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NOAA Technical Memorandum NWS NSSFC-5

THE OPERATIONAL METEOROLOGY OF CONVECTIVE WEATHER
VOLUME I: OPERATIONAL MESOANALYSIS

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Kansas City, Missouri

November 1982

U.S. DEPARTMENT OF COMMERCE / National Oceanic and Atmospheric Administration / National Weather Service

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" VOLUME I: OPERATIONAL MESOANALYSIS

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Kansas City, Missouri

November 1982

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PREFACE TO

THE OPERATIONAL METEOROLOGY OF CONVECTIVE WEATHER
VOLUME I: OPERATIONAL MESOANALYSIS

Primary causes are unknown to us; but are subject to simple and constant laws, which may be discovered by observation, the study of them being the object of natural philosophy.

-- Joseph Fourier, Theory of Heat

There is no other species on Earth that does science. It is, so far, entirely a human invention ... It has two rules. First: there are no sacred truths; all assumptions must be critically examined; arguments from authority are worthless. Second: whatever is inconsistent with the facts must be discarded or revised ... The obvious is sometimes false; the unexpected is sometimes true.

L.F. Richardson was a British meteorologist interested in war. He wished to understand its causes. There are intellectual parallels between war and weather. Both are complex. Both exhibit regularities, implying that they are not implacable forces but natural systems that can be understood and controlled. To understand the global weather you must first collect a great body of meteorological data; you must discover how weather actually behaves.

-- Carl Sagan, Cosmos

There is a growing accumulation of evidence to indicate that man has no direct contact with experience per se but that there is an intervening set of patterns which channel his senses and his thoughts, causing him to react one way when someone else with different underlying patterns will react as his experience dictates.

It is time, however, that we began to realize that much of what passes for science today may have been scientific yesterday but can no longer qualify because it does not make any additional meaningful statements about anything. It blindly adheres to procedures as a church adheres to its ritual.

-- E.T. Hall, The Silent Language

We can never have enough of nature. We must be refreshed by the sight of inexhaustible vigor, vast and titanic features, the sea-coast with its wrecks, the wilderness with its living and its decaying trees, the thunder

cloud, and the rain which lasts three weeks and produces freshets. We need to witness our own limits transgressed, and some life pasturing freely where we never wander.

-- Henry David Thoreau, Walden

These notes have been developed in an effort, however imperfect, to acquaint meteorologists in an operational environment with the basic concepts of convective weather systems. It is a sad fact of life that many of today's operational meteorologists have never been given a physical interpretation of the dynamics which are understood to govern the atmosphere and, in particular, convection. It is not my intent to be completely exhaustive, although the length of the text leads me to fear that it may be exhausting!

There are numerous threads which can be used to sew up the package I am trying to deliver. In trying to unravel them, I have at times assumed the reader knows things with which he/she may not, in fact, be familiar. Conversely, I have at times assumed the reader's ignorance of some basic ideas which I have felt important enough to explain in detail and, in the process, may have bored more advanced readers. I hope that both forms of exasperation never reach the breaking point.

In any work of this sort, it is easy to find the material one wrote a few months before somewhat less than satisfactory in light of new findings, recent publications, or just plain further thought. One has to stop the process of revisions somewhere, but I suspect we are at the start of an exciting new era in applied meteorology and here I am trying to summarize the proverbial "state of the art". Since I cannot hope to be completely up-to-date by the time this reaches the hands of the readers, I have tried to give enough material to bring the interested reader to the point of a self-sustaining, self-education process. If the reader is content to absorb only what is in these notes, my effort will not have succeeded.

While this preface is being written, Volumes II and III are still embryonic. The reader will note that there are many references to other parts of the text within the body of these notes. These internal references follow an outline-type of structure of the form I.III.A.3.b..., where the leading, underscored Roman numeral refers to the volume number. This is omitted when the reference is within the given volume. The second Roman numeral refers to the chapter in the volume, the capital letter to the sub-heading, and so forth. Since the second and third volumes are not

yet finished, I can only promise that they will be completed as rapidly as possible. Because these self-references generally concern amplifications or additional discussions of the referenced topics, it should not be terribly detrimental for them to be as yet unavailable. If the material were essential, it would have been included at that point in the notes.

The reader should also note that all footnotes in a given chapter will be collected at the end of that chapter. This is not the most convenient approach, but it happens to solve a nasty problem in trying to fit these notes into a readable text. My apologies for any inconvenience.

