DYNAMICAL INTERACTIONS BETWEEN LARGE CONVECTIVE CLOUDS AND ENVIRONMENT WITH VERTICAL SHEAR

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ABSTRACT

The physics of convective clouds depends not only upon thermodynamic processes but also upon the interactions with the environmental wind field. When vertical shear is present, a hydrodynamic (nonhydrostatic) pressure field is induced by relative motions near the boundaries of a large convective system which does not move with the ambient winds. This tends to make the cloud tilt away from the vertical, but at the same time vertical gradients of hydrodynamic pressure aid the formation of new convection. Expressions are derived for the accelerations concerned.

Quantitative estimates based upon experimental analogies show that vertical accelerations due to the induced pressure field may be of the same order as those associated with ordinary buoyancy forces. These account for the self-propagating nature of squall-line-type thunderstorms and their ability to continue at night when thunderstorms in stagnant air masses tend to die out. The hypothesis accounts for the observed movements of this type of rainstorm, somewhat toward the right of the winds in the convective layer, and is consistent with the preference of severe phenomena, such as tornadoes, for a particular flank of a rainstorm. Distinction is made between large and small clouds as affected by shear.

1. Introduction

Organized convective systems such as squall lines are generally associated not only with thermodynamic instability but with strong winds aloft as well [8; 10; 19; 25]. On the other hand, it is well known that with pronounced vertical shear the tops blow off small convective clouds and, being thus separated from supporting low-level updrafts, generally dissipate before reaching an advanced stage of development. This difference in behavior, and the tendency for squall lines to dissipate on moving into regions of weaker winds aloft [23], suggests that continuous regeneration of the large systems must be encouraged, rather than inhibited, by strong vertical shear.

This paper will be concerned mainly with certain internal mechanisms governing the behaviour of squall-line-type thunderstorms after they have formed rather than with their initial formation. Several characteristics distinguish such convection from ordinary “afternoon” thunderstorms. Among these is their strongly asymmetric character, with concentration of intense convective phenomena on a favored flank; connected with this is a marked tendency for new growth in a favored direction from the existing thunderstorm mass. Another feature is the ability of such thunderstorms to persist and even increase during the night hours after ordinary convection characteristically dies out. The present paper amplifies with quantitative evaluation of the processes involved, an explanation for this behavior given earlier by one of the writers [19].

2. Behavior of large rainstorms

The conclusions to follow have been generalized from a detailed analysis of surface and upper-air data for all severe convective situations (about twenty) in March and May 1949. For each case, hourly analyses of rainfall amounts were made throughout the period of interest, using data from the Weather Bureau Hydro-climatic Network of recording rain-gauge stations [31].

Convection associated with severe-weather phenomena sometimes takes the form of well-organized lines and sometimes occurs with isolated storms showing no particular organization. In either event, the patterns are characterized by distinct concentrations of moderate to heavy rains, typically 20 to 50 mi across, which will hereafter be called rainstorms. Individual rainstorms maintain their identities for many hours and in most cases can be tracked with confidence even though the raingauge stations are often 20 to 40 mi apart.

Use of hourly rainfall amounts must, of course, result in distortion due to storm movement, as well as smoothing of details of intensity. Such analyses adequately portray the broad-scale features, suitable for studying the storms in their synoptic settings. Detailed surface structures of thunderstorms and squall lines, best studied by use of radar and special networks, are well known through the works of Suck-
storf [28], Harrison and Orendorff [13], the Thunderstorm Project [30], Brunk [5], Williams [34], Fujita, Newstein, and Tepper [9], and others; large-area composites of radar photos have been constructed by Ligda et al. [14].

A good example of the history of a squall line is afforded by the situation of 20–21 May 1949, the first of three outbreaks of severe weather occurring on successive days with the same cyclone. The relation of the squall line at full development to the frontal system and flow patterns aloft is shown in fig. 1. The pronounced veering of wind with height and association with lower- and upper-level jet streams are characteristic.

