead to unwarranted alteration of kinematic properties in low levels where wind assessment should be more reliable.

Adiabatic vertical motions evaluated in the dry stable layers of the lower stratosphere were found to be an order of magnitude smaller than the kinematic estimates obtained at the same levels for the data analyzed here. Based on this condition and the judgment that kinematic vertical motions obtained from (6) were acceptable in the low tropospheric layers, an adjustment technique (O'Brien\(^5\)) is prescribed for the horizontal divergence, which varies with height, such that the vertical velocity satisfies some independently determined boundary conditions for the top and bottom of an atmospheric column, while achieving most of the compensation at highest levels where original computations are subject to greatest doubt.

The adjusted divergence \( D' \) and vertical motion \( \omega' \) are expressed as

\[
D'_k = D_k + \frac{k}{M} (\omega_K - \omega_T),
\]

\[
\omega'_p = \omega_p - \frac{k}{2M} (k + 1),
\]

where

\[
M = \sum_{k=1}^{K} k = \frac{1}{2} K(K + 1).
\]

Here \( K \) is the total number of layers and \( k \) again denotes individual layers. In the weighting factor on the right in (7) and (8), \( \omega_K \) is the unadjusted vertical velocity at 100 mb as obtained from (6), and \( \omega_T \) is the specified boundary value at the same level. As noted above, vertical motions evaluated near 100 mb from the adiabatic approach were found to be an order of magnitude smaller than \( \omega_K \). Furthermore, \( \omega_K \) averaged over the grid indicated essentially zero net vertical motion at 100 mb, for the region as a whole. For these reasons a choice of \( \omega_T = 0 \) was considered justified. Multiplication of (4) by \( \delta p \) provides an estimate of \( \omega'_p \) at the first constant pressure surface above ground (950 mb).

Distribution of \( \omega' \) near the level of maximum upward motion is shown in Fig. 11, superimposed on the PPI radar intensity contours as they appear at 1500 CST.

\(^4\) Finite difference schemes used in computing divergence and associated kinematic vertical velocity are similar to those described by Fankhauser (1965).

Reasonable qualitative agreement is demonstrated between the location of active convection and the configuration and orientation of strongest vertical velocity. Profiles of $D$, $D'$, $\omega$, and $\omega'$ given in Fig. 12 represent the average conditions over the grid interval containing the largest and southernmost storm echo in Fig. 11. As prescribed by the adjustment technique, the most significant change in the profiles occurs at highest levels. The maximum adjustment on $D$ in the 150–100 mb layer is equivalent to a mean wind speed discrepancy through this layer of 0.8 m sec$^{-1}$ over a distance of 20 km. The character of detailed hodographs presented by Danielsen and Duquet (1967) indicates that errors of this magnitude could easily accrue from averaging procedures and analytical techniques used to generate the wind components dealt with here.

Analyses appearing on the stacked and skewed horizontal planes in Figs. 13 and 14 provide three-dimensional impressions of the divergence and vertical motion fields that are resolved after operations (7)–(9) are applied to the original kinematic estimates obtained from (3) and (6). Levels selected for presentation correspond as closely as possible to the features of greatest significance appearing on the $D'$ and $\omega'$ profiles in Fig. 12.

In interpreting the results it should be recognized that, in the interval between the measurement of an event and its entry into the computational array, the data involved in the diagnostic treatment have been subjected to a considerable amount of subjective and objective smoothing in both the time and space domains. Gradients and scales of motion relevant to the kinematics and dynamics of individual thunderstorm circulations, if detected at all, are therefore undoubtedly filtered or damped further in the data-handling process. Nevertheless, there are medium-scale features retained in the analyses which relate well to the evolution of convective activity as shown in Fig. 2. Perhaps the most striking aspect is the noticeably relaxed character of both the divergence and vertical motion fields early and late in

![Fig. 11. Adjusted vertical velocity $\omega'$ (mb sec$^{-1}$) from (8) at 450 mb, 1500 CST 28 May 1967, and radar echoes as in Fig. 2d (1 mb = 10$^{-2}$ mb).](image)