As in any large work, numerous contributors have made these notes possible. The Chief of the Techniques Development Unit of NSSFC, Dr. Joseph T. Schaefer, has perhaps been most valuable as an encourager (and occasional pushes to complete this work are appreciated), a sounding-board for many of the topics contained herein, an editor, and a respected colleague. Dr. Robert A. Maddox of NOAA's Environmental Research Laboratories, Office of Weather Research and Modification, has provided many ideas, inspiration, and the encouragement only a "kindred spirit" can provide. The Deputy Director of the National Weather Service Training Center, Mr. Larry Burns, gave me the initial support to undertake this effort and confirmed my perception of the need for it in the first place. Numerous individuals have encouraged me by their interest, including Alan R. Moller (NWSFO, Fort Worth, Texas), Larry Wilson, Steve Weiss, Jim Henderson, and Mike Streib (all at NSSFC), as well as the usual host of those "too numerous to mention." Valuable reviews were provided by Profs. Walter J. Saucier, David A. Barber (North Carolina State Univ.), and Richard J. Reed (Univ. of Washington). Naturally, any errors and misinterpretations are my sole responsibility. Finally, Beverly Lambert has suffered through the numerous revisions and drafts and done an outstanding job with the manuscript preparation.

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Kansas City, Missouri
November, 1982

THE OPERATIONAL METEOROLOGY OF CONVECTIVE WEATHER
VOLUME I: OPERATIONAL MESOANALYSIS

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I. Introduction

A. Preliminary Remarks

Operational mesoanalysis is most often considered in the context of convective storms. Mesosystems significant to operational forecasting do not only encompass deep convection, as the patterns of heavy snowfall sometimes suggest, for example. Also, it is not clear that the process of mesoanalysis for convective storms can be transferred totally for application to, say, winter storms, although good analysis techniques are required in both areas. In any case, these notes will not address mesoanalysis associated with non-convective weather.

In order best to accomplish operational mesoanalysis, one should have a thorough understanding of synoptic-scale meteorology. Further, one should be familiar with convective storms and their dynamics. This is easy to say, but difficult to satisfy. No one person has a complete understanding of either one of these areas, especially the latter. Much remains to be learned about convective storm dynamics. Regrettably, there seems to have been a trend away from synoptic meteorology, both in the universities and within the operational arena as well (Doswell et al., 1981). Increasing dependence on numerical models has led to an overall decline in the skills of the synoptic meteorologist (Snellman, 1977). Additional evidence for this decline can be seen in the frequent reference here to texts and journal articles published in the 1950s. If more recent references were available, they would have been used, but the lack of interest in relating dynamic to synoptic meteorology (and vice versa) over the last two decades has led to the paucity of more recent references.

Realistically, these notes cannot provide a working knowledge of both synoptic meteorology and the dynamics of convective storms. Material in these areas will be covered, as it relates to the process of mesoanalysis, but the reader is urged to pursue these topics further by consulting the bibliographic references. Some general discussions of mesoanalysis are contained in Fujita et al. (1956), Magor (1959), Tepper (1959), and Fujita (1963). Pieces of the material concerning practical mesoanalysis are contained in the references, but to the author's knowledge, these have not been collected in one place. The presentation in these notes is essentially qualitative and non-mathematical, since a rigorous discussion is not necessary to the practicing mesoanalyst. Many ideas are presented without proof, but it is hoped that the reference material will be consulted when doubts arise.

B. Scaling Concepts

Under the general heading of "Operational Mesoanalysis" in these notes, a substantial variety of phenomena and concepts is presented. It is worthwhile to discuss this in terms of meteorological scales at the outset.