Fig. 2 shows the isohyetal patterns at various stages. From a few relatively small rainstorms near the cold front between 1400 and 1600 CST, the squall line grew to maximum extent and intensity by about 2200 CST, gradually shrinking after that time but lasting through the next day. As shown by the storm tracks in fig. 3 and the total rainfall in fig. 4, individual rainstorms mostly showed a northeastward drift while the line as a whole moved eastward. Thus a storm located in the southern part at one time moved a few hours later to the north end and gradually dissipated. Most of the rainstorms seen north of the warm front in fig. 2 had formed initially in the warm sector, reaching maximum intensity 4 or 5 hr after initiation, just after passing the warm front. Fig. 3 shows that most of the rainstorms lasted 8 hr or more, the longest track being 17 hr.

Initial formation of the squall line occurred very near the western edge of the region of potential instability (fig. 5) when the cold front first swept into the low-level moist tongue. Maximum development was reached near midnight when the rainstorms passed through the tongue of greatest instability.

It is clear that the general development took place...
in the setting created by large-scale synoptic processes [2; 17; 29; 32]. The convective systems, although feeding upon the environment created by such processes, behaved in a manner suggesting that they were self-propagating. This is shown by the fact that the squall-line rain area finally moved more than 1000 km east of the cyclone center (see, also, striking examples given by Brunk [5]). Also, the rainstorms were not simply carried by the winds, the tracks being rather on the average about 20 deg to right of the mean wind in the convective layer (see [20; 21] also).

3. Mechanism of rainstorm propagation

Character of circulation. Radar observations show that the individual rainstorms in a squall line generally consist of a large number of thunderstorm cells. Since the life of an individual cell is only about a half hour, continuous formation of new cells is necessary if the convective system is to persist for a long time. Harrison and Orendorff [13] concluded that squall lines are maintained through successive triggering of new convection by lifting of unstable air over a "pseudo-cold front" (hereafter called squall front) at the boundary of the rain-cooled air from existing thunderstorms. The presence of such a squall front is borne out by the cross section in fig. 6. The distribution of wet-bulb potential temperature, together with other evidence [19], suggest a circulation of the general nature indicated, the arrows representing average conditions of the flow with respect to the moving system. On the advancing (right-hand) side, the flux is indicated as upward although intermingled strong downdrafts can be presumed to be associated with the heavy rain in this region. Bergeron [3] (see especially his figure 8h) has postulated a circulation in squall lines in most respects similar to that shown here.

If we suppose that, when a mass of air in low levels encounters the squall front, the kinetic energy of relative horizontal motion is (through a process specified later) converted into energy for lifting an equivalent mass, the energy transformation may be represented by

$$\delta \left( \frac{V^2}{2} \right) \rightarrow \left( \frac{\Delta T}{T} \right) \delta z.$$  

In that case, air whose mean virtual temperature during lifting is $\Delta T$ colder than the undisturbed en-
environment could be lifted a distance $\delta z$ against gravity. Taking the initial relative horizontal speed $V_0$ to be 15 m sec$^{-1}$ (valid for the case in fig. 6), then if $T = 280K$ and $\Delta T = 4K$, $\delta z$ would be 800 m. Thus, the kinetic energy given up when the air encounters the squall front is of the order required for lifting potentially cool air far enough to set off the instability in many cases.

The available energy from this process depends upon the relative horizontal motions of the squall front and of the warm air which it encounters. It will be shown later that in a typical case only about half this relative speed, and correspondingly only part of the kinetic energy of relative motion, results from the ordinary outflow due to thunderstorm downdrafts spreading out in surface layers. The effectiveness of the squall front is due to a combination of this process and another process described next.

Vertical transfer of momentum and thunderstorm propagation. Considered as a whole, a large rainstorm consists of a mixture of updrafts and downdrafts, and a continual vertical exchange of horizontal momentum takes place within the cloud system. If the wind increases with height, as in fig. 7, this would lead to faster speeds in low levels, and slower speeds in upper levels, than are present in the wind field in which the cloud is imbedded.

