Formation of Mesolows or Pressure Troughs in Advance of Cumulonimbus Clouds

LEE R. HOXIT, CHARLES F. CHAPPELL AND J. MICHAEL FRITSCH

Atmospheric Physics and Chemistry Laboratory, Environmental Research Laboratories, NOAA, Boulder, Colo. 80302
(Manuscript received 7 June 1976, in revised form 24 August 1976)

ABSTRACT

Mesoscale lows or pressure troughs have been observed downwind of many mid-latitude cumulonimbus cloud systems, especially those that subsequently produce severe weather such as tornadoes, large hail, or damaging wind gusts. Case studies are presented which link the formation and existence of mesolows or troughs to subsidence warming in the upper troposphere and lower stratosphere. This subsidence warming, which is believed to result from the interaction of the convective cloud with a sheared environment, leads to an instability between the meso-γ and meso-β scales by hydrostatically reducing surface pressures, thereby organizing and increasing the low-level convergence ahead of existing convection.

1. Introduction

The variety and complexity of severe mid-latitude convection have been clearly depicted in analyses presented by Fujita et al. (1956), Magor (1959), Fankhauser (1971) and others. An important characteristic of this type of convection is the quasi-steady vertical circulations in the moist up- and downdrafts. These circulations are typically meso-γ in scale (Orlanski 1975) defined the horizontal scales of various mesoscale processes as meso-γ (2–20 km), meso-β (20–200 km) and meso-α (200–2000 km)]. In the vicinity of these cumulonimbus clouds, larger mesoscale circulations have often been observed to develop. These range from tens to a few hundreds of kilometers and encompass all of the meso-β scale. In turn, the meso-β circulations have been observed to enhance the meso-γ convective circulations, although such scale interactions are not well understood. Severe storms research and forecasting during the past 20–25 years have endeavored to determine the characteristics of these systems as well as the mechanisms responsible for their formation, organization and propagation. More recently, researchers involved in mesoscale numerical modeling have also confronted the problem of explaining the organization and development of mesoscale convective systems (Kreitzberg and Perkey, 1976; Fritsch, 1975).

Several features have already been associated with the triggering and propagation of mesoscale convective systems. These include frontal boundaries, dry lines, downdrafts, vertical shear, horizontal variations in boundary layer heating, and preexisting mesoscale circulations. Frontal boundaries can trigger convection through forced lifting of the boundary layer. Dry lines, frequently found over the southwest plains region, can also act to force deep convection (see, e.g., McGuire, 1962; Rhea, 1966). The importance of moist downdrafts and the subsequent cold outflow in initiating new updrafts has been discussed by Takeda (1965), Marwitz (1972) and others. The role of strong vertical shears in organizing updrafts and downdrafts into a quasi-steady state has been established by Newton and Newton (1959), Browning (1964), Newton (1966), Fankhauser (1971) and others. Horizontal variations in boundary layer heating due to morning cloudiness have been observed to modulate the areal distribution of severe convection (Purdom and Gurka, 1974). The existence of mesoscale mass and moisture convergence prior to convection has been documented by Lewis et al. (1974) and Modahl (1973).

For the most part, however, efforts to understand severe storms have focused on phenomena in or near convective clouds (Kropfl and Miller, 1975) or on preexisting features such as fronts, dry lines, or variations in surface heating. While these preexisting features strongly influence the formation of convection, they do not explain the subsequent mesoscale developments. Unlike synoptic-scale weather systems, where development can be attributed to barotropic or baroclinic instability, the genesis of mesoscale systems is not well defined. Fujita (1959) indicated that cold air produced by evaporation is the major factor in producing the meso-γ “bubble” high. However, the events leading to mesocyclogenesis have received little attention and may involve some instability much different from the classical instabilities defined in synoptic meteorology.

Analyses presented by Magor (1958, 1959, 1971), along with the 3 April 1974 case depicted in Fig. 1, are examples of mesolows with typical diameters of 100–200 km (large meso-β systems). The NHRE case shown in Figs. 2 and 3, plus other analyses by Fujita et al. (1956) and Barnes (1972), indicate smaller systems with
actions between convective clouds and their environment. An important part of these interactions appears to be upper-level subsidence downwind of the existing updrafts. Section 2 presents selected observations made in advance of squall lines or large cumulonimbus clouds. Section 3 considers the implications of downwind subsidence in the development of mesoscale circulations.