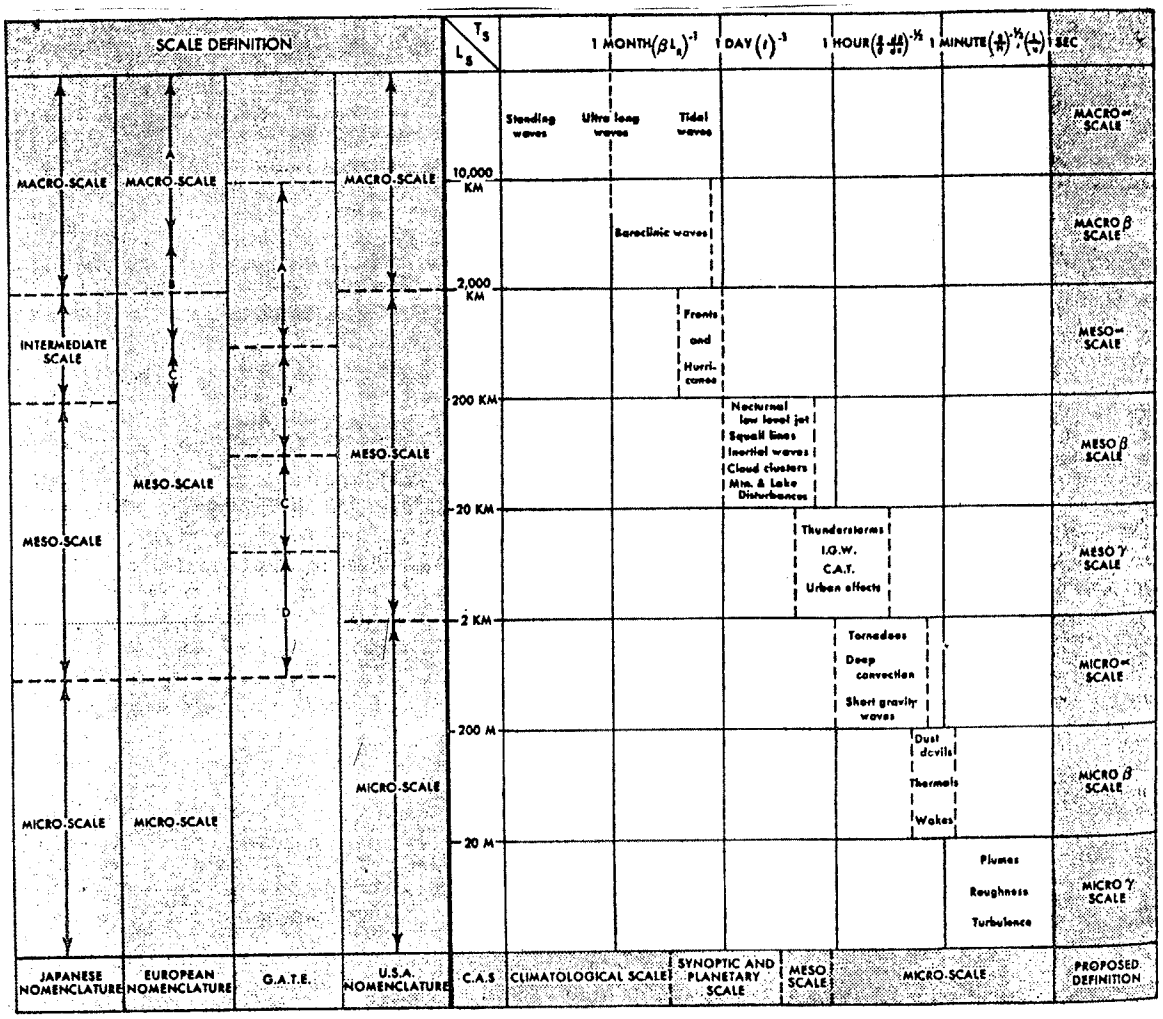


Fig. 1.1. Scale definitions and different meteorological phenomena with characteristic temporal and horizontal spatial scales (after Orlanski, 1975).

It should be emphasized that the notion of scaling is absolutely essential to understanding current and future meteorological thinking. We shall attempt to review current concepts on scales, from that of the extratropical cyclone (ETC) down to those phenomena at the observation limits of the present network of routine surface reports. One example of a proposed ordering of meteorological phenomena by scales is shown in Fig. 1.1.

The large-scale limit to our discussion can be given by some arbitrary order of magnitude estimates for scaling lengths (say, horizontal lengths of 10^3 km, vertical depths of 10 km, and time scales of 10^5 s [~1 day]. Note that these three values, suitably manipulated (as in Haltiner and Williams, 1980), can yield approximate values for most of the terms in the equations governing large-scale flows. The suitability of the manipulation hinges, in large measure, on knowing the answers we want before we begin. In other words, a formal scale analysis is essentially a way of justifying making mathematical assumptions to describe theoretically a problem for which we already have observed the answer! In the process, we can gain insights which may have not been previously obvious, and considerable physical understanding can be gained. Perhaps the most successful application of scale analysis is in the problem of our large-scale limit, the extratropical cyclone.

However, such a formal approach may not be the easiest to understand from an operational viewpoint and it suffers from a major deficiency: namely, on our lower scale limit, we do not have as clear a picture of the desired answer to be obtained. Instead, we consider a more physically-motivated way of establishing the scale of phenomena which draws heavily from the discussions by Emanuel (1980). By doing so, it is hoped that the reader can relate the discussion to observed daily weather events and will therefore be encouraged to pursue the topic as more formally developed in the references (Holton, 1979; Palmen and Newton, 1969; Haltiner and Williams, 1980).

