

ATMOSPHERIC CIRCULATION SYSTEMS

Their Structure and Physical Interpretation

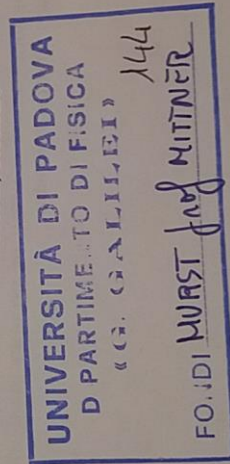
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PREFACE

False facts are highly injurious to the progress of science, for they often endure long; but false views, if supported by some evidence, do little harm, for every one takes a salutary pleasure in proving their falseness.

Charles Darwin
The Origin of Man, Chap. 6.

The objective of this book is to describe the essential nature of the principal circulation systems in the atmosphere, as they appear in light of our present knowledge. We have limited our discussion to the "lower atmosphere," roughly below the 100-mb surface. R. A. Craig, in his volume "The Upper Atmosphere," in this same series, has given a thorough treatment of the layers higher up. However, we also adopted this limitation because we wanted to discuss the processes most directly responsible for weather and climatological phenomena in the most ordinary sense. It is not yet clear how much these phenomena are influenced by processes in the upper atmosphere; while there are certainly connections, it is established that the main driving mechanisms for the lower atmosphere are in the lower atmosphere itself.

Despite the tremendous improvement of the observational data and their treatment during the past 20 to 30 years, these data are still too inadequate to

permit a complete description of the complicated nature of the diverse circulation patterns appearing in the atmosphere. It is therefore necessary to simplify and generalize by using proper methods of analysis founded partly on the principle of eliminating unessential phenomena. The limited observations available in a given system might be analyzed in different ways, it is the aim of synoptic meteorology to find the particular solution that is physically coherent. In the search for this, different hypotheses will appear, whose validity or falsity will be exposed when more observations are provided or more correct theories devised. In this book, the reader will find many such partial descriptions and interpretations, and we hope that the measure of his dissatisfaction will be his urge to improve upon them.

One of the leading objectives of meteorology is to find out, knowing the inputs of energy into the atmosphere and the properties of the fluid constituting it, what makes it operate in the way it does. Nature has devised the working parts most suited for converting energy between the various forms and for transferring the energy from one place to another as this is needed. The ultimate test of any hypothesis about a circulation system is whether in its final form the hypothesis accounts satisfactorily for these functions.

It is our purpose to describe the character of the circulation systems in light of their performance of these various functions. These systems differ in kind both according to the latitudes at which they are found, because the required functions differ in degree, and at a given latitude, because the circumstances of regional physiography require them to take different forms. Thus, to achieve any degree of completeness, we must undertake to describe a considerable variety of systems, sometimes in a very sketchy way. At the same time, we have attempted to provide enough references to enable the reader to track down a more adequate description of a particular phenomenon in which he may have a special interest.

There are practical limitations imposed upon what can be covered in a single volume. Perhaps the most discouraging aspect is that there are extensive areas which properly fall within the framework of the subject matter, but cannot be treated adequately in a few pages and which therefore must be omitted entirely. Secondly, it is disappointing to have to leave out a great deal of material as to the origins of present concepts. We have tried to strike a balance between including a fair amount of material of this sort and writing an account overloaded with references. Thus we have not given a complete historical summary, but where possible we have tried to indicate the sources of the particular aspects we have discussed. Only in Chapter 5 have we specifically included a brief historical review of the earlier develop-

ment of synoptic aerology and the polar-front theory, in order to stress their special importance for our present knowledge.

We have not included anything about aerological techniques or about weather forecasting, since both subjects are treated in available texts. We have also not attempted to present the material specifically in textbook form—for example, the derivations of equations... may be found elsewhere. However, we believe this volume will be suitable for use as an advanced text by those who have some grounding in basic theory and preferably also some practice in synoptic analysis.

Only a few references are made to the results of numerical model experiments in solving general and specific atmospheric circulation problems. Obviously a new era in the study of the atmosphere and its flow patterns has been inaugurated by this technique and by the use of satellite observations. The work in these areas is so vast in scope that their summarization cannot be attempted here, and is best left for experts in these specialties. With these omissions, it is possible that some readers will consider this book somewhat too qualitative or descriptive and perhaps not up to date in every respect. Nevertheless, the basic principles have not been changed by technological advances; and the traditional methods of synoptic aerology and the recent techniques play complementary roles in investigating the atmosphere.

It must be admitted that where alternative views exist on a given subject, we have not exposed all of them, with the excuse that to do this would so encumber the discussion as to destroy its continuity. We have exercised the authors' prerogative to choose the viewpoints discussed. It will perhaps be noticed that a large proportion of the illustrations is taken from our own publications or those of our close colleagues. This is partly a matter of convenience of access and partly a natural result of the fact that we find it easiest to write about material with which we are already familiar.

Largely for these reasons, our illustrations also show a geographical bias, especially toward synoptic systems over and near North America. Readers in other areas will recognize that these systems may differ in important respects from one location to another. Since one of our main objectives is to stress the physical principles, it is hoped that these will not suffer from the illustration of circulation systems of special types; the same basic principles apply regardless of locality. Nevertheless, we must apologize for not having extended the description to certain broad regions.

Thus, in many respects, this book is not what we had hoped it would be, but what we were able to make it. We trust that we have given an accurate account of the work of others. Such an aspiration is difficult to realize, since the meaning of what one reads is likely to be colored by his own thoughts on

Preface

the subject, so that in mentioning an author's work, one may either confound the original message or put the emphasis where the author does not intend it. At most, we pray that we have not obscured what is fact and what is hypothesis, and thereby violated Darwin's maxim used as a motto for this book.

In the preparation of this volume, we have been materially assisted by many persons as well as several institutions. We would specifically like to thank The University of Chicago, The University of Stockholm, The University of Helsinki, the United States Weather Bureau, and the National Center for Atmospheric Research for making available support and facilities that made it possible to carry out this work. We are personally indebted to Professors Sverre Pettersen, Bert Bolin, Bo Döös, and Lauri A. Vuorela, and Dr. Robert H. Simpson for their kindness in making the necessary arrangements. A number of our colleagues have given comments on parts of the manuscript or offered advice useful in its preparation. Ralph L. Coleman has been especially helpful in analyzing some of the figures and in drafting most of the new illustrations. We are grateful to Miss Margaret Johnstone, Mrs. Nancy Seiller, Mrs. Lois Gries, and Mrs. Eileen Workman, who cheerfully suffered through many drafts, for their expert typing and checking of the manuscript. We are particularly indebted to Professor Carl W. Kreitzberg for reading the manuscript and calling attention to a number of errors we would otherwise have overlooked.

March 1969

E.P.
C.W.N.

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LIST OF SYMBOLS

Special Conventions

C (E_1 , E_2)	conversion of energy from form E_1 to form E_2 , where the E 's may be either A_m , A_s , K_m , or K_s
$\dot{Q}$	dQ/dt
$\bar{Q}$	mean value around a latitude circle
$\bar{Q}$	mean value around an arbitrary boundary
$[Q]$	mean value over an area
Q'	local deviation from mean value
Q''	local deviation from area mean $[Q]$, where deviations (Q') from boundary means are also used
Q_0	value of Q at earth's surface (unless otherwise specified)
Q_g	geostrophic value
Q_x , Q_{xy}	component of Q in x -direction or in xy -plane, etc.

Common Symbols

Only symbols used repeatedly are listed. Others are defined as needed in text, sometimes using the same letters. Boldface letters indicate vector quantities.

a	radius of earth
c	speed of movement of a circulation system, such as a wave
c_p	specific heat of air at constant pressure
c_v	specific heat at constant volume
f	coriolis parameter, $2\Omega \sin \varphi$
g	acceleration of gravity
h	(1) thickness between two isobaric surfaces; (2) sensible heat content per unit mass

k	(1) kinetic energy per unit mass; (2) trajectory's curvature
k_m	kinetic energy of zonally averaged motions, per unit mass
k_z, k_φ	kinetic energies of mean zonal (k_z) and meridional (k_φ) motion, per unit mass
k_s	streamline curvature
n	distance normal to wind direction or to a curve as specified
p	pressure
q	(1) specific humidity; (2) heat content per unit mass
r	(1) radius in cylindrical coordinate system; (2) radius of flow curvature
s	distance downwind along a streamline
t	time
u, v, w	eastward, northward, and upward components of wind velocity
x, y, z	distance eastward or northward, and height above sea level
A	(1) area; (2) wave amplitude
A_e, A_m	integrated eddy or mean available potential energy within a specified volume
A_s	streamline amplitude
C	horizontal movement of an isopleth
D	(1) divergence (horizontal, unless otherwise specified by subscripts); (2) rate of frictional dissipation of kinetic energy within a specified volume
E	(1) mass of water evaporated from earth's surface, per unit time and area; (2) static energy ($c_p T + gz + Lq$)
F_{xy}, F_{xz}	convergence, in the poleward or vertical direction, of the eddy flux of eastward linear momentum
I	total energy within a specified volume
J	eddy flux of absolute angular momentum, in direction specified by subscript
K	total kinetic energy within a specified volume
K_e, K_m	integrated kinetic energy of the eddy or mean motions within a specified volume
K_φ	transfer of kinetic energy across latitude φ
K_y	mean kinetic energy of the meridional motions, integrated within a volume bounded by latitude circles and pressure surfaces
L	(1) wavelength (L_s , stationary wavelength); (2) length of a boundary; (3) latent heat of vaporization
M	total absolute angular momentum in ring bounded by two latitudes and pressure surfaces
$(M)_p$	downward mass flux through an isobaric surface, in specified area of cold air
M_r	absolute angular momentum about the central axis of a cyclone, per unit mass, at distance r from center
$(M_r)_r$	radial influx of absolute angular momentum through a cylinder at distance r from the center of a revolving system
$(M_r)_z$	upward flux of absolute angular momentum through an isobaric surface between two radii of a circulation system
M_φ	absolute angular momentum transferred poleward across latitude φ between two pressure surfaces

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M_1	total vertical flux of absolute angular momentum through an isobaric surface, over a specified area
P	(1) mass of precipitation reaching earth's surface, per unit time and area; (2) total potential energy within a specified volume
Q_a, Q_s	heat content of atmosphere or earth
R	gas constant for air
R_a, R_e, R_{ea}	latitudinal mean net (excess of absorbed over emitted) radiation per unit time and area received by atmosphere, earth, or earth-atmosphere system
Q_s	transfer of sensible heat from earth to atmosphere, per unit time and area
T	temperature (in some contexts, virtual temperature)
ΔT	temperature excess of an air parcel in a convective cloud, relative to the undisturbed environment temperature
V	total wind speed
V_a	scalar value of agostrophic wind velocity $V_a = (V - V_g)$
W	energy per unit area (excluding latent heat)
W_p	mean poleward flux of energy (other than latent heat) per unit length of a latitude parallel, in atmosphere or earth as designated by secondary subscript a or e ; W_p' includes latent heat flux
α	specific volume
β	df/dy , the variation of coriolis parameter poleward
γ	lapse rate ($-dT/dz$) of temperature with height
γ_a	adiabatic lapse rate ($-dT/dz$), dry or moist, depending on saturation
ζ	relative vorticity about a vertical axis
ζ_a	absolute vorticity ($f + \zeta$)
ζ_0	relative vorticity in an isentropic surface
η	absolute vorticity ($f + \partial\theta/\partial r$) in symmetrical cyclone
θ	(1) potential temperature; (2) angular measure in cylindrical coordinate system, increasing in cyclonic sense
θ_a, θ_w	equivalent and wet-bulb potential temperature
λ	longitude
μ	coefficient of eddy viscosity
ρ	density
σ	(1) vertical shear $\partial u/\partial z$; (2) horizontal wind speed inside a convective cloud; (3) area of a surface bounding a specified volume
γ	horizontal stress
φ	latitude
ψ	(1) Montgomery stream function ($c_p T + gz$); (2) angle between a frontal surface and the horizontal; (3) slope of a streamline in the vertical plane
ω	vertical wind component (dp/dt) in pressure coordinate system
Γ	lapse rate (dT/dp) in pressure coordinate system
Γ_a	adiabatic lapse rate (dT/dp), dry or moist process, depending on saturation
Φ	geopotential (gz)
Ω	angular velocity of earth's rotation
$\mathbf{F}$	(1) frontogenetic function, with components F_x, F_y, F_z ; (2) frictional force per unit mass
τ	tensor of the Navier-Stokes viscosity stresses

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ORGANIZED CONVECTIVE SYSTEMS IN MIDDLE LATITUDES

Over many regions of the middle latitudes as well as the tropics, a large portion of the total annual precipitation is derived from convective showers and thunderstorms, which also account for the greatest overall weather damage in many areas. Convective storms are also of particular interest because they manifest perhaps the most obvious interaction between circulations on large and small (or medium) scales. The occurrence of significant convection is strongly influenced by synoptic-scale disturbances, and convective clouds in turn are very effective agents for the vertical transfer of energy and horizontal momentum in amounts significant for the general circulation.