If fig. 7 were a cross section through an unbroken squall line extending indefinitely in the direction normal to the figure, the relative motion would imply horizontal convergence and divergence as indicated. In that case, vertical motions would be induced as shown near the cloud boundaries in the figure. Provided the air mass is potentially unstable and there is enough energy for lifting, growth of new convection would be favored on the downshear side.

For the case of a squall line made up of rainstorms separated by clear spaces or inactive cloud, the discussion above is not conclusive since the air might be imagined to divide and pass around the sides of an individual rainstorm without actual convergence and divergence. The consequences in that case may be inferred from hydrodynamics experiments relating to flow around obstacles. Although rainstorms have
Fig. 2. Continued. (See caption on page 484.)

Fig. 3. Tracks of individual rainstorms, squall line of 20–21 May 1949. Dashed lines connect rainfall centers observed at same time (date/hour below); dots on tracks show positions at successive hours. In circles at 3-hr intervals, intensity: open circle, maximum 1-hr rainfall less than 0.50 in; quarter filled, 0.50 in or more; half filled, 1.00 in or more per hr.
irregular shapes, it will be necessary for a tractable discussion to assume that they are analogous to circular cylinders acting, at any individual level, as obstacles in the wind field.

Fig. 8 shows the general character of the flow around a cylinder in a viscous fluid. Measurements show that a positive hydrodynamic pressure (induced departure from hydrostatic pressure) is found on the upstream side with respect to relative motion, with a negative departure on the downstream side.

If one considers the relative motions in fig. 7, the signs of the hydrodynamic pressures at rainstorm boundaries would be as shown in the parentheses. A vertical gradient of nonhydrostatic pressure would be present, and, because of differences between the gravity force (corresponding to hydrostatic pressure gradient force) and the vertical pressure gradient force at cloud boundary which is no longer hydrostatic, the induced vertical motions would be the same as indicated before.

4. Quantitative evaluations

Relative motions and induced pressure field. To appraise the induced pressures, it is necessary first to estimate the relative motions between in-cloud and ambient air. In the following, $V_0$ will denote the in-cloud wind velocity, $\bar{C}$ the average at a given level over the whole rainstorm, and $C$ the average at all levels. The latter is assumed to correspond to the mean motion of the cells of which a rainstorm is composed at a given instant, excluding effects of propagation (cell growth or dissipation at the edges). The wind velocity in the undisturbed environment is $V_s$, $\bar{V}$ being the average with height. The relative velocity

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![Fig. 4](image.jpg)

**Fig. 4.** Total rainfall for storm series of 20-21 May 1949 (only rain associated with squall line in preceding figures). Outer isohyet, 0.01 in, succeeding isohyets at intervals of 0.50 in; hatching, rainfall in excess of 1.00 in; cross-hatching, more than 2.00 in.

![Fig. 5](image.jpg)

**Fig. 5.** Stability index (500-mb temperature minus temperature of parcel lifted to 500-mb level from 850 mb; negative numbers indicate potential instability), and locations of rainstorms associated with squall line. First and third charts are interpolated, and are intended only to show pattern of stability.
between undisturbed environment and cloud column is $V_e = V_c - C$; subscripts $u$ and $l$ denote upper and lower levels.

Fig. 9 shows $V_e$ at various levels, for the average of all cases at locations of rainstorms in the situation discussed earlier. $C$ was estimated by a method suggested by data of the Thunderstorm Project on the movements of small radar echoes in cases where propagation was absent [30, p. 108–109]. These moved in the direction of the mean wind in the layer 2000 to 20,000 ft. However, they moved at a speed less than the average wind speed in this layer when the upper winds were strong, presumably because the air within the original up-drafts is drawn initially from the low-speed subcloud layer. Specifically, when the mean wind speed was 37 kn as in fig. 9, individual cells moved at 10 kn less than $V_e$. This difference was accepted for $V_e - C$ in fig. 9. Mean in-cloud motions at individual lower ($C_l$) and upper ($C_u$) levels were estimated on the basis of the following analysis.