2. Observations of organized deep convection and the downwind environment

The evolution of strong convective storms in the central United States can take many forms. Yet meteorologists working in this region note that certain sequential events are repeated in many cases. Typically, sky conditions for the first few hours after sunrise are characterized by broken or scattered stratus clouds. These conditions evolve to broken or scattered stratocumulus or cumulus humilis by late morning or early afternoon as planetary boundary layer heating continues. When lines of cumulonimbus clouds develop, cumulus clouds often dissipate 25–100 km ahead of the strong convection, suggesting that mesoscale circulations exist in advance of these cumulonimbus lines. The mesoscale circulations, apparently related in some way to the existing convection, may act to modify the environment generated through synoptic-scale processes and further enhance the potential for convection (Barnes, 1974).

More substantial evidence that mesoscale perturbations exist ahead of strong convective storms can be
found in detailed analyses of surface pressures and pressure tendencies. Some of the more impressive examples of mesoscale pressure changes occurred with the famous 3 April 1974 tornado outbreak (Hoxit and Chappell, 1975). Major features of their analyses are presented in Fig. 1. The following discussion focuses on events that occurred in Indiana.

A well-defined low-pressure trough with an imbedded mesolow (see Fig. 1a) was located ahead of a squall line some 500–700 km long. One hour difference in altimeter settings showed pressure falls greater than 4 mb h⁻¹ in west central Indiana. These pressure falls were the result of a combination of synoptic and mesoscale processes. An estimate of synoptic tendencies was obtained by space smoothing the 3 h tendencies for the period 1400–1700 CST using an 80 km grid. The synoptic tendencies were then subtracted from the observed tendencies providing estimates of mesoscale tendencies alone. Note in Fig. 1b that mesoscale pressure falls greater than 3 mb h⁻¹ were located 40–70 km ahead of the mesolow. Fig. 1c shows that the tracks of the tornadoes across northern Indiana coincided closely with the track of the mesoscale pressure-fall center. Apparently, most of the severe weather was associated with and encompassed by this mesolow (approximately 100–150 km in diameter).

The mesosystems in Fig. 1 were exceptionally large and intense. Figs. 2 and 3 show a more typical example of a mesolow preceding organized convection. Time series of radar echo position are included along with surface streamlines and mesoscale pressure perturbations for a major hail storm that moved through the National Hall Research Experiment (NHRE) network in northeast Colorado. Mesoscale pressure perturbations were determined by subjectively smoothing the pressure traces at NHRE stations to obtain the “synoptic-scale” pressure tendency, and then subtracting the smoothed traces from the actual ones.

Convective activity intensified as it moved into the surface network with a strong multi-cell storm imbedded in a squall line. Coincident with an increase in the size and intensity of echoes were the development and intensification of a mesolow approximately 15 km east and slightly south of the region of strongest echoes. During the latter part of the period, the mesolow was approximately 40 km in diameter with a maximum pressure perturbation between 1.5 and 2 mb. As the mesolow intensified, the low-level convergence northwest of the mesolow increased noticeably. The mesolow and radar echoes moved ESE at about 20 m s⁻¹, which was roughly 30–40° to the right of, and slightly slower than, the mean wind through the cloud layer.
The tendency for large and severe cumulonimbus cloud systems to move to the right of the mean cloud-layer winds has been known for some time; see, e.g., Hoecker (1957), and Newton and Fankhauser (1964). Several explanations for deviations to the left or right of the mean flow have been offered. Newton and Katz (1958) showed that when the wind veered with height, the maximum relative inflow into the storm occurred on the right flank of the storm, providing for new developments in this region. Fujita and Grandoso (1968) emphasized cyclonic or anticyclonic rotation of the clouds as the mechanism causing the right or left movement, respectively. Charba and Sasaki (1971) analyzed an event with both left and right moving storms in the same general area. They found that major inflow regions corresponded to the positions of surface mesolows, and suggested that the intensity and movement of these mesolows were important controls on the overall storm movements.