In this chapter we treat mainly a characteristic type of organized convective system connected with synoptic disturbances. The type of "air mass" convection initiated by surface heating will not be discussed, since the principles involved are described in many textbooks. It should, however, be stressed that the degree of "air mass" convection is also influenced by the larger-scale synoptic systems. This is well known to all forecasters, in terms of the flow patterns that determine the air masses over a given region, and in terms of the influence of general ascent or subsidence upon the stability and the depth of the moist layer. It also follows from the principles of the atmospheric water budget (Section 12.6). Since the total mean precipitation over a large region of shower activity is determined by the evaporation from the earth's surface and the net influx of water vapor, in the

13.1 Distribution of Severe Storms

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absence of such an influx the mean precipitation will correspond to the mean evaporation. Thus "air mass" showers can occur in a quiescent synoptic situation, but they may be significantly enhanced or suppressed by relatively feeble organized convergence or divergence in the lower layers.

13.1 DISTRIBUTION OF CONVECTIVE STORMS AND OF SEVERE STORMS

The frequency of thunderstorm occurrence varies greatly, owing to (among other influences) the character of the earth's surface. Over the eastern oceans in middle latitudes, 5 to 10 thunderstorm days per year are characteristic, while in some parts of tropical Africa and South America, thunderstorms occur on 180 to 200 days per year (World Meteorological Organization, 1956).

There is, however, no close relationship between the frequency of occurrence of thunderstorms and the damage they produce. In Fig. 13.1, Ludlam

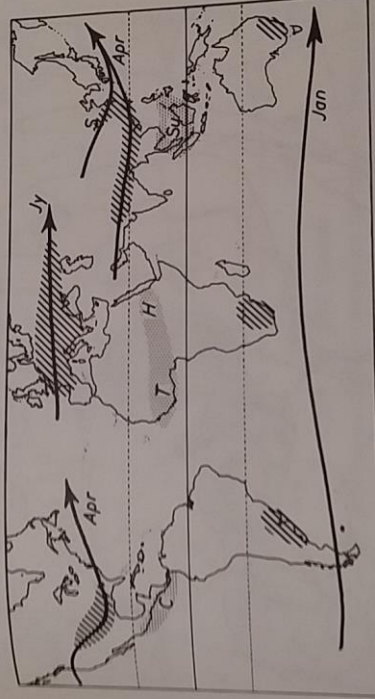


FIG. 13.1 According to Ludlam (1963): "Distribution of areas subject to severe squall-thunderstorms. In middle latitudes these areas (hatched) are closely related to the mean position of the jet-stream axis at the 500-mb level for the appropriate season. . . . In midsummer the jet over N. India shifts (to north of the Tibetan plateau), and the severe hailstorms [and squalls] there become infrequent. In the tropics the squall-storms are not accompanied by large hail. They mostly occur over regions (stippled) where monsoonal circulations produce well-defined intertropical fronts near which there is a lower-tropospheric wind shear not closely related to a middle-level jet stream." The regions T-H and Su are, however, overlain by strong mean easterlies at the 200-mb level (see Fig. 3.13b) in summer. Also, although the strongest winds in the Southern Hemisphere are at the latitude shown, a secondary maximum of mean westerlies passes directly over the squall areas.

(1963) has summarized the regions where severe squall thunderstorms are most common. In middle latitudes, these are in general the localities that most commonly come under the influence of the jet stream. Since cyclones are also associated with the jet-stream region, this is one indication of the frequent connection between synoptic-scale disturbances and severe thunderstorms. This association results from two influences: the action of synoptic-scale circulations in creating an air-mass structure favorable for convective overturning; and invigoration of the energy processes of the storm itself, due to the structure of the wind system in which it is embedded.

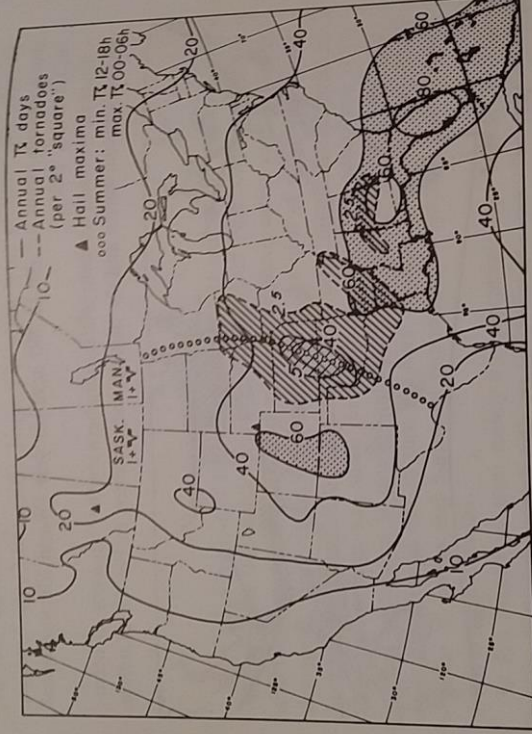
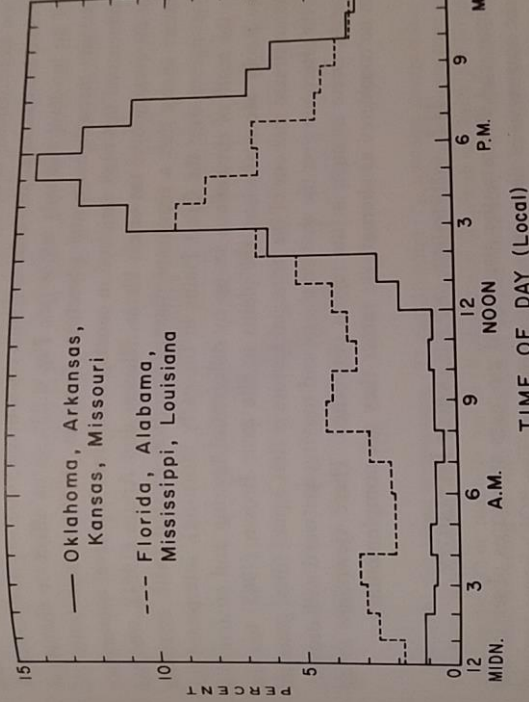


FIG. 13.2 Annual number of thunderstorm days (solid lines, taken with some alterations from World Meteorological Organization, 1956) and annual number of tornadoes per 2° latitude-longitude tessera (dashed lines, from House, 1963). Along the line of circles there is a nocturnal maximum and afternoon minimum in thunderstorm frequency.

The lack of any clear correlation between the frequency of thunderstorms and their severity is emphasized by Fig. 13.2. Although tornadoes (which occur with thunderstorms that are in other respects severe) occur in the broad region of frequent thunderstorms, the maximum number of severe storms coincides nearly with a relative minimum (< 40) of the total thun-

derstorm days per year. Most thunderstorms do not reach destructive proportions, for which special environmental conditions seem to be required. The western maximum in Fig. 13.2 is due to the frequent formation of heat thunderstorms over mountains; the southeastern one, to the direct influence of surface heating, the effect of which is greatly enhanced by the convergence of sea breezes during daytime over the Florida peninsula (Byers and Rodebush, 1949). In the Great Plains region, where there are fewer thunderstorms, the air is not normally so rich in water vapor as in the area farther east. However, this region is peculiarly suited, by virtue



surface heating is not so pronounced, the influence of insolation and outgoing radiation also frequently being suppressed by extensive cloud cover in the marine-air mass. Dynamical influences due to synoptic disturbances assume a relatively large importance, and these may initiate thunderstorms at any time of day or night.

In the central states, Fig. 13.3 also shows that the maximum occurrence is not in early afternoon when the surface temperature is usually highest, but is displaced toward late afternoon and evening. As indicated in Fig. 13.2, in parts of this region in the warmer part of the year, thunderstorms are more frequent in the early morning hours than in the afternoon. Meigs (1944) has connected this behavior with the diurnal variation in the strength of the low-level southerly winds (see Fig. 8.25), in its effect on changes of stability through horizontal advection. Sangster (1958) demonstrated the influence of this wind variation in causing maximum low-level convergence in the evening over the Great Plains. Bleeker and Andre (1951), who had earlier shown that a significant diurnal variation existed (with maximum convergence over the Great Plains in the early morning), interpreted this as the result of circulations set up by differential heating and cooling over the Rocky Mountains and the plains to the east. Brown (1962) has also demonstrated that one of the principal factors is that squall lines, generated in the afternoon over the western plains and moving eastward with ordinary speeds, arrive at night in the region concerned. These viewpoints should not be considered as alternatives; rather they are complementary.

13.2 GENERATION AND RELEASE OF INSTABILITY

Figure 13.4 illustrates the diversity of air-mass structures observed prior to the occurrence of destructive thunderstorms. The "Oklahoma tornado" sounding in Fig. 13.4a is also representative of situations that produce medium-sized hail; on occasion, appreciably greater instability may be present, producing very large hailstones. Studies in different regions by Donaldson (1958), Douglas and Hirschfeld (1959), and Schleusener and Grant (1961) indicate that the higher the top of a thunderstorm, especially when it penetrates the tropopause, the greater is the probability of hail from it. This association is due to the fact that strong upward currents are required both to carry the air in the updraft to levels where it becomes colder than the surroundings (and thus negatively buoyant) and to suspend or recirculate hailstones, whose terminal velocities increase with their sizes, long enough for them to grow. Other manifestations of storm severity are in a general way correlated with the conditions favorable for hail. Thus,

13.2 Generation and Release of Instability

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For example, tornadoes may occur with either of the sounding types shown in Fig. 13.4, but are more likely and more severe with the type in Fig. 13.4a. The common feature of these soundings, considering the overall depth of the troposphere, is the presence of both convective and conditional

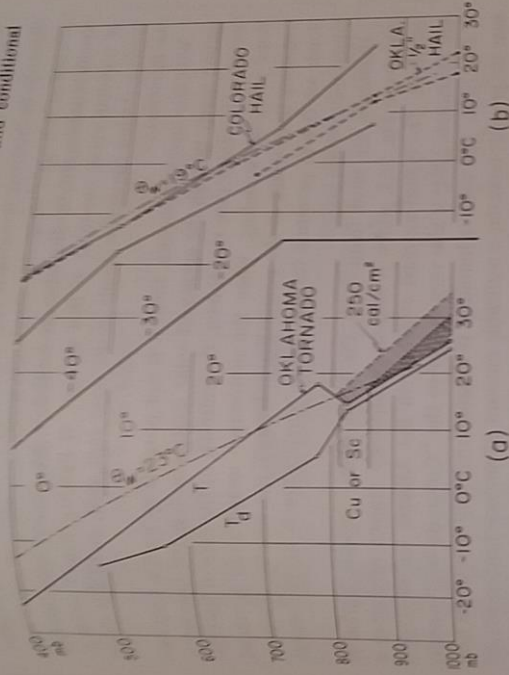


FIG. 13.4 (a) Mean temperature (T) and dewpoint (T_d) sounding shortly before the occurrence of tornadoes, after Fawbush and Miller (1953). (b) Dashed curves show mean sounding for 1/2-in. diameter hail in Oklahoma area (Fawbush and Miller). Solid curves show mean sounding for hail situations at Denver (Beckwith, 1960). Dash-dotted curves are moist adiabats corresponding to θ_w in lowest 100 mb, not considering the increase that would result from solar heating of this layer (the curve $\theta_w = 19^\circ\text{C}$ is valid for both soundings on the right).

instability. The degree of instability, however, may vary greatly, and the conditions required to release the instability may also differ.