It is convenient that our upper scale limit is the extratropical cyclone, since that weather system is probably the best understood. Without going into details, the essential physical mechanism driving the extratropical cyclone is known as baroclinic instability. The phenomenon itself (the ETC) was first described qualitatively by the so-called "Bergen School" (Bjerknes, 1919; Bjerknes and Solberg, 1921, 1922) via the "Polar Front Model." This is summarized in Fig. 1.2, which shows the basic structure and evolution of an extratropical cyclone. A variety of explanations were put forward to explain the underlying process during the ensuing decades, but the lack of adequate upper-air data prevented any satisfactory explanation for nearly 30 years. Then, the insights of Rossby (1940) and Charney (1947) provided the long-sought answer in quantitative terms which have come to be known as baroclinic instability.

This instability theory can be fairly easily summarized without mathematics. We begin with the fact that the north-south variation in solar heating results in a north-south temperature gradient. With the observation that this gradient is not uniformly distributed, but is concentrated in mid-latitudes, forming the so-called polar front, physical reasoning can be used to show that over the front the westerly winds must increase with height (with the increase being proportional to the strength of the temperature gradient)[1]. This increasing westerly wind with height, or vertical shear, intensifies as the unequal heating continues. The extratropical cyclone forms as the primary process by which this strong gradient is alleviated. In essence, the unequal heating stores up potential energy and when enough is stored up to trigger baroclinic instability, the developing cyclone draws on this reservoir of potential energy to drive the circulation (thus producing kinetic energy). When the reservoir drops below some critical level, the system then begins to decay and the circulation slowly winds down. Along the way, the storm has moved warm air upward and northward, while cold air has travelled downward and southward. Therefore, the flow has acted to relieve the strong temperature gradients which initiated the system.

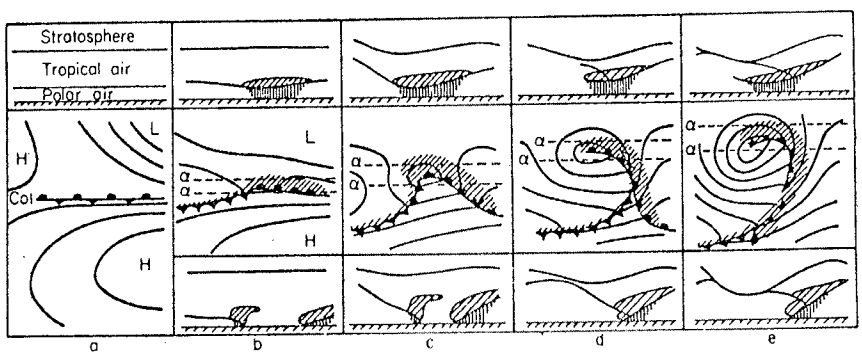


Fig. 1.2. Life cycle of extratropical cyclone (after J. Bjerknes, from Godske et al., 1957). In middle figures, thin lines are sea-level isobars. Top and bottom figures show schematic clouds, frontal surfaces and tropopause along lines α, a little north and south of ETC center. The times from stages a to c and from c to e are roughly one day in each case.

One of the observationally verifiable notions which has allowed treatment of baroclinic instability from a theoretical viewpoint is the basic validity of geostrophic balance on the scale of the extratropical cyclone. That is, the observed winds are pretty close to geostrophic, except perhaps near the surface. This observation has been incorporated in

the analysis of extratropical storms under the general heading of quasigeostrophic theory (see e.g., Holton, 1979, Chap. 6 and also II.B.1). Basically, the geostrophic wind is parallel to the isobars (or the contours, in pressure coordinates), with low pressure (heights) on its left, and with speed proportional to the magnitude of the pressure (or height) gradient (Fig. 1.3).

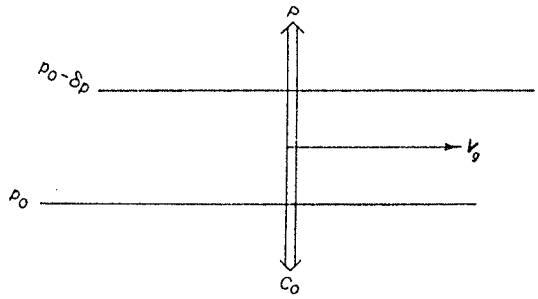


Fig. 1.3. The balance of forces for geostrophic equilibrium (after Holton, 1979). The pressure gradient force is denoted by P and the Coriolis force by C_0, while the resultant geostrophic wind is V_g.