The mean motion of the air within a slice at a given level within a rainstorm can remain unchanged with time, if the acceleration of air within it (due to external pressure forces) is offset by a continual import or export of momentum into or out of that slice, through its boundaries. If one considers only vertical
transport, this balance would exist if 

$$ \frac{dC}{dt} = \left( \frac{\partial V_e}{\partial z} \right)_m \approx \frac{\partial C}{\partial z} $$

(1)

In the last approximation, deviations of \( w \) from \( w_m \) and of \( V_e \) from \( C \) are neglected.

Let the hydrodynamic pressure \( w = (\rho - \rho_a) \) be the deviation of the actual pressure \( \rho \) from the hydrostatic pressure \( \rho_a \), arising from interactions between the cloud obstacle and the ambient wind field. The total external pressure force acting horizontally across a unit slice of the rainstorm is then \( D \Delta w_m \), if \( \Delta w_m \) is the mean difference upstream minus downstream pressure (fig. 8) averaged across the diameter \( D \) of the rainstorm and directed along \( V_e \). Equating this to the mass \( (\rho \pi D^3/4) \) times the mean acceleration of the

air within the slice, it follows that

$$ \frac{dC}{dt} = \frac{4 \Delta w_m}{\pi D}. $$

(2)

From (1) and (2), the condition for the mean horizontal velocity at a given level in a rainstorm to remain unchanged with time would be given by

$$ \frac{\partial C}{\partial z} = \frac{4 \Delta w_m}{\pi D}. $$

(3)

In that case, the air at a given level within the rainstorm could be maintained at a velocity different from the environmental wind, as in fig. 7.

In fig. 8, the numbers near the boundaries represent a “pressure coefficient” which yield \( w \) if multiplied by the kinetic energy per unit volume \( (\rho V_e^2/2) \) corresponding to a speed \( V_e \) of the undisturbed basic current relative to the obstacle. The particular values correspond to a Reynolds number of 0.3 \( \times 10^3 \), [12, p. 424]. In estimating \( Re = \rho V_e D/\mu \), the follow-

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2 Assumption of a circular cylinder as shape of the cloud column does not affect the pressure at upstream stagnation point, which is the most critical in later evaluations. For other shapes, pressure coefficients vary most for the lateral flanks; e.g., they are larger for elliptical cylinders elongated across the flow. The extent to which a cloud column presents an impenetrable obstacle cannot be established without observation. Such observations as do exist indicate a relatively sharp leading edge for the low-level gusts and for the temperature break, even at a considerable distance above ground (e.g., the case in fig. 6).
ing values were used: \( \rho = 10^{-3} \text{ gm cm}^{-3} \), \( V_r = 10 \text{ m/sec}^{-1} \), \( D = 30 \text{ km} \), and eddy viscosity \( \mu = 100 \text{ gm cm}^{-1} \text{ sec}^{-1} \).

In the range considered, a tenfold variation of \( Re \) is associated with only a 30 per cent variation in the pressure coefficients; it therefore suffices to consider \( \Delta \omega_m \) as being proportional to \( \rho V_r^2/2 \) for storms of all sizes. Then (3) can be written as

\[
D \propto \frac{V_r^2}{w \frac{\partial C}{\partial \varepsilon}}.
\]

(4)

Now, for a given environmental shear, the relative velocity \( V_r \) near cloud base or top will be larger as the in-cloud shear is smaller (fig. 7). Thus, for a given value of \( w \), a decrease of \( \partial C/\partial \varepsilon \) implies an increase of the right-hand term. It is thus apparent from (4) that, other things being equal, the larger the diameter of a cloud system the smaller will be the average vertical shear within the cloud and the greater can be the relative motions between cloud and environment as well as the induced hydrodynamic pressures. Thus, the tendency for the tops of cloud columns to be bodily blown away from the bases, inhibitory to the development of small clouds [11; 15; 24] is less pronounced in large clouds. This is intuitively evident from the fact that, with increasing cloud diameter, the mass to be accelerated increases more rapidly than the area presented broadside to the wind.