![Fig. 12. Vertical profiles of $D$, $D'$, $\omega$, and $\omega'$, 1500 CST 28 May 1967, derived by averaging four grid-point values surrounding southernmost storm in Fig. 11.](image)
Fig. 13. Hourly three-dimensional distribution of adjusted divergence $D' \left(10^{-4} \text{sec}^{-1}\right)$ computed from (7). Dimensions of horizontal array, 200 km $\times$ 200 km. Note different contour interval in lowest layer.
Fig. 14. Hourly three-dimensional distribution of adjusted vertical velocity $\omega' \, (\mu \text{b sec}^{-1})$, computed from (8). Horizontal dimensions, 200 km x 200 km.
the synoptic sequence and at all levels, at times when thunderstorm activity was virtually nonexistent.

Convergence at the 925-mb level gradually intensifies with the penetration of the frontal zone into the northwest portion of the grid as shown in Fig. 2. Note that the analyzed contour interval in this layer is four times greater than in the other layers aloft. Between 1300 and 1600 CST, during the period of most intense and widespread convective activity, the computed convergence near the front exceeds $-2 \times 10^{-4}$ m s$^{-1}$. Although a maximum of more than $-8 \times 10^{-4}$ m s$^{-1}$ is resolved at 1400 CST, the strongest radar echo development occurred between 1500 and 1600 and is apparently responsive to the midtropospheric divergence which develops with the passage of the wind speed maximum between 500 and 400 mb which was discussed in the preceding section and shown in Figs. 5 and 8.

The layer of divergence appearing at 725 mb persists throughout the observational period over the northwest sector of the grid and is strongest near the top of the moist layer (see soundings in Fig. 3, and vertical profiles in Fig. 12). This is apparently a synoptic-scale feature which seems to inhibit significant convective developments, except in preferred, strongly convergent regions at lower levels along the frontal zone. It is difficult to separate cause and effect when comparing changes in the divergence pattern at 225 mb to the storm formation and dissipation cycle illustrated in Fig. 2. Based on the continuity of features demonstrated in Fig. 6, however, and their agreement with the wave characteristics shown in Fig. 7c for the synoptic scale, it appears as though storm formations are more a response to, rather than a cause of, the divergence resolved at this level.

In Fig. 14, the single feature best related to the distribution and intensity of precipitation echoes is the field of vertical motion at 450 mb. Distinct centers of upward motion with magnitudes of 20–25 mb sec$^{-1}$ are, at all times, well correlated with the location of the strongest thunderstorms. During periods of greatest upward motions, compensating regions of subsidence appear at middle levels (700 and 450 mb) on the flanks and to the rear of stormy areas.

Panofsky (1951) has pointed out, since the sum of the velocity differences evaluated in (3) is usually on the order of 1 m sec$^{-1}$, that regardless of the scale over which differences are taken, the order of magnitude of computed divergence is apt to be strongly influenced by the size of the computational area. It is appropriate, therefore, to compare values shown in Fig. 13 with others from similarly dense sounding networks. Matsumoto et al. (1967) report maximum convergence in the lowest layers along bands of wintertime convection of $-3 \times 10^{-4}$, while Kreitzberg (1968) found convergence near an occluded front on the order of $-4 \times 10^{-5}$ m s$^{-1}$. Even though these values relate to comparable sounding networks, the question of the dependence of computed divergence on observational density and frequency may still be extant since the convergence shown at 925 mb in Fig. 13 and that reported by Matsumoto are derived from dense surface network winds, while Kreitzberg's computations at lowest levels are based on winds taken from the more widely spaced mesonet network rawinsondes.

Using kinematic methods, Elliott and Hovind (1964, 1965) arrived at vertical motions associated with Pacific cyclones entering Southern California by applying time-space conversions to serial rawinsonde data after considering circulations to be steady within systems moving with a fixed translational velocity. In so doing, their data density approaches that of the networks mentioned above and their reported maximum upward and downward motions compare closely with those shown in Figs. 11 and 14.