However, one might easily be led to ask some potentially embarassing questions about this state of balance. For example, if the geostrophic wind is so good at approximating the true wind, how do pressure systems deepen (or fill)? If the wind speed happens to be non-geostrophic (i.e., ageostrophic) for some reason, how do the winds and/or pressures re-adjust to geostrophy? Since the geostrophic wind is not divergent[2], can we say then that vertical motion is unimportant for baroclinic instability?

We shall not explore the answers to all these questions in these notes, but once again refer the reader to the references. However, the subject of how the winds and pressure field come to adjust themselves to a state of near-geostrophic balance happens to be relevant to the issue of scale. The manner in which the adjustment occurs depends on the scale of the pressure system (Rossby, 1938). Specifically, on the small scale, the pressure field changes to fit the winds while on the large scale, the winds adjust to fit the pressure field. But how small is "small" and how large is "large"? It turns out that we can define a length scale called the Rossby radius of deformation, λ, which is related to the problem of geostrophic adjustment. Physically, the adjustment is accomplished by gravity waves which travel at relatively fast speeds. If we take this gravity wave speed and divide it by the Coriolis parameter[3] (the reciprocal of the Coriolis parameter defines a time scale appropriate to geostrophic balance), we obtain the Rossby radius of deformation. This can be interpreted as the influence radius of the gravity waves which accomplish the adjustment. For length scales much less than λ, the gravity waves have time to reach any point in the system and they act to adjust the pressure field. For length scales much greater than λ, gravity waves can not penetrate the

entire system and the winds have time to adjust to the pressure.

Just how large is λ ? It happens that λ is about 1500 km, which is a length of the same order as that of the ETC. Since these disturbances are neither much larger nor much smaller than λ , we can conclude that "synoptic-scale" systems adjust both their wind and their pressure fields to maintain a state of near-geostrophic balance. Such a conclusion should be readily apparent to those who deal with the weather operationally. The Rossby radius of deformation also provides a useful clue to the behavior of the "short wave" troughs in the atmosphere, and the smaller-scale features in the jet stream. Since these smaller features have lengths perhaps as small as 300 km, one expects their pressure fields to react to non-geostrophic winds rather than vice-versa. Again, operational experience supports this conclusion.

We have established our large-scale limit as the Rossby radius of deformation. In doing so, we have made a somewhat less arbitrary choice than is often made, since it is based on well-accepted theory and observational experience. That is, baroclinic instability (which is widely accepted as the dominant physical mechanism in extratropical cyclones) requires both wind and pressure perturbations to operate, limiting the scales of these weather systems to near the Rossby radius of deformation.

Can we motivate a definition similarly for what we call "mesoscale" - i.e., our lower limit of consideration in this section? The main issue in developing a physical-dynamical definition of mesoscale is whether or not there exists a dominant, scale-dependent instability which forces mesoscale systems. Emanuel (1980) has suggested the so-called "symmetric" instability for this purpose, but he also leaves open the possibility that other processes may exist and be physically significant. His basic definition of mesoscale is that on such a scale, both Coriolis accelerations and ageostrophic advection are important.[4] This approach seems entirely reasonable, and symmetric instabilities do, indeed, operate on such length scales (~100 km). Further, this scale definition turns out to lie at about the resolution limit of operational surface data. Hence, this is probably the best choice for our lower scale limit, even if the dominant physical process is not as clearly established as on the larger scale.

It does seem clear that on scales below 100 km, the Coriolis acceleration becomes more dynamically irrelevant,

while on scales much larger than 100 km, the ageostrophic contribution to advection becomes decreasingly significant. Quantitatively, this is accounted for by the Rossby Number (Ro) - the ratio of the actual to the Coriolis acceleration. Thus, Ro is small for length scales of 1000 km or more, and large for scales below 100 km. Around 100 km, Ro~1, which says that the Coriolis and actual accelerations are about the same.

Theory suggests that for these intermediate scales, a wide variety of instabilities are possible and the actually occurring combination of parameters may determine which process is most unstable in a given situation. This variety of theoretical instabilities is plausible when we realize that a much greater range of phenomena is seen to exist on the mesoscale than on larger scales. The ETC is by far the dominant form of weather system operating at scales near the Rossby radius (at mid-latitudes), whereas we shall see that a lot of fundamentally different phenomena occur in the mesoscale range.