This may be appreciated by comparison of the two Oklahoma soundings. The tornado sounding is characterized by great potential instability,¹ but the upward penetration of air from the moist layer is inhibited by the

¹ In our usage, "potential instability" will refer to a combined condition of "convective instability" ($\partial\theta_w/\partial z < 0$) and "conditional instability" ($\gamma > \gamma_w$) when a deep layer is considered. In general, either of these latter types of instability may exist without the other. The terms *potential* and *convective instability* are sometimes used synonymously.

inversion near 800 mb. In spring, for which season this sounding is most representative, the potential insolation (heat accumulation from incoming minus outgoing radiation, under clear skies, between the coolest and warmest part of the day) is roughly 250 cal/cm² at the latitude concerned. If a layer Δp in thickness is warmed by an amount ΔT , the heating required per unit area is $c_p \Delta T \Delta p/g$. The maximum possible heating (hatched area) would be inadequate to eliminate the inversion. Moreover, with this type of sounding about one-third of this amount of heating would suffice to produce shallow convective clouds in the layer indicated, and these reduce the insolation.

With this thermal structure, then, it is not possible to realize deep convection purely through the agency of solar heating unless the inversion is considerably weaker than that shown. By contrast, in the case of the Oklahoma small-hail sounding, when the convective temperature is reached (requiring three-tenths of the potential insolation), the ensuing convective clouds are free to penetrate to high levels, since there is no inversion to inhibit their vertical growth.

Hailstorms are most frequent in the High Plains regions marked by triangles in Fig. 13.2, which is seldom invaded by marine tropical air in low levels. The typical air-mass structure in the Alberta region (Longley and Thompson, 1965) is also very similar to the "Colorado hail" sounding in Fig. 13.4b. The relative humidity is characteristically low in the layers near the ground; however, a realizable amount of surface heating will start convection. At Denver (Beckwith, 1960), hailstorms commonly occur some time after a cold-front passage; northeasterly winds increase the moisture content in low levels, and orographic lifting is also influential.

Longley and Thompson found it difficult to differentiate clearly between the stabilities and lower-level moisture contents on major-hail days and days without hail in Alberta. The synoptic parameters indicated, however, that afternoon hail is most likely in Alberta when there is an upper-tropospheric trough over British Columbia in the morning, with an approaching cold front and a low-level warm tongue ahead of it. These are conditions under which general ascending motions are to be expected; it may be noted that a lifting of the High Plains sounding by even as little as 50 mb makes this air mass much more susceptible to thunderstorm formation from surface heating.

The foregoing considerations indicate that it is necessary to consider the dynamical factors associated with synoptic systems as well as the vertical structure of the air mass observed at a given time. This is reflected in the procedures for forecasting severe local storms, described by Showalter and

fuls (1943), Fawbush *et al.* (1951), and the staff of the Severe Local Storm Forecast Center (U.S. Weather Bureau, 1956). The importance of synoptic-scale processes for generation of severe convective storms is also recognized in tropical regions (Section 14.10).

The conditions that contribute to generate convective instability and that lead to its release were in general terms outlined rather completely by Norman (1938). Figure 13.5 illustrates these conditions schematically for

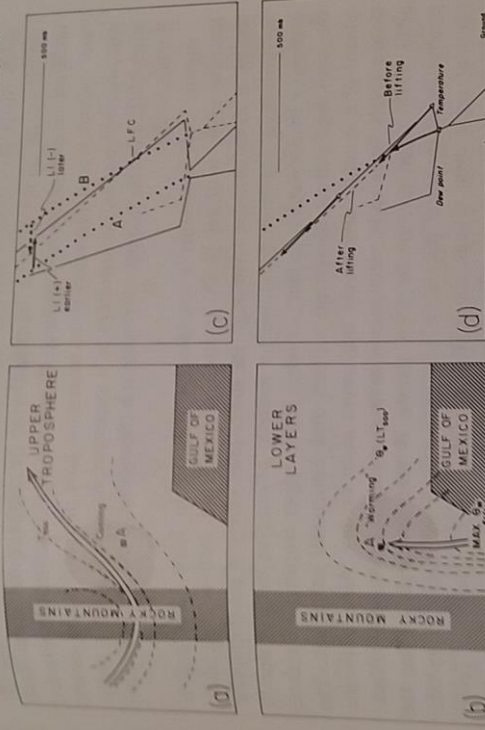


FIG. 13.5 Schematic illustration of processes in formation and modification of potentially unstable air mass. In (a) and (b), arrows are axes of high-tropospheric and low-level jets. In (a), 500-mb isotherms are indicated; in (b), wet-bulb potential temperature in moist layer, which when lifted adiabatically reaches temperature LT_{500} at the 500-mb level. In (c) and (d), temperature and dewpoint curves; lines of circles are moist adiabats corresponding to θ_w of moist layer. In (c), the lifted index (LI) indicates potential stability at first, changing to potential instability later. In (d), sounding curves before (solid) and after (dashed) lifting. (From Newton, 1963.)

a common situation east of the Rocky Mountains. For examples of real situations, see Sugg and Foster (1954) and U.S. Weather Bureau (1956). One of the essential processes is differential horizontal advection in the lower and upper troposphere, the importance of which was emphasized by Means (1944) and which has been discussed further by Miller (1955). It should be noted that horizontal advective processes do not release the

instability, although the upward motions leading to such release tend to be associated with the regions where differential advection contributes to growth of potential instability.

On the approach of a cold trough aloft from the west (Fig. 13.5a), there is usually an increase in the southerly current over the Great Plains region, often accompanying cyclogenesis east of the Rocky Mountains. Air with high-moisture content is thus drawn northward from the Gulf of Mexico, as shown in Fig. 13.5b, where the wet-bulb potential temperature (θ_w) of the air in the moist layer is identified with its temperature when lifted unmixed to the middle troposphere. Comparison of this "lifted temperature" (which would characterize the innermost part of a convective cloud) with the environment temperature at a corresponding level gives a useful index of the possible intensity of convection (Showalter, 1953; U.S. Weather Bureau, 1956).

Figure 13.5c represents a sounding at point A in Figs. 13.5a and b. Increase of θ_w in lower levels, in itself or combined with cooling aloft, results in destabilization. Although the lower-level changes are dominated by horizontal advection (modified by diabatic heating or cooling due to proximity to the earth's surface), cooling in the upper troposphere reflects the combined effects of vertical motion and horizontal advection. The advent of a synoptic disturbance may occur night or day. Also, advection in low levels tends to be strongest in early morning (Means, 1944; Blackadar, 1957) when the low-level jet stream is best developed. Thus, the generation of potential instability, and on occasion its release, may take place at night even though the additional influence of insolation favors daytime onset of convection. The importance of synoptic circulations varies with the situation. Occasionally, extreme instability may be created in a previously stable region, mainly by differential advection, in 12 hr or so. On other occasions, particularly in the warmer months, an area may be occupied for extended periods by marine tropical air, in which case differential advection is unimportant and insolation dominates the onset of convection.

As discussed in connection with Fig. 13.4a, when a stable layer is present it may not be possible to realize deep convection through surface heating alone. Beebe and Bates (1955) have demonstrated that such a stable layer may be eliminated by organized vertical motions, in the manner shown by Fig. 13.5d. If, as is usual, the inversion layer is convectively unstable, lifting results in the air beneath it becoming saturated before the air above (Rossby, 1932) and thus cooling at a different rate. The lifting required to vitiate a strong inversion may be of the order of 100 mb. Hence, ordinary cyclone-

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scale vertical motions (of, say, 5 mb/sec) can eliminate the inversion in 6 hr.²

As pointed out by Fjølke (1951), the inversion plays an important role in the generation of potential instability. If a strong inversion is present, initially it prevents deep convective overturning. Thus the potential instability may increase to a high degree in the manner of Fig. 13.5c if the differential advection and related processes are pronounced and go on long enough. Finally, when a mechanism is imposed to eliminate the inversion, the stratocumulus or slanted cumulus clouds, which may earlier have been formed in the marine layer, are free to penetrate upward and form cumulonimbus.

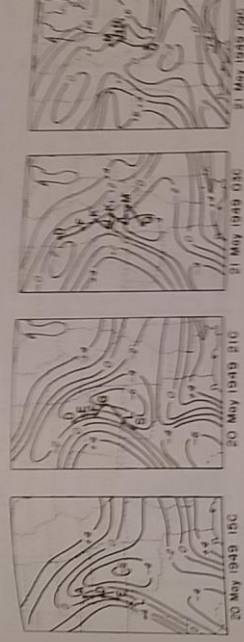


FIG. 13.6 Showalter stability index (500-mb temperature minus temperature of parcel lifted to 500-mb level from 850 mb; negative values indicate potential instability). Dots on lines, identified by letters, show successive locations of heaviest precipitation in individual rainstorms associated with squall line. Stability isopleths in first and third charts were interpolated; there were no upper-air observations at those times. Potential instability was present in some regions outside the zero isopleth, where conditions at 850 mb did not represent the maximum θ_{e0} of the low-level moist air. (From Newton and Newton, 1959.)

The mechanism initiating convection may be, as has been suggested, simply the organized slow vertical motions connected with synoptic disturbances, combined with insolation. In some cases the first convection is clearly associated with a cold front.³ Figure 13.6 illustrates the behavior of a squall line initiated by a cold front that, moving more rapidly eastward

² Horizontal advection is often an additional factor in removing the inversion locally. Advection may first contribute to warming in the lower middle troposphere and then to cooling after passage of a warm tongue at 850 to 700 mb (U.S. Weather Bureau, 1956.)

³ Analyses of the circulations in the vicinity of cold fronts, by Clarke (1961) and Brundage (1965), indicate upward motions of 50 to 70 cm/sec at 500- to 1000-m height in a relatively narrow band preceding them.

than the tongue of potentially unstable air, swept into its western edge and set off a line of thunderstorms there. These progressed eastward through the tongue of instability, first increasing in intensity and then mostly dying out as they encountered the stable air further east. This is a fairly common pattern of behavior. Note that since the tongue of instability may itself move eastward with appreciable speed, the extent of the swath swept by a squall line of this sort is likely to be considerably greater than the breadth of the unstable tongue at a given time.

The western boundary of the tropical marine layer is often very sharp, as in the example analyzed by L. D. Sanders in Fig. 13.7 (U.S. Weather

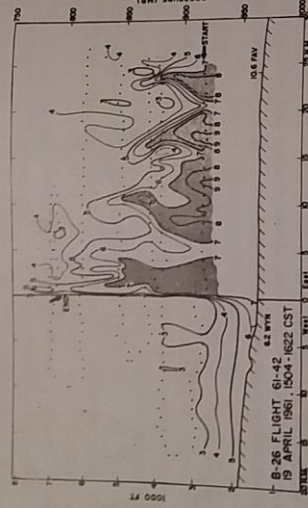


Fig. 13.7 Isopleths of mixing ratio (g/kg) in vicinity of boundary between continental and maritime tropical-air masses over northwest Oklahoma, Apr. 19, 1961. Constructed from stepwise aircraft traverses shown by dotted lines. Analysis by L. D. Sanders, from U.S. Weather Bureau (1963).