Marwitz (1973) and Foote and Fankhauser (1973) presented evidence that significant non-hydrostatic pressure perturbations exist near cloud base in some Northeast Colorado hailstorms. These non-hydrostatic processes, maintained by the dynamics of a vigorous updraft, may create a surface mesolow in the inflow region of the storm. Scale analysis indicates, however, that these pressure perturbations would likely be in the immediate vicinity of the updraft, so that the resulting low would typically be of meso-$\gamma$ scale. Non-hydrostatic processes may have induced the mesolow shown in Fig. 3. However, the increase in size to 40–50 km in diameter by 1600 MDT suggests mechanisms operating on the meso-$\beta$ scale were also involved. Those cases referenced earlier documenting the occurrence of larger meso-$\beta$ systems support this suggestion. Since processes on the meso-$\beta$ scale are largely hydrostatic, surface pressure falls infer a net warming of the atmospheric column; thus, we may consider what processes might produce this net warming.

Direct measurements of warming ahead of squall lines have been made on several occasions. J. Schaefer (personal communication, NOAA, NSSFC Kansas City, Mo.), found middle and upper tropospheric warming of several degrees celsius by examining serial rawinsonde data taken during squall-line passages through the National Severe Storms Laboratory (NSSL) network. Williams (1957) found temperature increases up to 6°C h$^{-1}$ at 300 mb some 1 to 2 h prior to passage of a funnel-bearing squall line at Omaha, Nebr.

Fig. 4 presents a reanalysis of a portion of the case
examined by Williams. We have extended the analysis to 200 mb to better document the temperature and humidity changes in the upper troposphere. Table 1 lists the launch times and maximum ascent levels of those soundings used to construct Fig. 4. An ascent rate of 300 m min⁻¹ was assumed for all soundings. Wind data obtained from the 1730 CST sounding are given in Table 2. Unfortunately, most soundings were terminated at or below 200 mb. Likewise, surface barograph traces in the region of the upper-level measurements, radar photography, cloud top information, and movement and extent of the squall line are not available. Yet, temperature and humidity variations prior to and during squall-line passage provide an excellent example of the modifying effects of organized convection.

The temperature change prior to arrival of the squall line is most pronounced in the upper troposphere. (The increase in temperature at 200 mb was ≈13°C). As the middle and upper troposphere were warming, the lower troposphere experienced slight cooling. This cooling extended into the middle troposphere just ahead of the gust front. The 2350 CST sounding (not included in Fig. 4) recorded −35.3°C and −58.2°C at 300 and 200 mb, respectively. Thus upper tropospheric temperatures returned to values close to the pre-line regime. An area of rapid warming is shown in the mid-troposphere a few minutes after the passage of the gust front. This warm-

![Fig. 4. Time-height cross section of temperature and relative humidity obtained from serial rawinsonde ascents at North Omaha, Nebr., 21 June 1957.](image-url)
ing is thought to be associated with the updraft core and is positioned above a region of significant cooling near the surface. Simultaneously, the upper troposphere is undergoing significant cooling. A third region of warming near 700 mb is indicated behind the main updraft region.

Evidence of sinking in the middle and upper troposphere prior to squall-line passage is also found in the humidity analysis. With time, relative humidities of 30% or less move downward from 350 to below 450 mb, reflecting significant drying in this region. In contrast, the region from 350 to 250 mb shows a marked increase in moisture after 1900 CST. The continued warming as humidities increased suggests the arrival of anvil cirrus along with significant sinking.

Of several possible warming mechanisms (i.e., latent heat release, warm air advection, adiabatic warming by subsidence, or absorption of long or short wave radiation), evidence suggests that subsidence dominates. For example, Schlesinger (1973a, b) presented results of a two-dimensional model simulation of deep convection in both moderate and strong shear. Subsidence was indicated 40-80 km downwind of the major updraft regions in several different simulations. However, it is likely that some of this downward motion was due to the lateral boundary conditions employed. Fankhauser (1974) utilized the NSSL mesoscale rawinsonde network to compute vertical motions along a cross section through the severe-weather-producing squall line of 8 June 1966. His results showed sinking motion in excess of 10 cm s⁻¹ in the mid-troposphere some 20-40 km downwind of the updraft region (see Fig. 5). Williams (1957) used the adiabatic method to compute vertical motions for the case shown in Fig. 4 and found sinking motions of 50 cm s⁻¹ and greater in the upper troposphere 1 to 2 h ahead of the squall line passage. In addition, during high-altitude NASA aircraft flights, P. Kuhn (personal communication, NOAA, ERL, Boulder, Colo.) has observed sinking motions near the tropopause (altitude ≈13 km) some 30-50 km downwind of large cumulonimbi in the Texas Panhandle.