Further, it is not clear on this scale what sort of dominant force balances exist, if any, analogous to geostrophic balance on the large scale. Hopefully, future research will provide some insight into mesoscale instabilities and allow a clearer picture to emerge of what "mesoscale" really implies about the dynamics of systems. At this time, it seems plausible to suggest that friction and latent heat are likely to have larger roles than they play in baroclinic instability. Since these two factors have proven difficult to treat in theoretical models, considerable time may elapse before we can treat mesoscale processes on the same level as we now deal with the ETC.

Finally, the density and frequency of upper air data may well prove to be the barrier to our mesoscale understanding that they once were on the large scale. It is difficult for meteorologists to attempt an explanation of phenomena they have not routinely observed, since the mathematics of atmospheric flow allow a bewildering variety of solutions. Only by careful comparison with observations can plausible theories be selected from the vast array of candidates. Since "mesoscale" observations are still not routinely available, only limited conclusions can be drawn from the limited mesoscale data.

CHAPTER I FOOTNOTES

[1] P. I-4: This physical reasoning is based on the concept of the thermal wind (see e.g., Holton, 1979, p. 68ff and also II.B.2), which is in turn an application of the geostrophic wind law, valid only for large-scale flow.

[2] P. I-5: This is not exactly true, as we shall see in IV.B.

[3] P. I-5: The Coriolis Parameter (often denoted by "f") can be thought of as the vorticity of the earth about the local vertical. Thus, at the north pole, where the local vertical is also the earth's rotation axis, f is simply the earth's vorticity (twice its rotation rate, or $1.4584 \times 10^{-4}\ s^{-1}$). Since the local vertical increases its departure from the earth's rotation axis as one moves away from the poles, the Coriolis parameter decreases with latitude, and vanishes at the Equator. The rate of decrease in f is slow at high latitudes (f is $1.0313 \times 10^{-4}\ s^{-1}$ at 45°N), but increases rapidly, reaching its maximum at the equator itself. Coriolis parameter changes signs upon crossing into the Southern Hemisphere so, for example, the Southern Hemisphere geostrophic wind blows with low pressure on its right.

[4] P. I-6: In the case of large-scale motions just described, the advection of atmospheric properties is dominated by the geostrophic contribution. In fact, this is a cornerstone of quasigeostrophic theory.

II. Upper-Air Data Analysis

A. General Remarks

It must be pointed out immediately that the network of upper air observations is entirely inadequate for any true mesoanalysis. With routine soundings over the U.S. only available every 12 h, at an average separation of about 400 km, no analysis can be considered mesoscale.

Nevertheless, this is where mesoanalysis should begin. It cannot be overemphasized that a forecast should start with a 4-dimensional mental picture of the atmosphere. Thus, some of the analyst's most important efforts should be directed toward developing this 4-dimensional understanding. With the development and application of sophisticated remote sensing tools (specifically, radar and satellite imagery), new understanding of many aspects of convection has been obtained rapidly. It should be completely obvious that analysis should not be done without examination of all the available data. The process of analysis is, in no small part, heavily dependent on the skill of the analyst at integrating a variety of data into a unified picture (i.e., a synthesis). Although these notes by themselves cannot provide the reader with all the necessary knowledge to interpret remote sensing data, some elements will be presented in those areas where such data can be crucial in the analysis process.

Remote sensing data can have a real impact on the upper-air analyses, in two related ways. First, the position and strength of upper air systems can be refined, based on the cloud and precipitation patterns. Second, and more importantly, information from the data-void areas (e.g., over the oceans) may have a real impact, either directly (e.g., a feature in the Gulf of Mexico which can move onshore later in the forecast period) or indirectly (e.g., a misanalyzed short wave trough which results in a faulty numerical prognosis [Hales, 1979a]). See Anderson et al. (1974) or Weldon (1979) for applications of satellite imagery to the synoptic scale analysis problem.

A 4-dimensional understanding is possible, even with limited time contraints, using centrally analyzed charts at the standard upper levels. As detailed by Maddox (1979b), these upper air and surface maps can and should be enhanced to emphasize features of importance to convective storm forecasting. At SELS, analysis of upper level charts is done by hand, as well. Although subjective analysis has numerous drawbacks from a theoretical and aesthetic

viewpoint, it is an excellent way of accomplishing several worthwhile goals. These include: (1) all of the data are subjected to examination, thus pinpointing erroneous observations, convection-contaminated soundings, and so forth; (2) the process of "drawing lines" forces an awareness of the significant upper-air features; and (3) an analysis of upper-air maps can be accomplished which is oriented toward mesoanalysis - i.e., the heavy smoothing necessary for large-scale modelling purposes can be avoided. Many texts exist to help guide the process of synoptic-scale analysis (e.g., Saucier, 1955; Petterssen, 1956a; Godske et al., 1957).