Bureau, 1963). In this case the moisture transition took place within about 1 km in the horizontal; the varying depth of the moist air to the east was probably associated with gravity waves set up at the western boundary. That boundary, which is normally characterized by a shift from southerly winds in the marine air to southwesterly winds in the dry air, is in the warmer months a semipermanent zone of convergence. This has long been known as a favorite place for thunderstorms to form, even without the intervention of cold fronts, as brought out by the statistics of Rhea (1966).

It will be noted in Fig. 13.6 that the potential instability was also quite pronounced in the southern region where no thunderstorms developed. This is because the stronger ascending motions were confined to the northern

portion, where there was a well-developed, upper-level wave and jet stream (Section 6.2). The inversion capping the moist layer was more or less undisturbed farther south so that penetrative convection could not take place there.

Faulk (1951), Fawbush *et al.* (1951), and others have emphasized the role of orography in the climatological distribution of severe storms. The reason for their great frequency in the Oklahoma-Kansas region (Fig. 13.2) is that this region is conveniently close to a source of moist air, in the Gulf of Mexico, and just east of a high plateau. Tropical marine air moving northward from the Gulf is constrained by the Rocky Mountains from spreading westward. The low-level jet stream, advecting air with high θ_w northward (Fig. 13.5b), tends to achieve strongest development a few hundred kilometers east of the Rockies. This, along with the circumstance that cold upper troughs frequently move in from the Pacific Ocean, makes this region particularly susceptible to the generation and release of strong potential instability.

13.3 SYNOPTIC STRUCTURE OF SQUALL LINES

Although severe weather is on many occasions associated with isolated large thunderstorms, it usually reaches most destructive proportions with squall lines that sweep over large areas during their lifetimes. The general structure of an extensive squall line is well illustrated by Fig. 13.8, taken from Bergeron (1954).

An outstanding feature of the thunderstorm region behind the squall line is its general association with an area of high pressure. This is attributed to cooling in the lower layers, owing primarily to the evaporation of rain and to some extent also to melting of snow or hail or both. Surface winds diverge with a strong component across the isobars (Tepper, 1950) because the air does not remain long enough in the region of strong pressure gradient for the Coriolis acceleration to be effective. With the usual pressure rise of about 2 to 5 mb, the approximate Bernoulli relationship $\Delta p/\rho = \Delta V^2/2$ suggests that air parcels traveling from the high to the region of lower pressure should attain a speed of about 18 to 28 m/sec. On the average, this is somewhat diminished as a result of surface friction, but locally stronger winds may occur.

At the leading edge of the thunderstorm system, a sharp wind shift is generally observed so that, as in Fig. 13.8, the squall line has the appearance of a cold front. A rapid drop of temperature takes place near the wind-shift line, along with a decrease in θ_w , despite a general increase of relative

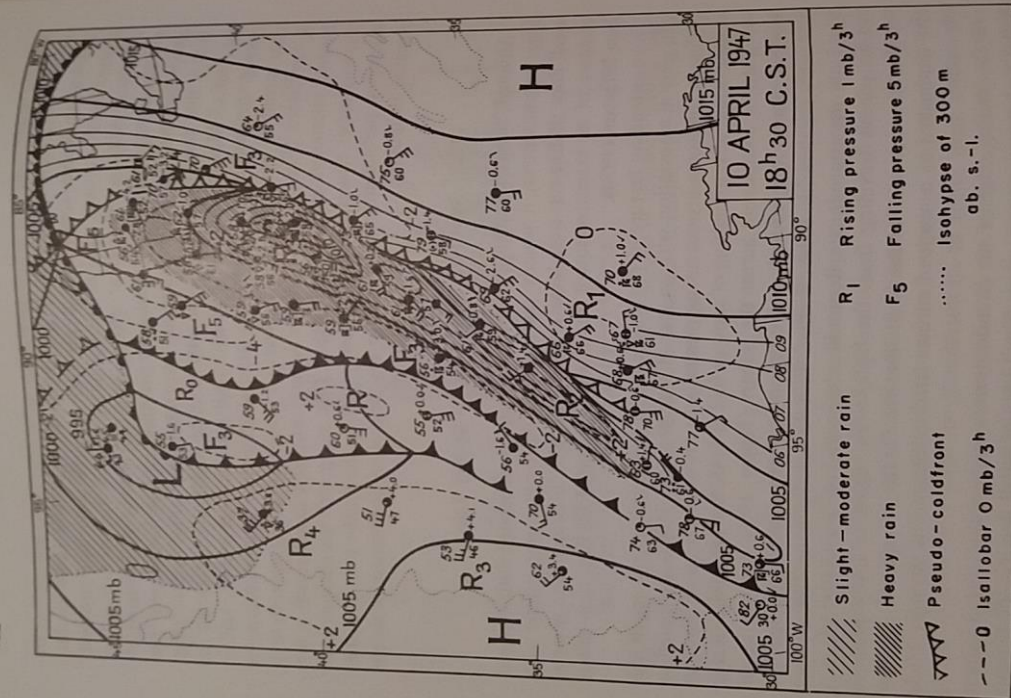


FIG. 13.8 Weather map with convective system over central United States, at 1830 CST Apr. 10, 1947. (After Bergeron, 1954).

13.3 Synoptic Structure of Squall Lines

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humidity. Although squall lines most commonly occur in the warm sectors of cyclones (or in tropical air well removed from frontal systems), they often extend far north of warm fronts and occasionally form some distance behind cold fronts; in both cases the thunderstorms are in the warm air above the front.

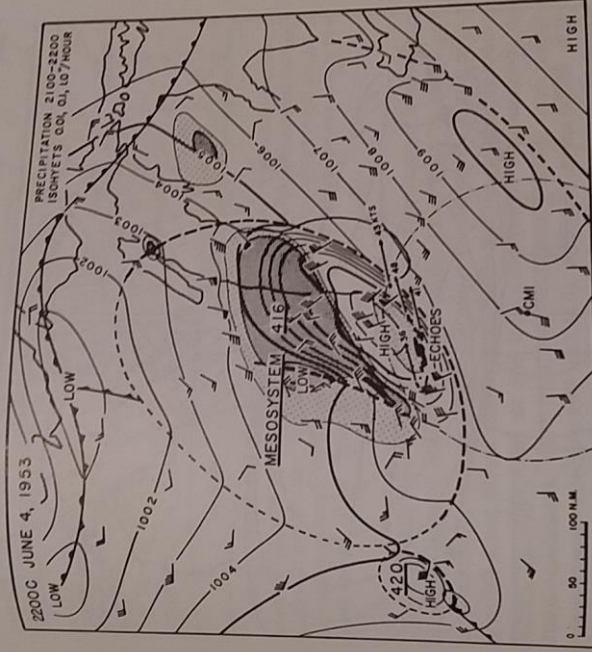


Fig. 13.9 Surface chart for 2200 CST June 4, 1953. Sea-level isobars at 1-mb intervals. In wind symbols, a full barb is 5 knots and a flag is 25 knots. Dashed lines are boundaries of cold-air domes of mesosystems. Isohyets are for precipitation (stippled) during previous hour. (After Fujita and Brown, 1958.)

Some squall lines appear as single outbreaks of thunderstorms that achieve great lengths, as in Fig. 13.8. In other instances, several discrete "mesosystems" (Fig. 13.9) may appear successively. At the stage shown, the cold air produced by the thunderstorms has spread out over a much larger area than is covered by the precipitation itself. Fujita (1959) has demonstrated that the mass of cold air produced is proportional to the cumulative mass

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of rain that has fallen from the convective system and to the amount that has been evaporated into the subcloud layer. Figure 13.10 shows how, as this cold air is continuously augmented and diverges from the region of active thunderstorms, the area covered by the mesosystem expands with time.

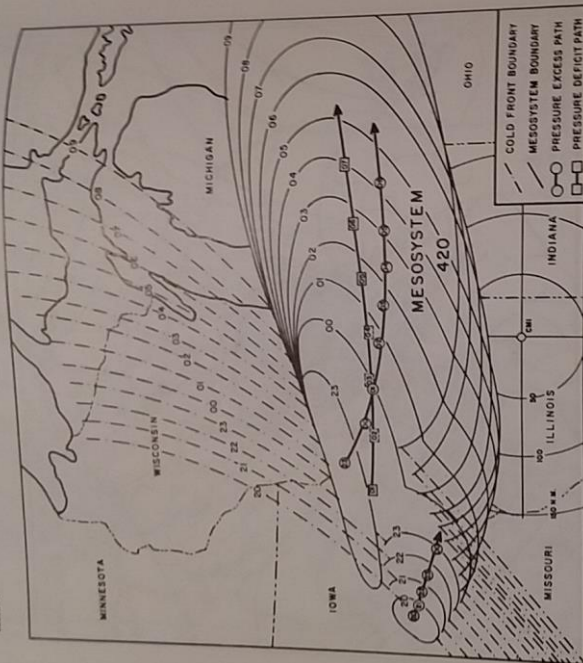


FIG. 13.10 Isochrones (hours in CST) of cold front (dashed) and boundary of the smaller mesosystem (420) in Fig. 13.9, on June 4-5. Arrows indicate tracks of high center (circles) and of a low center to its rear, these features being analogous to those observed in mesosystem 416 of Fig. 13.9. (After Fujita and Brown, 1958.)

In Fig. 13.9 a small-scale low is seen to the west of the precipitating system. This feature, whose cause has not been conclusively explained, is characteristic of the mature stage of a mesosystem (Brunk, 1953; Williams, 1953; Fujita, 1955); it has also been observed in England (Pegley, 1962). Intense pressure dips and strong winds may occur with its passage (Brunk, 1949).

13.4 General Aspects of Storm Circulation

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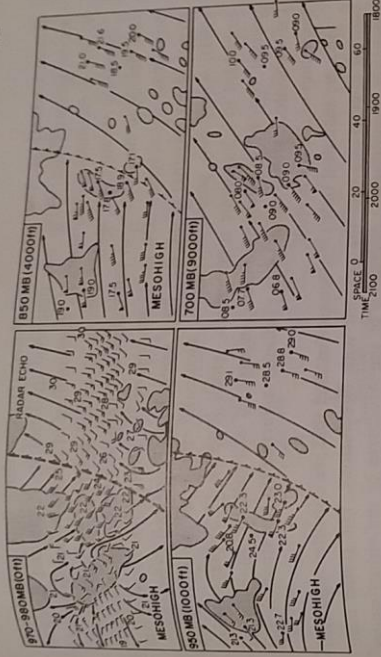


FIG. 13.11 Wind field of a squall-line system passing over the Thunderstorm Project network in Ohio, on June 29, 1947. Dashed line is squall line; shaded areas are radar echoes from precipitation. The numerous observations represent a combined analysis of space and time data from a dense surface network and serial radiosonde ascents from several stations. (After Fujita, 1959.)

The wind shift observed at the ground on passage of a squall line decreases in intensity upward. This is illustrated by Fig. 13.11, where the shift is seen to be relatively weak already at 850 mb and undetectable at 700 mb. Observations show (Byers and Braham, 1949) that the temperature contrast between the rain-cooled air and the undisturbed air is greatest near the ground and becomes quite small near cloud base. Thus the excess pressure beneath the storm (Figs. 13.8 and 13.9), which reflects the weight of the cold air above, rapidly decreases in magnitude upward, as does the disturbance of the wind field associated with it.