The 3.5 mb h⁻¹ surface-pressure falls shown in Fig. 1b indicate roughly 0.4°C h⁻¹ increase in the mean temperature of the column. If the warming is assumed to occur in the 100-500 mb layer, the layer mean temperature increases about 1°C h⁻¹. The sinking required to warm this amount was computed using the 1200 GMT, 3 April 1974, Nashville, Tenn., rawinsonde data as representative of the upper-level thermodynamic structure over Indiana that afternoon. The vertical motions ranged from −25 cm s⁻¹ at 500 mb and −10 cm s⁻¹ at 300 mb to −5 cm s⁻¹ at 150 mb. These values are certainly reasonable and similar to those given by Fankhauser.

In summary, there is considerable evidence of upper-level subsidence and associated warming downwind of many organized convective systems. The warming occurs above regions where surface mesolows or pressure troughs often exist. The magnitude of sinking in the upper troposphere and lower stratosphere required to reduce surface pressures 2-4 mb h⁻¹ is only tens of centimeters per second. Next, an attempt is made to
isolate some of the factors that may contribute to
donwind subsidence, which in turn determines the
strength and location of lower pressures. The potential
for downwind sinking to contribute to the organization
and maintenance of convective systems is also discussed.

3. Downwind subsidence and mesoscale
organization

Vertical circulations in and near a large cumulonimbus
cloud, embedded in moderate to strong vertical
wind shear, are shown schematically in Fig. 6. The
solid arrows represent a composite of several observa-
tional studies; the dashed arrows are based to a great
extent on indirect calculations. Subsidence in the stable
regions of the lower stratosphere would obviously be a
very effective warming mechanism. Unfortunately, we
have little data on the thermal response of the strato-
sphere ahead of strong convective clouds.

Vertical circulations and some possible variations are
illustrated further in Figs. 7a and 7b. Upper tropo-
spheric vertical motions are shown again in schematic
form. Fig. 7a represents a situation where calm or
easterly low-level winds are overlain by increasing
westerly flow aloft. With all the environmental flow in

Fig. 6. Schematic model of the vertical circulations in and near large cumulonimbus clouds in moderate to strong shear. Solid arrows are based on observations; dashed arrows are based mostly on calculations. Shading depicts region of downwind subsidence. (Adapted from Fritsch, 1975.)

Fig. 7a. Vertical motions in the upper troposphere in and near a cumulonimbus cloud. Magnitudes are indicated by the size of plus and minus signs; several meters per second for the largest to several centimeters per second for the smallest. Cloud movement is shown by dashed arrow. East winds in boundary layer, west winds aloft.
the east–west direction, the cirrus anvil and subsidence should exist in some symmetric pattern to the east of the main updraft. Significant sinking is shown in the wake of the updraft (Fulks, 1962), and over a broader region well downwind of the updraft. Upper-level air would move eastward relative to the cloud (assuming steady state) and warm air would advect downstream from its source in the sinking regions (assumed here to remain stationary relative to the cloud). Fig. 7b presents a convective cloud imbedded in a flow more characteristic of the severe storm environment. Southerly boundary layer winds veer and increase in magnitude becoming west–southwest or westerly in the upper troposphere. For these conditions we assume the clouds move with the mid-tropospheric flow, while the warm air generated in the upper troposphere-lower stratosphere moves with the upper tropospheric winds. Once the warmer air is created by the downwind subsidence field it will move faster and to the right relative to existing cells. The mesolow or trough created in the lower level then increases the boundary layer convergence ahead and to the right of existing cells, providing a mechanism for new cell development on the southeast flank of old cells. This process may have contributed to the rightward movement of the system shown in Figs. 2 and 3.

Figs. 8a and 8b extend the concepts portrayed in Figs. 6 and 7 to the case of squall-line development. Fig. 8a illustrates how subsidence downwind of several cells combines to produce upper-level warming over a region 50–100 km in horizontal dimensions. Fig. 8b shows a three-dimensional view for the same cloud configuration. Downwind subsidence is indicated in both the anvil regions and in the surrounding cloud-free air. Two processes may be acting with the vertical shear of the environmental wind to produce this sinking. First, the compensating subsidence known to occur around convective clouds may be organized downwind of the updraft by the environmental flow. Second, upper-level winds move eastward relative to the penetrating updrafts, with the result that the cloud appears somewhat as a moving barrier to the upper-level flow. While observations show that this flow tends to deflect around the cloud (Fankhauser, 1971), it is possible that some environmental air may be forced over the clouds near the tropopause or in the lower stratosphere. The "lee-wave" effect could be a significant warming mechanism in the stable environment of the lower stratosphere.