Observations given by Byers and Battan [7] show that with strong shear even relatively small radar echoes (\( D = 5 \text{ km} \)) change their tilt from the vertical at only 50 to 75 per cent of the rate indicated by the ambient wind shear. For a rainstorm several times this diameter, it appears reasonable to suppose that the average in-cloud shear could be as little as 1/3 that of the environmental winds.

With that assumption, \( C_t \) and \( C_u \) in fig. 9 have been taken to be displaced from the mean cell velocity \( \bar{C} \) by 1/3 of the difference between \( \bar{C} \) and the wind velocity at the appropriate levels. The relative motion vectors are shown for the middle of the lower-level moist layer and for the upper part of the cloud at a level where vertical motions are likely to be vigorous.

Table 1 gives the values of \( \rho V_r^2/2 \) for various levels from the ground (960 mb) up to 300 mb. The hydrodynamic pressure \( \omega \) (in millibars) to be expected on any particular part of the rainstorm boundary can be determined by multiplying the value in the last column of table 1 by the pressure coefficient as shown in fig. 8, having regard for the direction of the relative wind.

At 500 mb, from the values in table 1 and the detailed pressure distributions given in Goldstein [12], \( \Delta \omega_m \) would be about 0.4 mb. Taking 1/3 the environmental shear (fig. 9) of 18 m sec\(^{-1}\) in 6 km between 700 and 300 mb, \( \partial C/\partial \varepsilon \approx 10^{-3} \text{ sec}^{-1} \). If \( D = 30 \text{ km} \), (3) would be satisfied if \( \omega_m = 2.5 \text{ m sec}^{-1} \), about a third of the median draft velocity measured by Thunderstorm Project aircraft. In this evaluation (equation 1), only mean values have been considered. The probable positive correlation between \( w \) and \( \partial V_r/\partial \varepsilon \) in individual updrafts and downdrafts in the upper part of the cloud [30, fig. 90] suggests that momentum transport in these “eddies” would decrease the required size of \( \omega_m \) computed above.

For the 910-mb level, again taking 1/3 of the ambient shear, \( \partial C/\partial \varepsilon = 3 \times 10^{-3} \text{ sec}^{-1} \). From (3), \( \omega_m \) would be \(-0.5 \text{ m sec}^{-1} \) (negative because \( \Delta \omega_m \), along \( V_r \) in fig. 9, is nearly the opposite direction from \( \partial C/\partial \varepsilon \), along the hodograph toward the right). This is a reasonable value for the mean downdraft velocity, requiring a mean outflow through the rainstorm boundary, of 7.5 m sec\(^{-1}\) in the lowest 500 m.

It is concluded that realistic values of vertical motion can account for enough transfer of horizontal momentum to establish and maintain large (10 m sec\(^{-1}\)) differences between mean motion of in-cloud air and the ambient winds, at individual levels in the upper and lower parts of a large rainstorm.

It will be noted from (3) that the more vigorous the vertical motions, the more readily large horizontal accelerations from outside forces can be counteracted and the greater \( V_r \) can be. Strong in-cloud shear would be favored by the attenuated vertical motions near the top of an established convective system, and the upper parts (e.g., the anvil cirrus) may be readily carried away from the main body of the cloud (being replenished by cloudy air from below).