7. Quantitative assessment of data credibility

Since inherent limitations in both the accuracy of the conventional rawinsonde system and the methods adopted to synthesize its output become more critical as the space and time intervals decrease, it is necessary to establish the quantitative reliability and consistency of upper-air analyses at the mesoscale. It has been demonstrated in the preceding section that after compensating for apparent errors involved in the measurement and analysis of wind, the three-dimensional distribution of divergence and vertical motion demonstrates reasonable qualitative agreement with the location and intensity of thunderstorm developments. The importance of latent heat contributions to the thunderstorm energy budget (McLaughlin, 1967) dictates, however, that a network designed for the study of convective dynamics must also be capable of resolving moisture and temperature distributions to a sufficient degree of accuracy. An analytical and data consistency check, involving both kinematic and thermodynamic properties of moist air, is therefore adopted.

The first law of thermodynamics can be written as

$$dh = c_p dT - \alpha d\rho,$$

(10)

where $c_p$ is the specific heat capacity at constant pressure and $\alpha$ specific volume. If the differentials are taken with respect to time and the individual temperature change is expanded into its components, the resulting thermodynamic energy equation is

$$\frac{dh}{dt} = \epsilon_c \left[ \frac{\partial T}{\partial t} + \frac{\partial T}{\partial x} + \frac{\partial T}{\partial y} \omega \left( \frac{\partial T}{\partial p} - \frac{\alpha}{c_p} \right) \right],$$

(11)

which expresses the rate of diabatic heating for a unit mass of air.

An alternative estimate of nonadiabatic heating near and within thunderstorms is obtained by assuming that
the major diabatic contribution derives from the condensation of water vapor. We first consider the individual change of mixing ratio $r$ in expanded form, i.e.,

$$\frac{dr}{dt} + \frac{\partial r}{\partial x} + \frac{\partial r}{\partial y} + \frac{\partial r}{\partial p} = 0.$$  

(12)

Multiplication of (12) by the latent heat of condensation $L$, and use of the mass continuity equation, leads to

$$\frac{dh}{dt} = -L \left[ \frac{\partial r}{\partial t} + \frac{\partial (\omega r)}{\partial x} + \frac{\partial (\omega r)}{\partial y} + \frac{\partial (\omega r)}{\partial p} \right],$$  

(13)

which also expresses the heating rate per unit mass. Theoretically, (13) should tend to overestimate heat release since all converging water vapor is presumed to condense. On the other hand, other energy transfer processes such as radiation, friction and conduction may work in such a way as to partially compensate for the overemphasized contributions of water vapor condensation.

For comparative purposes, the right sides of (11) and (13) are evaluated on the layered mesh described in preceding sections. To minimize truncation and preserve the maximum information content, time differences are taken from the hourly input, and time-centered averages of all other terms are derived. Integration from the surface to 100 mb then gives the net diabatic heating for a vertical air column of unit cross-sectional area on the half hour. The results are shown in Fig. 15, where $Q_e$ and $Q_m$ represent the summation through 18 layers of (11) and (13), respectively.

At most synoptic times, the point-to-point quantitative agreement between the independent energy assessments is gratifying. As anticipated, the net positive heating contribution calculated from both methods is found to maximize in areas occupied by the strongest radar echoes.

In the practical evaluation of (11) and (13), vertical motion is taken from (8) which includes adjustments for the assumed source and character of wind errors. Wind components appearing in other terms do not, however, include error compensation. In view of the results in Fig. 15, it would appear, therefore, that the failure of (6) to produce realistic kinematic vertical motions at high levels derives mainly from the cumulative effect of unresolved nonlinearities in the wind analyses, rather than from shortcomings in the measurement of the winds. It should be noted that inadequacies in the evaluation of wind at higher levels have a minimal effect in (13), since most of the water vapor convergence takes place in the moist lower layers where confidence in wind resolution is greater (Bradbury, 1957; Fankhauser, 1965).

In the foregoing discussion it has been presumed that temperature and moisture gradients were dealt with effectively. The breakdown in the comparative energy estimates at 1430 CST in Fig. 15 suggests that this is not completely so. Since the $Q_m$ pattern agrees well with the location and intensity of radar echoes at this time, we must look with suspicion upon the definition of moisture content, known to be a highly variable atmospheric constituent and the least reliable aspect of rawinsonde measurements.