The magnitude and position of the upper-level warming may depend on several factors. The vertical shear is apparently the dominant factor in organizing the subsidence on the downwind side, while the size, duration, and juxtaposition of updrafts may modulate the amount of subsidence. In Fig. 8b the most sinking and greatest surface pressure falls are shown ahead of the two strongest cells. Boundary layer winds respond to the developing mesolow leading to an increase in convergence ahead and to the north of the low center. This provides an enhanced supply of low-level moisture for the existing cells in the middle of the line. Although a closed cyclonic circulation often develops in the low-level flow relative to the moving mesoscale system, observations indicate that these closed circulations exist in the actual wind field only for the large and intense meso-β scale systems. Interaction of the bubble-high and mesolow accelerates the southern portion of the gust front southeastward producing a wave on this boundary. Forced lifting by the gust front triggers new cloud development along the southern boundary of the line. The events are self-propagating with the new cells contributing to lower pressure ahead of the line as they grow and mature.

It follows, then, that mesolows or troughs ahead of deep convection can exhibit many variations in in-
tensity and size. Factors that cause and modulate the sinking downwind of large convective clouds are not satisfactorily understood, and obviously need much additional study. Furthermore, lower tropospheric temperature cooling may in some cases compensate for upper level warming, and meso-cyclogenesis will not occur.

Evidence presented in the previous section shows the surface isallobaric field and the middle and upper tropospheric subsidence to be meso-β scale. These meso-β adjustments to processes initially on the meso-γ scale result in "mesodynamic instability." The lower surface pressures increase boundary layer convergence of mass and moisture into the region ahead of the convection, creating the potential for even greater convective activity. Equally important is the temporary suppression of convection in the subsidence region. This occurs through a slight reduction of available buoyant energy and/or generation or enhancement of an inversion separating the sinking air from the converging boundary layer air mass. As shown in Fig. 6, downwind subsidence, coupled with the convection upstream, contributes first to a capping, then to a sudden forced release of the convective instability.

Circulations depicted in Figs. 6–8 inherently contribute to organizing and maintaining strong convection. If the triggering mechanism is quasilinear, such as a cold front or dry line, the mechanism acts to maintain the squall line. If a cumulonimbus system interacts with its environment to concentrate the downwind sinking, a mesolow may form. These mesolows are frequently associated with long and severe tornadoes or with tornado families such as those shown in Fig. 1.

To date, most observational efforts concerned with severe mid-latitude convection have focused on clouds and their immediate environment, especially in the lower atmosphere. These results suggest that more attention should be given to securing observations in the upper troposphere-lower stratosphere region, especially 25–100 km downwind of existing clouds.

4. Summary

Evidence presented suggests that subsidence warming in the upper troposphere and lower stratosphere, downwind of cumulonimbus clouds, is the mechanism responsible for formation of meso-β scale surface lows or pressure troughs. Surface pressure falls of 2–4 mb h⁻¹ can be produced hydrostatically by sinking of the order of tens of centimeters per second in the 100–500 mb layer.

Interaction of convective clouds (meso-γ scale) with a sheared environmental flow (meso-α or synoptic scale) appears to organize the subsidence field downwind of existing clouds leading to the formation of intermediate (meso-β) circulations. Response of boundary layer winds to the meso-β pressure perturbations increases the convergence of mass and moisture ahead of the existing cells, maintaining or enhancing the amount of convection. Also, the subsidence may suppress convective activity until the forced lifting along the gust front releases the convective instability, thereby acting as an organizing mechanism. These processes constitute a mesodynamic instability between the meso-γ scales (the cumulonimbus clouds) and the meso-β scale.

Acknowledgments. The authors wish to thank Mr. James Fankhauser of the National Center for Atmospheric Research for providing the data for the 21 May 1973 NHRE storm presented in Figs. 2 and 3. The rawinsonde data for the North Omaha special series were furnished by the National Climatic Center, Asheville, N. C., Mr. Horace Hudson provided copies
REFERENCES