Vertical accelerations. It will now be shown that the hydrodynamic pressure field induced by relative motions of the above magnitude is significant for convective processes. The vertical acceleration may be written (\( \rho = \rho_h + \omega \))

\[
\frac{d\omega}{dt} = - \frac{1}{\rho} \left( \frac{\partial \rho_h}{\partial \varepsilon} + \frac{\partial \omega}{\partial \varepsilon} \right) - g,
\]

where \( \rho \) is the density of the air column under con-
sideration, which may or may not be within cloud.
If the undisturbed environment of that column is
assumed in hydrostatic balance and has density \( \rho_e \),
then \( \partial \rho_e / \partial z = -g \rho_e \) and, identically,
\( \partial \sigma / \partial z = -g \rho_e \partial \omega / \partial \rho_h \). When these substitutions are made into (5),
\[
\frac{dw}{dt} = g \left( \frac{\rho_e}{\rho} - 1 \right) + g \frac{\rho_e \partial \sigma}{\rho \partial \rho_h}.
\] (6)

If we write \( T = T_e + \Delta T, T \) being the virtual temperature in the column considered and \( T_e \) that of the undisturbed environment, the equation of state gives \( (\rho_e / \rho) = (1 - \sigma / \rho)(1 + \Delta T / T_e) \), which on substi-
tution into (6) yields
\[
\frac{dw}{dt} = g \left( \frac{\Delta T}{T_e} - \frac{\sigma}{\rho} + \frac{\partial \sigma}{\partial \rho_h} \right) \approx g \left( \frac{\Delta T}{T_e} + \frac{\partial \sigma}{\partial \rho} \right).
\] (7)

Here, several verifiably small higher-order terms have
been dropped. The second term in the middle expression, though not always insignificant, is in the following application generally much smaller than the third term. The last term states that, if the hydrodynamic pressure decreases upward, an upward-acting force exists independent of the ordinary buoyancy force due to density anomaly, expressed by the first term.

In fig. 10, the relative streamlines corresponding to the basic relative velocities \( V_{tr} \) and \( V_{br} \) in fig. 9 are sketched. Locations of positive and negative pressure are shown near cloud boundaries at individual levels. On the near side of the rainstorm, positive pressure is found in lower levels beneath a pressure deficit in upper levels, favoring upward acceleration. On the left side (in the figure), the pressure distribution would contribute to downward acceleration; on the other sides where a pressure deficit is found at both levels, comparatively weak upward or downward accelerations would be present.

With use of the numerical values in table 1 and fig. 8, the greatest pressure deficit or excess at both upper and lower levels is 0.35 to 0.40 mb. Thus, the hydrodynamic pressure variation through the convective cloud layer could be around 0.7 mb. In the 500-900-mb layer, \( g \partial \sigma / \partial \rho \) in (7) would be 1.8 cm sec\(^{-2}\), giving an increment of upward speed of 5.5 m sec\(^{-1}\) if acting on a rising parcel for 5 min. Thus, new updrafts forming on the boundary of a large rainstorm, in the right sector to benefit from upward-directed hydrodynamic pressure forces, could be expected to have additional vigor beyond that due to buoyancy in the ordinary sense.
For the subcloud layer and lower part of the cloud layer, it is essential to take into account the outflow due to downdrafts spreading out in lower levels beneath active thunderstorms, since at any point on the periphery this would add to the relative motion caused by momentum transfer, as in fig. 11. It has been assumed that an outflow velocity \( V_0 \) of 10 m sec\(^{-1} \) exists at cloud boundary, the in-cloud velocity \( V_c \) at each point on the periphery (dashed wind symbols) being the sum of this radial outflow velocity and the velocity \( C_t \) representative of mean conditions at a lower level of the cloud system. The environmental wind velocity for the 910-mb level is plotted in fig. 11 as \( V_{10} \), vector differences between this and the in-cloud wind being indicated at various points by solid wind symbols. Under the assumed conditions, the maximum relative velocity amounts to 18 m sec\(^{-1} \) into the rainstorm on the SE side, with practically no relative motion at cloud boundary on the NW side.