As Fig. 2c indicates, the CHK and WAT ascents scheduled at 1400 CST were released within mesoscale high pressure systems associated with decaying thunderstorm complexes. These are characteristically colder and have lower specific humidity than their surroundings (Newton, 1950; Fujita, 1959). Undue weight given to these soundings in the analyses apparently leads to unrepresentative moisture gradients. According to (13), the local increase in moisture, demonstrated between 1400 and 1500 CST on the CHK series in Fig. 4, tends to compensate the positive contributions that advective terms make toward increasing $Q_m$. Doubt is further cast on the resolution of moisture distributions by the fact that, contrary to other times, the water vapor flux convergence evaluated at 1430 CST fell far short of accounting for areal averages of observed point rainfall at mesonetwork surface stations.

Although the results in Fig. 8 serve to establish the internal consistency of the rawinsonde data and lend credence to the applied analytical techniques, the physical significance of the energy estimates remains in question. Rainfall measured by a dense network of surface gages provides a basis for examining the physical relevance of the computations. Precipitation rates recorded during the direct passage of the largest storm in Fig. 2d correspond to an associated latent heat release rate per unit storm area that is one order of magnitude larger than the largest positive heating indicated in Fig. 15 at 1530 CST. On the other hand, by comparing Fig. 15 with Figs. 2d and 2e, it can be seen that the ratio of the area covered by positive diabatic heating at 1530 CST to that occupied by radar echoes is on the order of 10 to 1. Furthermore, when the total heat production associated with the entire positive heating area is compared to the total latent heat production from all thunderstorms within the region enclosed by the computational grid at the same time, their orders of magnitude are comparable. It appears, therefore, that although the analyses fail to resolve the details of kinematic features and energy processes relating to singular convective circulations, the net effect of storm presence is realized within the overall region containing the thunderstorms.

8. Conclusions

The results of these analyses are strongly influenced by the filtering processes applied in the various steps of data reduction and synthesis. In addition, the scales and magnitudes of the resolved kinematic and dynamic
Fig. 15. Successive hourly comparisons of net diabatic heating (10^6 cal m^-2 sec^-1) resulting from vertical integration of Eqs. (11) right and (13) left. Grid dimensions are the same as in Fig. 2.
features are clearly a function of the original sounding frequency and spacing. For a typical Oklahoma thunderstorm, translating within the 9- and 10-station rawinsonde network, detection capability, as discussed in Section 2, should approach 80%. Failure to diagnose known characteristics of individual thunderstorm circulations demonstrates that high detection probability does not necessarily lead directly to the definition of relevant convective features at the same scale.

Definability can be assessed from another point of view. Basic sampling theory says that for a given sampling interval \( s \), the smallest scale-length resolvable will have dimensions \( 2s \). With the average station spacing of 80-90 km and a sounding interval of 90 min, the unambiguous resolution resulting from a purely objective data handling approach would be restricted to features greater than 175 km in length or 3 hr in duration.

Comparison of storm size and location, station spacing and sounding interval, with the distribution and intensity of the analyzed divergence, vertical motion and thermodynamic energy fields indicates that the level of resolution lies considerably above that predicted by sampling theory and somewhat below that expected from the standpoint of detection probability. Retention of features that are smaller than those relating to sampling theory is permitted on substantial subjective grounds based on the four-dimensionally quasi-continuous nature of network serial ascents. At the same time the objective and subjective smoothing involved in the data reduction and analyses, combined with the basic observational limitations in time and space, does not permit reliable extension of scales downward to include singular thunderstorm processes. Measurements pertaining to internal thunderstorm circulations were not likely to be obtained in any case since balloon release into centers of active convection was avoided whenever possible.

Although it is unlikely that the sounding system, in itself, will lead to a complete diagnosis of thunderstorm dynamics, synthesis of network data in the manner outlined here appears to provide a reliable and internally consistent framework within which more detailed and smaller scale thunderstorm observations, as obtained from radar, aircraft and surface mesonetwork stations, may be viewed. Finally, the analyses seem to provide the logical link between the synoptic and mesoscales which would permit investigation of the role that thunderstorm processes play in the general circulation of the atmosphere.

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