The forecast day generally begins with the morning (1200 GMT) soundings. The data at that time are relatively free of convective contamination. This is somewhat less true in the late spring and summer, when convection may continue through the night and on into the next day (note the discussion by Maddox, 1980b). Nevertheless, the morning analysis should allow the forecaster to develop a relatively clear picture of the synoptic-scale setting for the afternoon's and evening's developments.

B. Upper Air Chart Analysis

If the analyst has the option of contouring the constant pressure level charts, rather than simply enhancing the facsimile (or AFOS) products, the basic process is relatively straightforward. At 850 and 700 mb, the Severe Local Storms Forecast Unit of NSSFC (SELS) analyzes for height (30 m contour interval), temperature (2°C isotherm interval) and dewpoint temperature (2°C isodrosotherms, starting with 8°C at 850 mb and 0°C at 700 mb). At 500 mb, heights (60 m contours) temperatures (2°C isotherms) and 12-h height changes (30 m isallohypses) are analyzed. At 250 mb, isotachs (20 kt interval) and axes of maximum wind are depicted. Examples of SELS-type analyses shall be shown in III.V.

Within some limits, the development of this basic set of charts follows standard analysis practice (see Saucier, 1955, ch. 4). As described by Miller (1972), the analyst should avoid drawing closed isopleths whenever possible, even at the occasional expense of creating long, narrow "ribbons". There is good evidence that the atmosphere really does tend to create such features and the basic idea is to emphasize the source regions.

An important departure from synoptic scale practice is a heavy emphasis on 12-h changes in the observations. The

SELS routines which plot the upper-air data provide a 12-h change for all plotted variables, including the winds. Rather than emphasizing chart-to-chart continuity, the severe weather analyst needs to recognize the significance of the chart-to-chart changes. Of course, some effort should be made to develop time continuity, but the upper air data by themselves are too sparse in space and time to provide a clear picture of the often subtle features which move through the synoptic-scale patterns. Short wave troughs, wind maxima, vorticity "lobes" and small-scale temperature anomalies are frequently too small to be analyzed in detail unless 12-h height changes, backing/veering patterns of the wind, and thermodynamic changes are examined.

There are two complicating factors in evaluating the change fields: normal diurnal variations (e.g., Harris, 1959) and the contamination of the rawinsonde observations by convection. The analyst should know and recognize the expected diurnal changes (e.g., high 700 mb temperatures at 00 GMT over the mountains; roughly 20 m 12-h height rises at 12 GMT or falls at 00 GMT in mid-latitudes at 500 mb). While diurnal effects are at least conceptually easy to account for, convection can produce large changes that are less easy to adjust. Studies by Ninomiya (1971a,b), Maddox (1979a, 1980a), and others have shown that large thunderstorm complexes (up to 500,000 km^2, often lasting for 8 hr or more) can have a dramatic influence on even synoptic-scale rawinsonde networks. Since the effects of convection cannot be isolated, the correction of convectively contaminated data is basically not possible. Radar and satellite data should be examined routinely during analysis so that the analyst can exercise caution in interpreting the data within convective regions.

Most, if not all, of the effort spent by a severe weather analyst/forecaster in examining upper-air data is directed toward finding where upward vertical motion will occur in regions of moist, unstably stratified air (Beebe and Bates, 1955). This being the case, the real job of analysis should be directed toward this end, not merely drawing lines on the charts.

1. Vertical Motion

By way of introduction, one might ask the physical reason for a meteorologist's preoccupation with vertical motion. The production of "weather" requires condensation and the most common way the atmosphere produces condensation is adiabatic cooling by expansion. This

results from lowering the pressure. Since in horizontal motion parcels tend to travel parallel to isobars, no important change in pressure results. Local pressure falls do, of course, occur but their magnitude is so small in comparison to what is required to saturate parcels that their effect is not significant (but pressure falls are important in other ways, of course; see III.D). Since pressure surfaces are so closely packed in the vertical through the troposphere, a small vertical displacement can result in a large change in pressure. Naturally, this is reflected in the normal state of large-scale hydrostatic balance, where the relatively large vertical pressure gradient force is compensated for by gravitational acceleration. What small vertical accelerations occur are quite negligible in comparison, but still are our primary source of weather. For meteorologists the physical significance of vertical motion is ultimately the reduction of pressure following a parcel which results in condensation.