For \( V_c = 18 \) m sec\(^{-1} \) at 910 mb, \( (\rho V_c^2)/2 \) is 1.7 \( \times 10^4 \) gm cm\(^{-1} \) sec\(^{-2} \), corresponding to a maximum positive pressure, on the SE side of the rainstorm, of 1.7 mb. From table 1, in the levels 700 to 850 mb, \( \psi \) is less than 0.1 mb. If we consider the mean for the layer 910 to 700 mb, on the SE side of the rainstorm \( \partial \psi / \partial p \) would be about (1.6 mb)/(200 mb). From (7), the upward acceleration due to the hydrodynamic pressure gradient can thus amount to 8 cm sec\(^{-2} \), equivalent to the buoyancy acceleration corresponding to a temperature excess of 2.2\( C_t \), or more if a thinner layer is considered.

Triggering of instability by "lifting." Equation (7) shows that in the presence of a hydrodynamic pressure field, a layer of air may be accelerated upward even if it is colder than the environment, so long as

\[
\frac{\Delta T}{T_c} < - \frac{\partial \psi}{\partial p}
\]  

(8)

This makes it possible to draw a distinction between the triggering of new thunderstorm cells by downdrafts spreading out in a stagnant air mass and triggering by the combined effects of this and the momentum-transfer process with vertical shear.

The above analysis suggests that the combined effects may result, on a certain flank of a rainstorm, in a value for \( \psi/\partial p \) around 1.6 mb. According to the criterion (8), if \( T_c = 290 K \), a layer of air 290 mb thick could be lifted until its mean temperature is 1.6°C cooler than the environment. Fig. 12 shows a sounding taken 7 min before passage of the squall line in fig. 6. A pressure differential of 1.4 mb could in this case produce a "lifting" of a deep layer to the degree shown by the "modified" dashed sounding curve and would clearly suffice to set off the available potential instability.

Consider, on the other hand, the same conditions without the presence of vertical shear. With a radial outflow of 10 m sec\(^{-1} \), \( (\rho V_c^2)/2 \) would in lower levels correspond to a hydrodynamic pressure of 0.5 mb (on all flanks of the rainstorm), with a decrease to zero at the top of the outflow layer. The possible degree of lifting would be only one third that of the combined effects above. For the same sounding, as indicated by the "modified" dotted curves, only a relatively thin layer of air could be lifted mechanically.

With greater stability in lower levels (as with nocturnal cooling), the energy available for lifting due to spreading out of the downdraft alone might not suffice to set off new convection. Additional energy made available through transfer of kinetic energy from aloft, in the case of strong vertical shear, would make it much more likely that potential instability could be triggered when surface layers are stable. This distinction accounts in large part for the fact that squall-line-type thunderstorms can continue through the night hours when most thunderstorms tend to dissipate.

Movement of rainstorms. For the case represented by figs. 9 to 11 in which the wind veers with height, growth of new convection is most favored on the right-hand side of a rainstorm, with respect to the mean wind in the convective layer. The total movement of a rainstorm reflects in part a drift of existing cells with the mean wind and in part an anomalous movement to right due to this propagation. In fig. 9, \( R \) is the mean rainstorm velocity obtained from tracks of isohyetal maxima in the synoptic example considered, showing the indicated deviation. Similar results were found in other cases summarized by Newton and Katz [21]; a qualitative check of the twenty cases analyzed for the present study revealed the same tendency.