13.4 GENERAL ASPECTS OF STORM CIRCULATION

Because of the difficulties and hazards of making direct measurements in intense convective systems, aerological descriptions of them are fragmentary, and the character of their circulations must be largely inferred from indirect evidence.

This evidence suggests that there is a variety of structures as well as a wide range of storm sizes and intensities; the larger and more severe storms have in general more highly organized, more asymmetrical, and more per-

sistent circulations. The "air mass" thunderstorm, which has been extensively explored and which was described in detail by Byers and Brahman (1949), is characterized by sporadic development of relatively small "cells." These pass through an updraft phase, to mixed updraft and downdraft, and finally to downdraft, an individual cell perishing after about a half-hour from its inception. On the other end of the scale, some large severe storms are apparently dominated by a single huge "supercell" (Browning, 1964), which may persist for hours. Generally, these severe storms are embedded in a baroclinic environment, which (Section 13.2) plays an important role in the generation and release of convection; in addition, the associated vertical shear has a profound influence on the structure and mechanics of the storm itself.

Here we shall deal only with certain features that we consider most relevant to interactions with the synoptic-scale circulations. The reader interested in a comprehensive discussion of the structures of individual storms and groups of storms is referred particularly to Browning *et al.* (1965), who provide the most complete and detailed description up to the time of this writing.

The broad relationship between a large, organized, convective system and the associated synoptic disturbance is illustrated by Fig. 13.12, taken from Fankhauser (1965). This shows that the general lower-tropospheric confluence and upper-tropospheric diffluence associated with the cyclone and upper trough were concentrated in the vicinity of the convective system. Computations of the air flow across the dotted boundary in Fig. 13.12 give the vertical distribution of mean divergence over the enclosed area, shown by curve D_p in Fig. 13.13. This indicates marked convergence in the lowest layers, weak convergence through most of the troposphere, and increasing divergence above 400 mb to 200 mb, near the tropopause level that corresponded approximately to the general top of the cloud system. Fankhauser's further analysis shows that weak and indifferent vertical motions were present in the middle troposphere outside the near-vicinity of the squall line. Hence, it may be concluded that *nearly all ascent of air associated with the cyclone was taking place in the convective system*. The computed water-vapor influx, represented by

$$\oint_L \int_{z_{500 \text{ mb}}}^{p_0} \nu_n q \frac{dp}{g} dL$$

where ν_n is the inward component of the wind across the boundary L and q is specific humidity, is represented by curve M for 50-mb layers. The vertically integrated vapor accumulation was within 5% of the mass rate of

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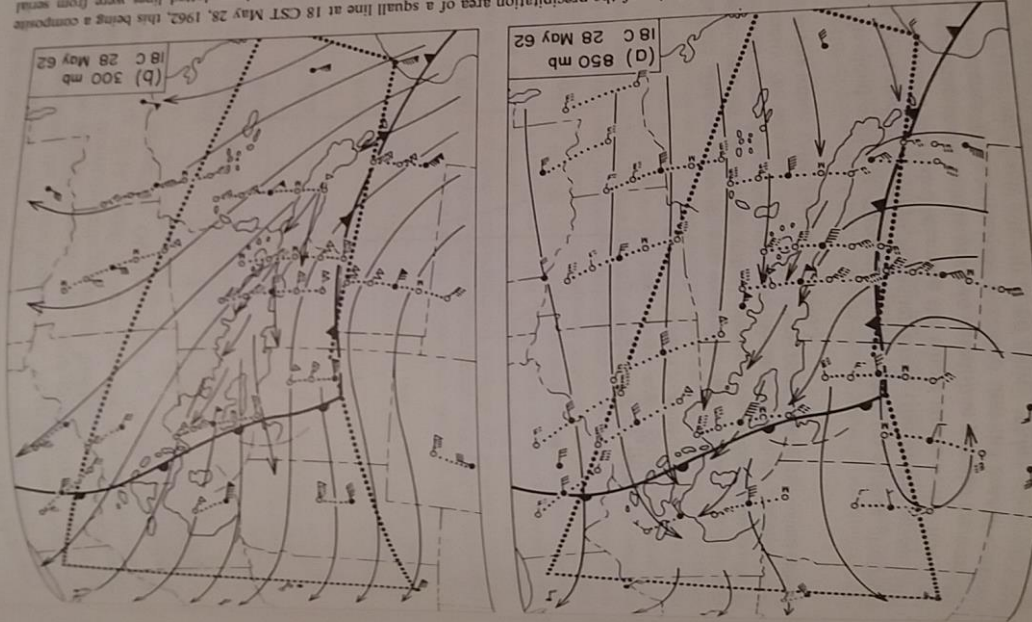


FIG. 13.12 The thin lines represent boundaries of the precipitation area of a squall line at 18 CST May 28, 1962, this being a composite of the radar echoes observed from several stations. The wind observations plotted on open circles along dotted lines were from serial ascents made 3 to 6 hr before and after the time of the figure, displaced according to the speed of movement of the system. Surface fronts and streamlines at (a) 850 mb, and (b) 300 mb. (After Fankhauser, 1965.)

rainfall estimated from analyses of hourly rainfall data. A fairly close agreement was obtained earlier in similar computations by Bradbury (1957).

On a finer scale, the general character of the circulation in a convective system can be deduced from the wind field in its vicinity together with applications of thermodynamic principles. The result of one such analysis is shown in Fig. 13.14, a time section through a large thunderstorm in a squall line that passed over Oklahoma City in late afternoon on May 21,

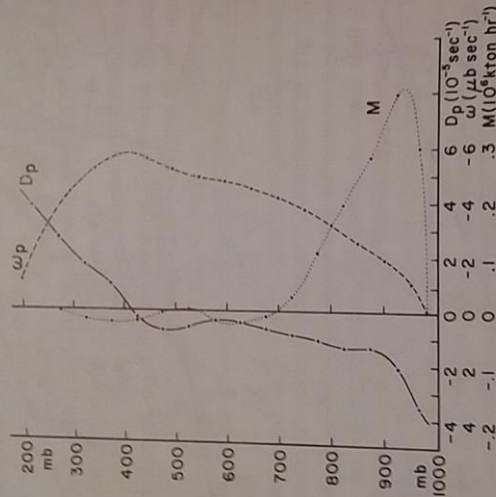


FIG. 13.13 The mean, over the area enclosed by dotted lines in Fig. 13.12, of the horizontal divergence (D_p) and the vertical velocity (ω_p); scales at bottom. Curve M shows the horizontal flux of water vapor per 50-mb layer into the region enclosed by this boundary. This value goes to zero at the lowest point because the volume considered is over sloping terrain. (After Fankhauser, 1965.)

1961. The cloud outline is schematic, being based on surface observations and on radar profiles observed at earlier times (the limited upward tilt of the radar antenna precluded scanning the whole storm at this time). Other storms occurring in the vicinity prior to squall-line passage are omitted for the sake of clarity. Isotach analyses indicated that the general winds near tropopause level were from west 50 to 60 knots. Thus, there was a pronounced veering and an increase of wind speed with height, characteristic

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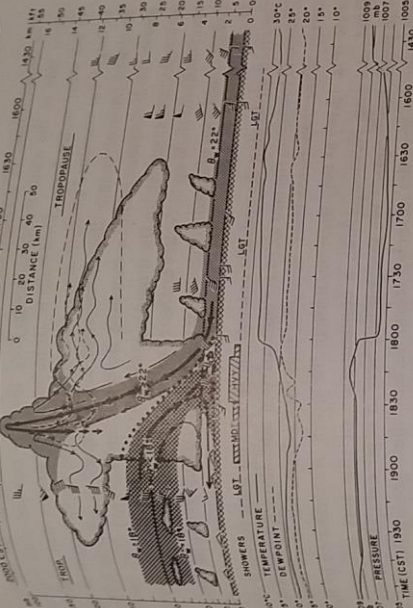
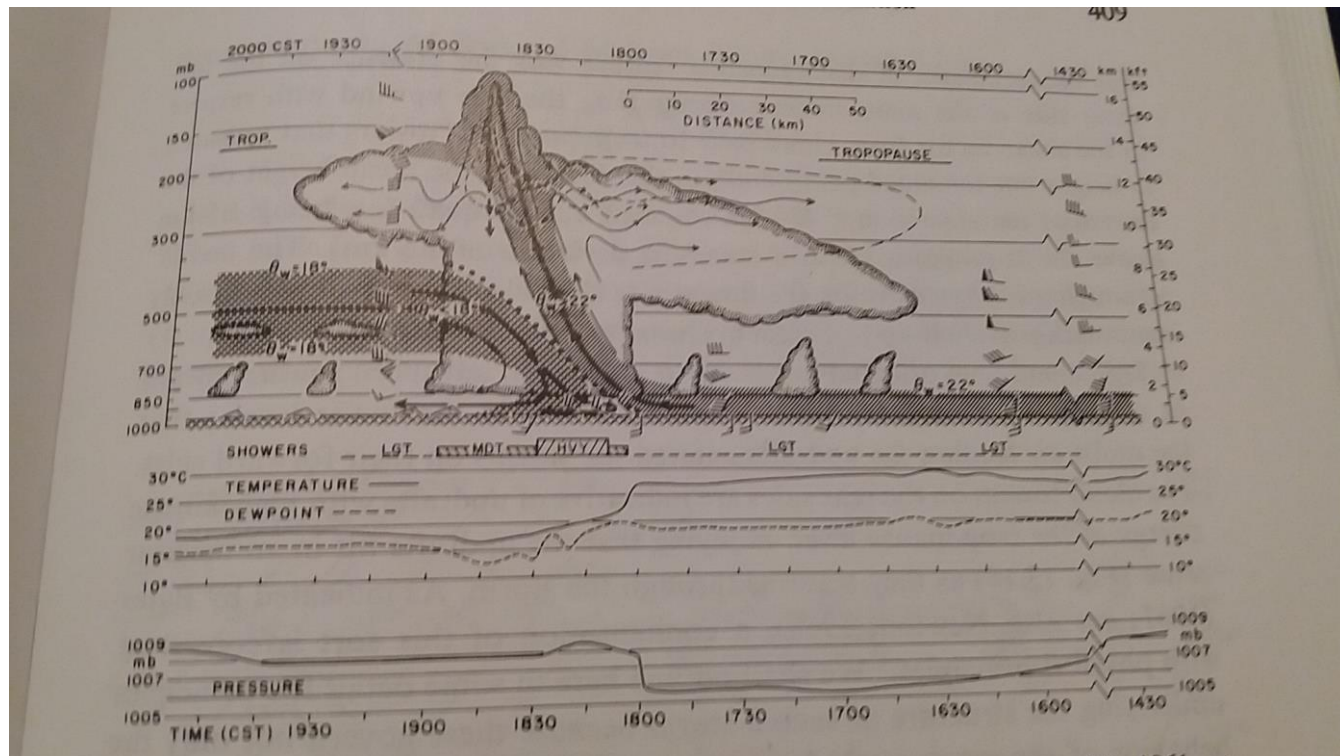


Fig. 13.14 Partly schematic cross section through squall line of May 21, 1961, as it passed Oklahoma City. The vertical exaggeration is fivefold. Winds plotted with directions relative to north at top of figure. Hatching indicates depth of air, with θ_w in excess of 22°C ahead of squall line, and its probable extent in updraft and upper portion of cloud. Crosshatching indicates extent of air, with θ_w less than 18°C , based on sounding behind storm and on surface observations. A similar layer at closely corresponding heights was present before squall-line passage; this is omitted for the sake of clarity. Shown at bottom are periods of precipitation, temperature, dewpoint, and sea-level pressure traces (somewhat smoothed). Heavy arrows, axes of main drafts; light arrows, relative streamlines, dashed where air emanates from core of stratospheric tower. At right, long dashes suggest outline of mass of air plume originating in storm and spreading out essentially horizontally; the radar-detected cloud plume, at lower elevations, consists of small precipitation particles that have partly fallen out of the air plume. The dimensions of the stratospheric tower are about the maximum attained and may at a given time be considerably smaller in a pulsating storm. Anvil length is typical for a storm about 2 hr old. (From Newton, 1966.)

of the environment of most severe storms. In the discussion to follow, the circulations are, for simplicity, treated only in two dimensions.⁴

In this particular squall line, the cloud profile indicated in Fig. 13.14

⁴ The important influences of wind veer with height have been discussed by Browning and Ludlam (1962) and Browning (1964) in relation to the three-dimensional configurations of severe storms and the distribution of hail and rain, and by Newton and Fankhauser (1964) in regard to the direction in which an individual storm moves, which partly controls its intake of water vapor to balance the rainout.



was characteristic. A distinguishing feature was the presence on the upshear side of the main storm columns (i.e., the side upwind with respect to the shear vector from cloud base to tropopause) of towers that pulsed in height but remained persistently in this location (the mean height of the towers of ten storms in a 200-km segment of the squall line being 1.5 km above the tropopause, with a standard deviation of 1.5 km). The anvils, although most extensive on the downshear side, also protruded as a heavily mammatiform shelf up to 50 km upshear from the storm cores.

This squall line, oriented northeast to southwest, advanced southeastward at a speed of 11 m/sec against southerly winds in the lower levels; thus the moist air feeding the updrafts entered the storms on their forward sides. The cloud towers on the rear sides are indicative of updrafts with high kinetic energy at the tropopause level, implying that the updrafts tilt in an upshear sense (Fig. 13.14) as they traverse through the storm. As indicated by Bates (1961; see also Newton, 1966), a configuration of this sort is consistent with physical principles.^a We shall outline here in some detail the principles underlying the structure of such a storm because these govern not only the behavior of the storm itself but also, as will be seen in Section 13.6, the way in which the convective cloud acts to transfer properties significant for the general atmospheric circulation.

At the considerable heights to which air parcels rise above the tropopause, they become colder than the ambient air and must subsequently acquire large downward velocities. Thermodynamic analysis suggests, however, that the downdraft must diminish rapidly in intensity below the tropopause and that *no appreciable amount of the air originating in the updraft is likely to return to the lower troposphere*. The considerations involved are illustrated in Fig. 13.15, in which the thin lines correspond to a composite sounding representative of the undisturbed environment of the storm of Fig. 13.14.

Normand (1946), Stommel (1947), and Schmidt (1947) introduced the important concept of entrainment; the drawing in of air from the environment of a convective cloud and its mixing into the saturated air of an updraft or downdraft modify both thermal properties and vertical momentum of the draft. Because the surface-to-volume ratio is greater for a small than for a large-diameter draft, a small cloud is more influenced by the mixing-in of air through its sides than is a large one. From cloud observa-

^a The upshear canting of the draft is due in part to a strong tendency of the air in a large-diameter draft to conserve its horizontal momentum despite drag forces exerted by the ambient winds, and in part to the movement of the storm as a whole, which causes the air elements during their rise to lag the foot of the updraft on the advancing side of the storm.

tions and theory, Malkus and Williams (1963) concluded that

$$\frac{1}{M} \frac{dM}{dz} \approx \frac{1}{D} \quad (13.1)$$

where D is cloud diameter and M is the vertical mass flux at a given level. "Thus trade cumulus sized clouds ($D \sim 1$ km) entrain a mass flux comparable to their own initial mass flux in 1 km, while thunderstorm sized towers (D up to 10 km) can shoot through most of the troposphere before experiencing one-to-one dilution." In an illuminating numerical experiment, Squires

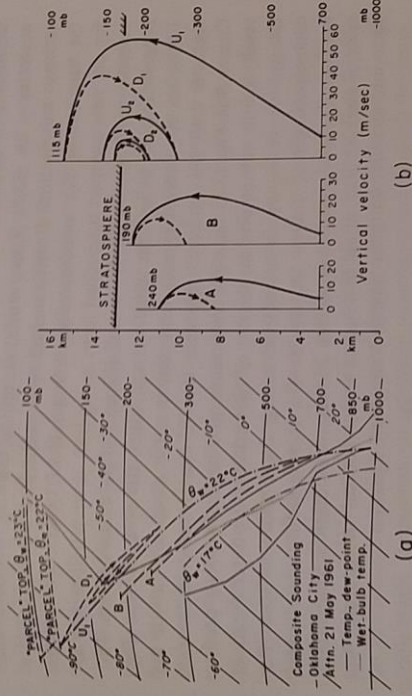


FIG. 13.15 (a) Representative sounding at Oklahoma City, afternoon of May 21, 1961 (thin solid lines, temperature and dewpoint; dotted line, wet-bulb temperature); plotted on a tephigram. For meaning of other curves, see text. (b) Vertical velocities in updrafts (solid lines) and downdrafts (dashed lines) under different assumptions described in text. (From Newton, 1966.)

and Turner (1962) demonstrated the influence of varying cloud diameter and environmental relative humidity upon the vigor of the updraft, its condensed water content, and the vertical development achieved by the cloud.

Since the towers of large cumulonimbus clouds often penetrate to maximum heights predicted from "parcel" theory, the air in the *cores* of their updrafts is evidently at times undiluted by mixing with the environment. However, since the diameters of drafts even in large storms are not likely

to exceed about 10 km. Eq. (13.1) suggests that significant entrainment is to be expected. Consequently, the outer sheaths of drafts must undergo strong mixing. Then, under the influence of buoyant accelerations, the air in an updraft core would rise to the greatest height, whereas air parcels in the outer portions of the draft, having decreased buoyancy due to chilling, would not rise so far.

Curves U_1 in Fig. 13.15 show the characteristics of an unmixed updraft parcel, having $\theta_w = 22^\circ\text{C}$ corresponding to the average in the initial layer below 700 mb. Such a parcel would rise to about 16 km. With a draft diameter of 7.5 km, the overall entrainment into the draft would be 1/3 km, according to Eq. (13.1). If, as a crude assumption, the inner protected core occupied half the draft diameter and all the entrained air were mixed into the outer sheath, this would correspond to a rate of 1/6 km. To account for the mass transport determined later, an updraft is required whose diameter is much smaller than the overall diameter of the cloud volume. This suggests that the bulk of the cloud through which the updraft rises (Fig. 13.14) has only weak buoyancy, and thus that the cloud air outside the vigorous draft has a temperature near that of the storm environment. Curves B in Fig. 13.15 show the temperature and vertical velocity profiles resulting from entrainment (at the rate of 1/6 km) into the outer sheath of the updraft, of cloudy (saturated) air having these characteristics. For comparison, curves A show the results of a corresponding entrainment, but from the unsaturated environment. In these computations the entrainment of inert air (i.e., of air that has no initial vertical momentum) was taken into account, using the modified buoyancy equation

$$\frac{d}{dt}(wm) = mg - \frac{\Delta T}{T} \quad (13.2)$$

where ΔT is temperature excess over the environment and m is the mass of an air parcel, initially a unit mass, that is augmented according to the entrainment rate.

At the height of its maximum upward penetration, an air parcel is colder than its environment and is subjected to a downward buoyancy acceleration. Analyses of the motions of cloud elements (e.g., by Anderson, 1960; Sundbørg, 1961) suggest that air rising in the cores of cloud towers subsequently diverges outward, making it susceptible to mixing with the environment. Thus, although some air may rise undiluted, it is very unlikely that the returning downdraft is not subjected to some degree of mixing. For the case U_1 , the further behavior of an air parcel entraining at the rate of 1/6 km is indicated in Fig. 13.15. In this case, it undergoes an oscillation

that is damped partly because it gradually approaches the environment temperature in the process of entrainment and partly because inert air is incorporated, according to Eq. (13.2). Even with the very small mixing rate of only 1/20 km, a downdraft parcel could descend no lower than to 340 mb, and after damped oscillation would approach equilibrium near the 350-mb level.

The conclusion from this analysis is that *essentially all air in the updraft must, after undergoing oscillations, remain in the upper troposphere, eventually passing into the expanding anvil of the storm.* This behavior, and the rising of air in different parts of the draft to varying heights depending on their different degrees of mixing, is indicated schematically by the streamlines in Fig. 13.14. It is obvious, of course, that all these oscillations cannot coexist; rather they would interfere with each other and be expressed as "turbulence."

The spreading out of air in the surface layers beneath thunderstorms (Section 13.3) indicates that there must be a substantial downward mass flux in the lower part of the cloud system. In this example, the lowest θ_w in the surface layers was 17°C. It is apparent from Fig. 13.15a that the cold air beneath the thunderstorm must have originated in the middle troposphere where the environmental θ_w was correspondingly low. It may then be inferred that there is a separate downdraft circulation in the lower part of the cumulonimbus (Fig. 13.14), as indicated in other analyses by Newton (1950), Bates (1961), Fuls (1962), and Browning and Ludlam (1962).

On entering the storm, the air taking part in the lower downdraft circulation has moderately high momentum characteristic of the environment in middle levels. Its momentum enables it to move forward through the storm, converging with and undercutting the warm air in advance of it. *The mechanical lifting provided by the cold downdraft thus continually regenerates the updraft, prolonging the storm circulation.* The vertical circulation in the lower downdraft is caused by gravitational sinking, its coldness being due to chilling by partial evaporation of water condensed in the sloping updraft and falling through the downdraft. Normand (1946) recognized the desirability of some such process, pointing out that the energy is greatly augmented if a thunderstorm is "organized to take in potentially cool air at high levels as well as potentially warm air at the lower levels."

The principle was discussed earlier by Normand (1938). Desai and Mal (1938) recognized its applicability to the severe springtime thunderstorms of northeast India. They pointed out that the fall of both atmospheric temperature and wet-bulb temperature, observed at the ground with the onset of these storms, could be explained by evaporative cooling of dry

continental air descending from levels above the moist layer. Harrison and Orendorff (1941) were the first to describe the essential structure of squall lines over the United States, observing that the squall line is a self-regenerative mechanism of the character described above.

The present conception of a highly organized, severe storm circulation has been most strongly influenced by the analysis carried out by Browning and Ludlam (1962) of a giant hailstorm over England. Through an ingenious and thorough interpretation of radar and other observations, they were able to deduce the main features of the storm circulation and trace its influence on the growth and redistribution of hail and other hydrometeors. They showed also that in its early and late stages, the storm was composed of numerous cells that propagated in a systematic manner related to the vertical shear, as suggested by Newton and Newton (1959). During the time when the storm was best developed and disgorged huge quantities of hail, however, it was dominated by a persistent steady circulation, basically similar to that in Fig. 13.14, but with a more erect updraft in upper levels.

13.5 MASS AND WATER BUDGETS

The mass budget of a squall-line thunderstorm may be estimated from the winds in its environment together with the considerations mentioned in Section 13.4. For a "solid" squall line^a moving with speed c , the fluxes of air and of water vapor into or out of an individual storm are given by

$$\int (c - v_n) D \frac{dp}{g}; \quad \int (c - v_n) q D \frac{dp}{g} \quad (13.3)$$

where v_n is the wind component along the direction toward which the squall line is advancing, q is specific humidity, and D is distance along the line between thunderstorms, taken to be the width of an individual storm. The vertical profiles in Fig. 13.16 indicate a flux of air (and water vapor) into the forward side of the storms below 700 mb, a flux outward from the rear side below 810 mb, a weaker flux into the rear side between 360 and 810 mb, and a strong efflux above 360 mb where the balloon passed through the anvil.