Point \( M \) in fig. 11 indicates where the greatest relative velocity between in-cloud and environmental wind is found; this nearly coincides with the location of the strongest upward-directed hydrodynamic pressure gradient. \( R \) in fig. 9 nearly coincides with this point, suggesting that the average movement of the rainstorms is governed by movement of the portion of the squall front of the rainstorm where most active propagation takes place. It cannot be verified whether or not the assumed value of \( V_c \) is valid for the particular synoptic case concerned; however, this was based on typical values for the observed mean outflow in various published cases. In Katz’ study [21], the mean movement of rainstorms was about 32 km (8 kn less than \( V_c \)) in a direction 25 deg to the right of the mean wind in the layer 850 to 500 mb (or the 700-mb wind), with a standard vector deviation of 10 kn.
Tornadoes. In the twenty series analyzed, tornadoes were predominantly associated with rainstorms large enough to be followed for several hours by use of rain-gauge network data. As shown by fig. 13, which summarizes occurrences for the case described in section 2, and by fig. 14 which includes results from other cases, the tornadoes were on the whole found in close proximity to an upper-level jet stream. The same was true for the 850-mb level. Thus, tornadoes were associated with the type of rainstorm discussed above, in which in-cloud and ambient-wind velocities may differ appreciably. Most tornadoes occurred with squall-line type situations [25]; however, many occurred with isolated rainstorms or in the early stages before rainstorms joined to form a solid squall line, in accord with an observation by Brunk [6].

In many cases, one or two of the most persistent rainstorms accounted for the majority of severe local storms reported. Long tornado tracks, when these occurred, tended to parallel the rainstorm tracks. Mook [18] found that tornadoes in the United States tend to be associated with a particular 1000-500-mb thickness line (18,600 ft) near the zone of strongest vertical shear. Moreover, he observes that this line (more nearly parallel to the vertical shear vector than to the mean wind direction) lies roughly parallel to the normal tornado tracks in the areas investigated.

Although the exact mechanism is not clear, the
possibility is suggested that tornadoes, and "tornado cyclones" [4], may derive their rotation from mechanically generated eddies due to the presence of the cloud as an obstacle in the wind flow. Hypotheses for tornado formation related to this general notion have been advanced by Wegener [33], Markgraf [16], Fulks [11], and Penn, Pierce, and McGuire [22].

Tornadoes appeared to occur mainly on the right-hand sides of rainstorms, in agreement with findings from radar observations by Stout and Huff [27] and others. Occurrence on that side is consistent with the above conclusions since, presumably, tornadoes should be associated with the most vigorous cumulus convection.

5. Conclusion

The above results appear to furnish explanation for the several features outlined in the Introduction, distinguishing large-scale convection in the presence of strong vertical shear from that without shear. A principal conclusion is that the induced hydrodynamic pressure field can produce vertical accelerations independent of and comparable in strength with ordinary buoyancy forces.

Suitable observations do not exist for direct verification of the mechanism discussed. Aircraft measurements shown by Simpson and Starrett [26] indicate pressure deviations aloft near the boundaries of convective bands in hurricanes (which have much the same character as middle-latitude squall lines), comparable with the hydrodynamic pressures deduced above. Although Simpson and Starrett propose a different explanation, it is possible that these pressure deviations arise from the processes discussed here. Their observations indicate that it should in principle be possible to measure details of the pressure field aloft in the neighborhood of squall lines over water areas (such as the Great Lakes).

Assertion of the general importance of vertical shear does not mean that strong shear is essential in all cases. It was shown that there is enough energy available in the ordinary thermodynamically-caused outflow from thunderstorms to trigger convection when the air is sufficiently unstable in lower levels. The significance of vertical shear is that kinetic energy drawn from the wind field in which a storm is imbedded provides an important additional source which may be put to work in setting off and increasing the vigor of convection.

Effects of shear on small cumulus clouds appear to be different from those indicated for large systems. It was shown that the smaller a cloud the more rapidly it will be accelerated toward the environmental wind velocity. Observations of trade cumuli [15] show relative speeds of 1 to 2 m sec⁻¹; estimated hydrodynamic pressures could produce vertical accelerations equivalent to buoyancy with a temperature excess of 0.1C. What effect this may have must be considered in combination with the effects of entrainment and evaporative cooling at cloud boundaries. In any event, as demonstrated by J. S. Malkus [15] and by Scorer and Ludlam [24], the rapid removal of cloud towers by the wind and the rising of new towers from the same base produces an appearance of building on the upshear rather than the downshear side as in the case of a large rainstorm.

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REFERENCES