^a Squall lines are ordinarily composed of discrete thunderstorms, which as individuals conserve their identities for long periods of time. In the present example, thunderstorms were spaced at average intervals of 20 km along the line. Although the radar-detected cores of heaviest precipitation were distinct, the cloud mass between these was continuous, as was the squall front at the surface.

13.5 Mass and Water Budgets

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Based on expressions (13.3), Fig. 13.16, and the scheme of Fig. 13.14, the mass fluxes in various branches of a single large thunderstorm are shown in Fig. 13.17. Of the 700 kton/sec of air entering the updraft in low levels, 20 to 60% circulates through the tower above the tropopause,⁷ the remainder rising to more limited heights in the strongly entraining outskirts of the updraft and spreading out directly into the anvil. An updraft mass

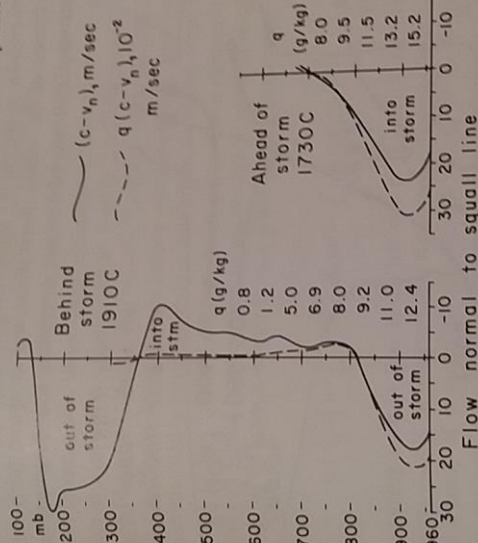
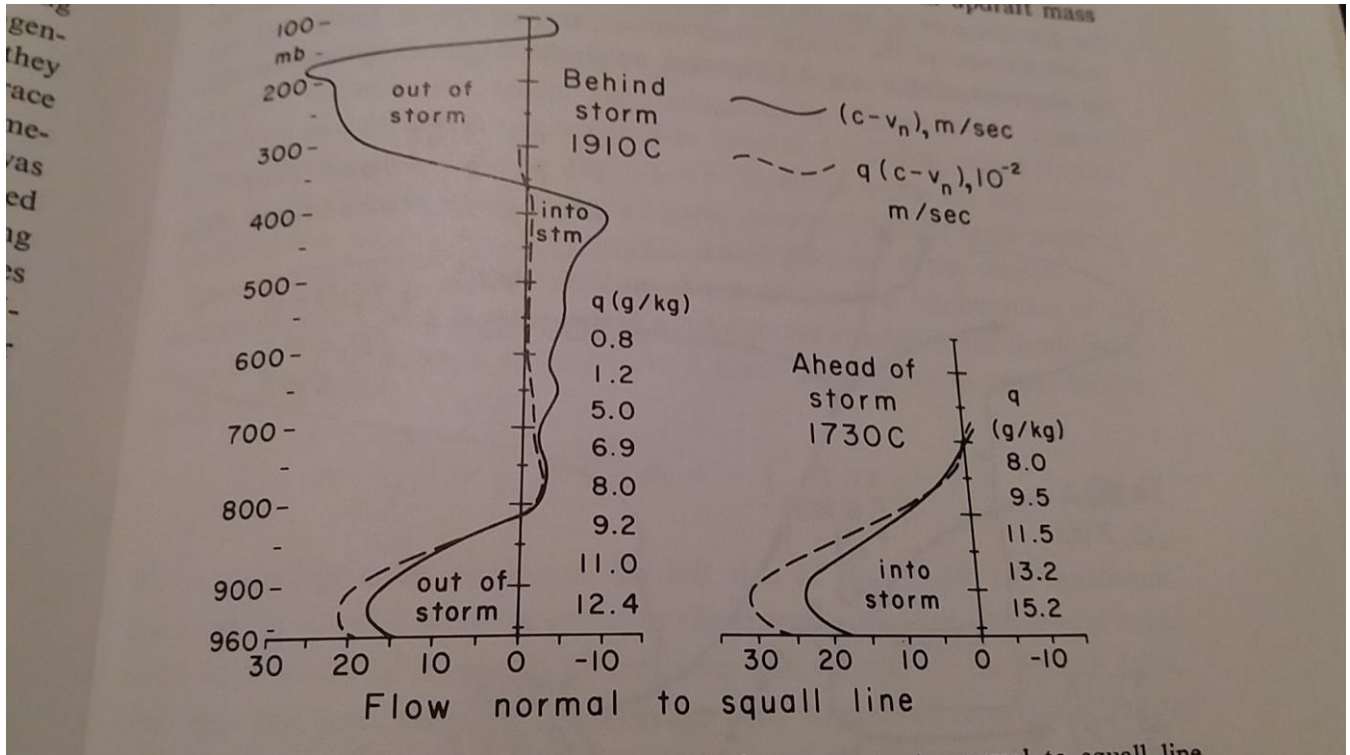


Fig. 13.16 Vertical profiles of relative velocity components normal to squall line (solid) and of the product of relative velocity and specific humidity (dashed), based on soundings at 1730 and 1910 CST in Fig. 13.14. (From Newton, 1966.)

flux of this magnitude would lead to an expansion of the anvil volume at the rate of $2 \text{ km}^3/\text{sec}$ (consistent with observed rates of anvil growth). The mass flux in the lower-tropospheric downdraft, 400 kton/sec, is about 60% of that in the updraft.

The flux of water vapor entering the foot of the updraft is 8.8 kton/sec. With a mean mixing ratio of 0.9 g/kg and an assumed water substance

⁷ This estimate was made by an indirect method described by Newton (1966). For an average storm with strongly pulsating towers, a mean flux of 150 kton/sec upward and downward through the tropopause is realistic. For a "steady state" storm whose tower is broad and remains persistently high above the tropopause, as much as 400 kton/sec of air may circulate through the stratospheric tower.



density of 0.3 g/m^3 , 0.6 kton/sec of vapor and 0.6 kton/sec of water are exported in the anvil expansion, leaving 7.6 kton/sec for precipitation aloft in the main storm column. From Fig. 13.16, the vapor flow aloft into the rear of the storm is 0.7 kton/sec , and the outflow toward the rear in low levels is 4.3 kton/sec , requiring 3.6 kton/sec of this rain to be evaporated into the downdraft. This leaves a residual of 4.0 kton/sec water available for precipitation at the earth's surface. Allowing for uncertainties in representativeness of the various observations, this compares fairly well with the observed rainfall rate, 4.7 kton/sec , determined by averaging the precipi-

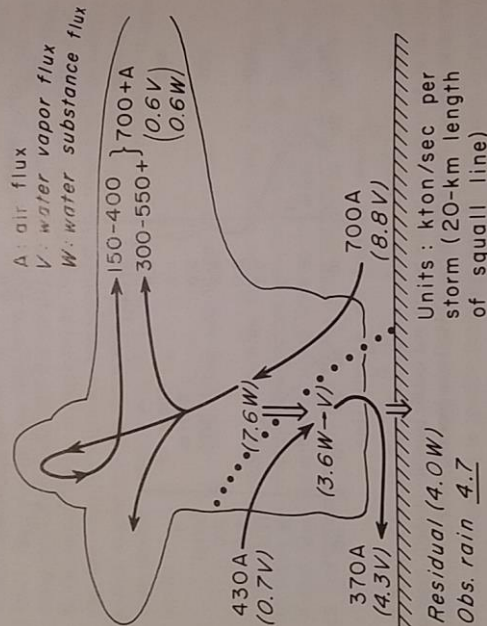
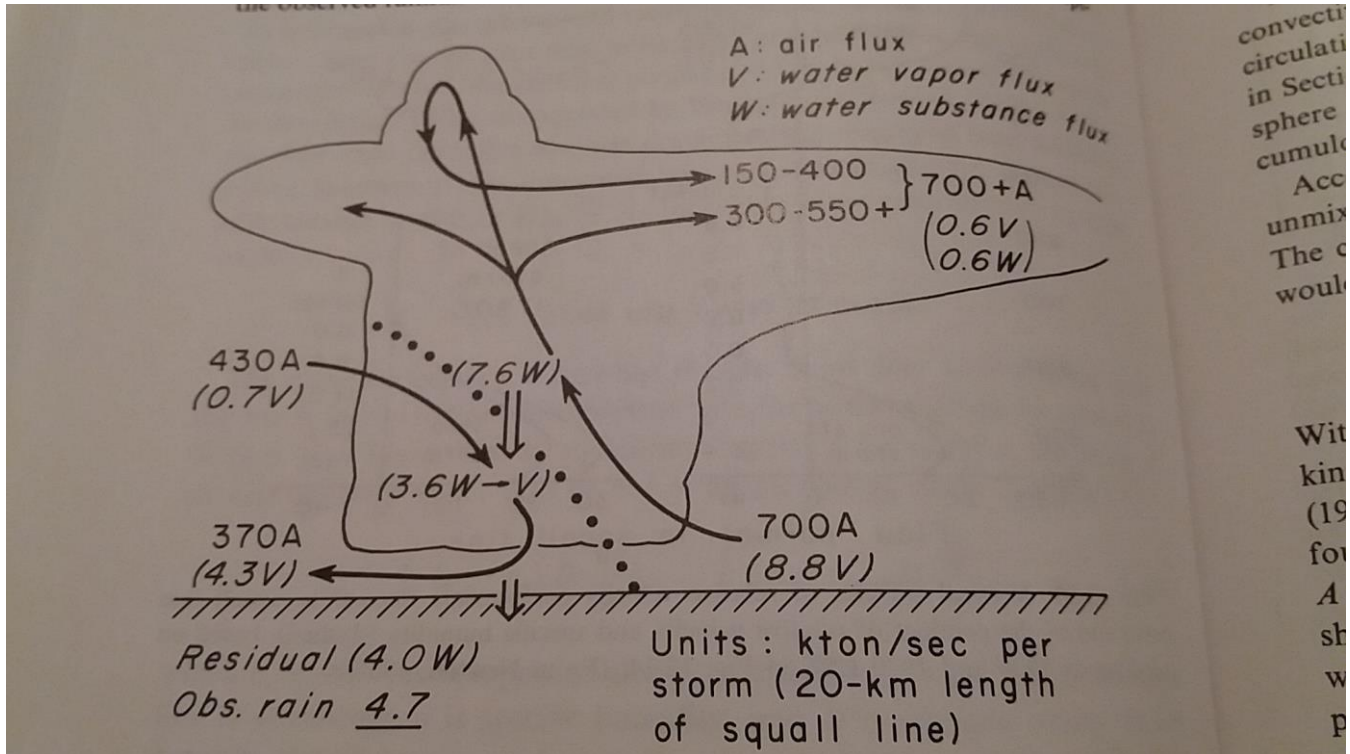


Fig. 13.17 Principal branches of the circulation, with fluxes of air (A), water vapor (V), and water (W), corresponding to a single thunderstorm in squall line. (From Newton, 1966.)

tation over the U.S. Weather Bureau "Beta" mesometeorological network south of Oklahoma City, over which the squall line passed.

It should be noted that the mass fluxes in Fig. 13.17 are representative of a large intense thunderstorm; they are more than an order of magnitude greater than the fluxes determined by Braham (1952) for typical "air mass" storms. Also, Braham found in the latter that the surface rain was only about 10% of the vapor flux through the storm, while in the large storm



discussed above, the rain reaching the ground was over 50% of the vapor flux entering the updraft. It may also be noted that since the influx and efflux almost balance within the upper troposphere, the amount of rain deposited is essentially equivalent to the moisture convergence in the moist layer below 70 mb (compare Fig. 13.13).

13.6 VERTICAL TRANSFER OF PROPERTIES

The vertical transports of heat, angular momentum, and water vapor in convection systems have never been evaluated in relation to the general circulation as a whole. As mentioned in Section 2.5 and discussed further in Sections 14.4 and 17.2, the heat transfer from the low to the high troposphere in the Hadley circulation is largely accomplished by high-reaching cumulonimbus clouds covering only a small portion of the tropics.

According to the "parcel" theory of convection, the temperature of an unmixed parcel would be prescribed by the moist adiabat from cloud base. The change of its kinetic energy from a lower level z_0 to an upper level z would be

$$\Delta \left(\frac{w^2}{2} \right) = \int_{z_0}^z g \frac{\Delta T}{T} dz = - \int_{p_0}^p R \Delta T d \ln p \quad (13.4)$$

With a sounding such as that in the left side of Fig. 13.18, the maximum kinetic energy should be achieved near the tropopause level. Petterssen *et al.* (1945), in a study of cumulus clouds over the vicinity of the British Isles, found that the tops rose on the average only to a level (corresponding to Δ in Fig. 13.18) where, according to the parcel concept, the buoyancy should be greatest. This was so also in the sample shown in this figure, in which only a very small fraction of the clouds grew to stratospheric heights predicted by parcel theory.

The "slice" concept (Bjerknes, 1938; Petterssen, 1939), wherein the environment is warmed by dry-adiabatic descent of the air between clouds (thereby diminishing their excess temperature), can partially account for the limited developments of cumuli. Stommel's findings (1947) indicate that the diminution of their buoyancies through entrainment has a great additional influence; Eq. (13.1) indicates that only a favored few clouds that manage to grow to large diameter can survive the erosive influences of entrainment and penetrate to high levels.

Thus, in any approach to the appraisal of heat transfer by cumulus convection, it will be essential to determine the spectra of their sizes under varying environmental conditions and to take into account the differences

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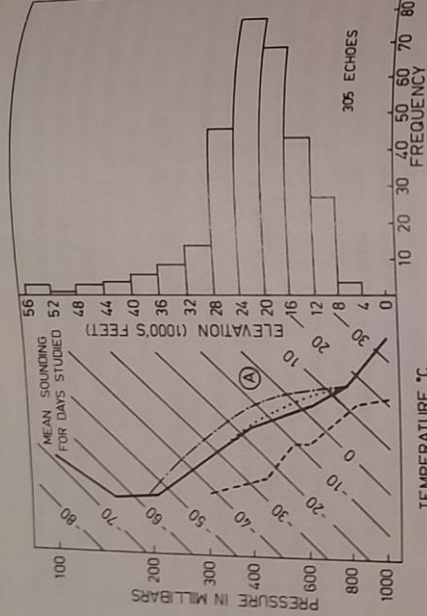


FIG. 13.18 Frequency distribution of the greatest heights reached by individual radar-echo tops in convective clouds over Texas. The mean sounding (temperature, solid; dewpoint, dashed) for the ten warm-season days studied is shown at left. Added dotted line is schematic; shows qualitative influence of strong entrainment. (After Clark, 1960.)

of thermal and kinematic properties with cloud size.⁸ In a crude sense, the vertical eddy-heat transport over a given horizontal area A is given by $\int \rho w_c \Delta T dA$. Since, according to Eq. (13.4), the vertical velocity is proportional to $\int \Delta T^{1/2} dz$, the vertical heat flux (say, through the 500-mb level) should be roughly proportional to

$$\int \int \Delta T^{1/2} dA dz$$

Thus the large clouds, whose extensive horizontal areas tend to be associated with large buoyancies and great vertical development, must predominate in the mechanism of vertical heat transport between the lower and upper troposphere.

⁸ A first attack on this problem has been made by Kasahara and Asai (1967) on the basis of a population of clouds of uniform size. The size of an individual cloud can be varied, and both entrainment and the compensating descent between clouds are taken into account in their model. This model was devised to study the two-way interaction between the clouds, as a mechanism for transporting and generating energy, and the influence of the large-scale motion field upon the production of clouds.

As far as the whole atmosphere is concerned, the heat gain from a single large storm, given by the latent heat of condensation times the observed mass rate of precipitation at the ground, would (for the case in Fig. 13.17) be 2.8×10^{12} cal/sec. From data derived from radar measurements for eight squall lines as given by Austin *et al.* (1961), an equivalent average value of 2.45×10^{12} cal/sec (per 20-km length of squall line) was determined, which is closely comparable. The vertical energy transfer may alternatively be estimated from the expression

$$\int \rho w(c_p T' + Lq' + gz') dA \approx \sum F_z(c_p \bar{T}' + Lq' + gz') \quad (13.5)$$

where the primes indicate deviations of temperature, specific humidity, and height at an isobaric surface within the storm from the values in the environment,⁹ and F_z is the mass flux (positive upward) in a branch of the circulation. The value of the height deviation z' is unknown, but inclusion of values as large as 50 m influence the result by less than 10%.

Assuming properties prescribed by Figs. 13.17 and 13.15a, with saturation in both updraft (at $\theta_w = 22^\circ\text{C}$) and downdraft (at $\theta_w = 17^\circ\text{C}$), the upward energy flux is found to be 1.6×10^{12} cal/sec through the 700-mb surface (where both updraft and downdraft fluxes must be considered) and 2.2×10^{12} cal/sec through the 450-mb surface (decreasing at higher levels). Allowing for the discrepancy between the results of the moisture-flux budget and the observed rainfall in Fig. 13.17, the foregoing evaluation suggests that the heat transported upward in the middle troposphere is essentially equivalent to the total heat realized from condensation, this being transferred into and remaining in the upper troposphere. In the preceding computations the heat of fusion has not been considered. The freezing and subsequent sublimation of small ice particles entirely in the upper troposphere would evidently not influence the heat transfer. On the other hand, if all the precipitation fell as hail (a condition approached in some storms), the heat transfer would be augmented by 13%, since freezing occurs in the upper troposphere with melting in the lower troposphere or at the ground.

Fankhauser (1965) found that over a 6-hr period, the squall line in Figs. 13.12 and 13.13 precipitated 8.4×10^{12} kg of water. The corresponding release of condensation heat, 23×10^{13} cal/sec, may be compared with the

⁹ In expressions (13.5), the upward mass transport through an isobaric surface within the storm is assumed to be compensated by an equal downward mass transport outside it of air having the properties of the environment. Since this compensation may take place distant from the storm, the latter condition may not be satisfied exactly. On the right side, the use of mean (barred) values implies a neglect of small eddies and of systematic variations of properties across the drafts.

value of 20×10^{15} cal/sec found by Palmén and Holopainen (1963) for the partially nonconvective rainfall shield of a winter cyclone. Since, as indicated earlier, virtually all upward mass flux in the cyclone of Fig. 13.12 took place in the convective system, it is likely that this was also true of the vertical heat flux. In such a cyclone with a well-developed convective system, it is often observed that little ordinary frontal precipitation occurs, presumably because (Fig. 13.12a) the warm, moist air is intercepted and overtaken before it reaches the fronts.¹⁰

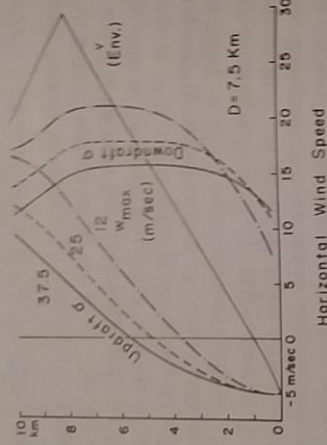


FIG. 13.19 Horizontal wind speed (σ) within the updraft of a thunderstorm, embedded in an environment wind field V , for cases wherein the maximum vertical velocity (at tropopause level) has the three indicated values. The horizontal velocities indicated for the downdraft were calculated on the supposition that the downdraft extends through the whole depth of the cloud. As indicated in Figs. 13.14 and 13.15, the downdraft must in reality consist of two separate branches, one in the upper part of the storm and the other entering the storm in the middle troposphere. Thus, the downdraft wind profile is qualitatively valid only in the uppermost levels and in the lower troposphere. (After Bates and Newton, 1965.)

Only a crude estimate can be made to illustrate the effect of thunderstorms in transferring momentum vertically. Figure 13.19 shows a computation (Bates and Newton, 1965) of the horizontal speed within an updraft, which would result if a draft of a 7.5-km diameter were subjected to drag forces

¹⁰ In the cyclone model developed by Bjerknes and Solberg (Fig. 5.1) from observations in northwestern Europe, the precipitation was well organized along fronts, and very little rain was assumed to occur in the warm sector. Their classical model is in this respect not applicable to some cyclones in lower latitudes (e.g., in the United States) because these often contain potentially unstable air not characteristic of cyclones in higher latitudes, particularly in the cooler seasons.

while rising through an environment with vertical shear fairly typical of severe storm situations (an updraft of this diameter would, with a vertical velocity of 25 m/sec at 500 mb, carry a mass flux of the magnitude in Fig. 13.17).

If we assume that the upward mass flux in the storm is compensated by a downward mass flux in the environment *at the same latitude*, the contribution to the vertical flux of earth angular momentum may be neglected, assuming also that the downward flux of relative angular momentum is characterized by the wind velocity in the immediate storm environment. The upward flux of angular momentum may then be approximated by

$$\int \rho v u a \cos \varphi \, dA \approx \sum F_z \bar{u}' a \cos \varphi \quad (13.6)$$

where the conventions are as in Eq. (13.5).

In the middle troposphere (say, at 500 mb), $\bar{u}' \approx 0$ in the downdraft; in the updraft $\bar{u}' \approx -15$ m/sec according to Fig. 13.19, and $F_z = 700$ kton/sec from Fig. 13.17. The corresponding angular-momentum flux for a single storm of this magnitude is then -5.5×10^{23} gm cm² sec⁻² through the 500-mb surface at latitude 35°. In the lower troposphere, momentum is carried downward by the downdraft and air with deficient momentum (though by a smaller magnitude than that above) is carried upward, the computed net transfer at 700 mb being nearly the same as that at 500 mb.

In the squall line of Fig. 13.12 (nearly 1500 km in length), the total rainfall rate evaluated by Fankhauser was about 80 times that of the single storm represented in Fig. 13.17. The vertical shear of the west-wind component was about half as great as in Fig. 13.19. Correspondingly, assuming the vertical mass flux to be proportional to the mass rainfall rate, the vertical flux of angular momentum through the middle troposphere, accomplished by the entire squall line, would be -2.2×10^{26} gm cm² sec⁻². It must be emphasized that this is a raw estimate.¹¹ While it can probably be taken as typical for a well-developed squall line, it should also be noted that a squall line of this magnitude is a special event.

¹¹ Note also the uncertainties arising from the assumptions stated before Eq. (13.6). We may, however, compare a cyclone in which convective ascent occurs with one in which the ascending air is stable and rises slowly. In the latter case, the horizontal velocity accommodates nearly to the general wind field. With a wind profile as in Fig. 13.19, there would in this case be a large upward transfer of momentum through the middle troposphere, whereas the corresponding upward transfer by convective clouds would be small. This comparison concerns only the transfer within the precipitation region; the overall vertical transfer of momentum depends in either case upon the properties of the compensating air that descends elsewhere.

A complete evaluation of the momentum transfer by thunderstorms would require a knowledge of their populations and consideration of the different conditions under which they occur. However, it may be pointed out that the average torque required to balance the angular-momentum flux for the whole latitude belt between 30° and 50°N (Table 1.1) in summer is $-6 \times 10^{25} \text{ gm cm}^2 \text{ sec}^{-2}$; in winter $-31 \times 10^{25} \text{ gm cm}^2 \text{ sec}^{-2}$. The value given above for a single squall line is 37% of the required amount in summer and 7% in winter, suggesting that *vertical angular-momentum transfer by convective storms is very significant during the warmer months; it is probably of little significance compared with other processes in winter when the frequency of large convective systems is comparatively low.*

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