In examining upper-air data to locate "features", a basic problem is the diagnosis of regions of upward vertical motion. Upward motion on the scale of the upper air data is in the range of a few cm s^{-1}. This illustrates the essentially horizontal nature of large-scale flow, since the vertical component can be less than a tenth of one percent of the horizontal wind. However, since this upward motion is sustained for long periods, it can have dramatic effects. If one maintains a 5 cm s^{-1} upward motion for 24 hr, the net vertical lift is more than 4 km! Further, if the parcel started at a pressure of 1000 mb, that amount of lift reduces the pressure to about 600 mb. Since the surface of the earth and the tropopause act effectively as bounding surfaces for vertical motions, a region of upward vertical motion must have convergence at its base and divergence at its summit. This is a consequence of the law of mass continuity. Thus, divergence undergoes a change in sign with height, leading to the concept of the so-called level of nondivergence. Actually, this "level" is rarely at the same height from

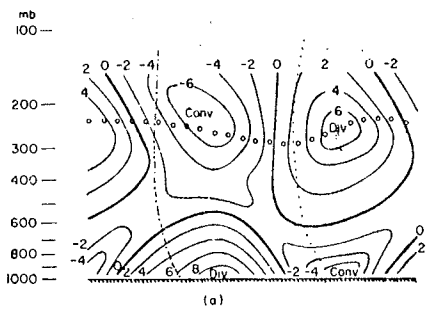


Fig. 2.1. Vertical cross section (after Fleagle, 1948) of horizontal divergence relative to trough and ridge lines (dotted and dash-dotted lines, respectively). Divergence contours in units of 10^{-6} s^{-1}.

place to place and time to time. Rather, it is typically a sloping surface (Fig. 2.1), as described by Charney (1947). Therefore, the axis of strongest vertical motions may be somewhat tilted away from the vertical. The interest in divergence is, therefore, an extension of the need to assess large-scale vertical motion.

A basic effort in analysis is to infer upper level divergence from such features as short-wave troughs, jet maxima, vorticity advection, and so forth (see McNulty, 1978; or Kloth and Davies-Jones, 1980 for discussions on these topics). Owing to several difficulties, we often must rely on such subtle approaches to diagnose divergence. One basic problem is that we have available only 12-hourly samples: in the morning when mesosystems may not be well developed, and again in the evening when convection is usually already underway. Organized regions of upper-level divergence are hard to follow as a result. Another, frequently mentioned problem is exemplified if we consider a 5 cm s^{-1} upward motion at a height of 5 km. This implies that the average low-level convergence in the layer from the surface to 5 km is 10^{-5} s^{-1}. This, in turn, suggests horizontal wind differences in the range of 1 m s^{-1} over a distance of 100 km. Small changes in the data (say 10% of the observed wind speed) can result in a large change (in the range of 100%) in the calculated divergence and, hence, the vertical velocity.

Given the small magnitude of synoptic-scale vertical motion and the modest changes in horizontal wind needed to produce it, the role of quasigeostrophic theory becomes more clear. For most purposes, and specifically for horizontal advection, the geostrophic flow is good enough. The divergence needed for vertical motion is not contained in the geostrophic wind, but the theory can be used to evaluate it. In effect, the vertical motion is the result of a secondary flow (much weaker) which is required to maintain a state of near-geostrophic (and hydrostatic) balance. This secondary circulation is a cornerstone of quasigeostrophic theory (and explains why the term is <u>quasi</u>geostrophic) and its validity is seen in its value for diagnosis of real weather systems.

Vorticity advection is widely accepted as an indirect means of locating large-scale upward motion. By vorticity advection, we mean a pattern of height contours and vorticity isopleths as shown in Fig. 2.2. For this indirect method to work, a variety of assumptions is necessary. The first two assumptions are that the actual winds are closely approximated by the geostrophic winds (which parallel the contours) and that the vorticity field

is derived from the height field (i.e., is essentially geostrophic) and so is moving slower than the winds. Under these restrictions, a parcel moves through the vorticity pattern, and finds its original vorticity different from that of its environment. Another assumption is that the basic process by which the parcel changes its vorticity is divergence (or convergence), so if a parcel is moving into regions of lower vorticity (as in a region of positive vorticity advection [PVA]) there must be a tendency for divergence to bring the parcel's vorticity down to that of its environment.

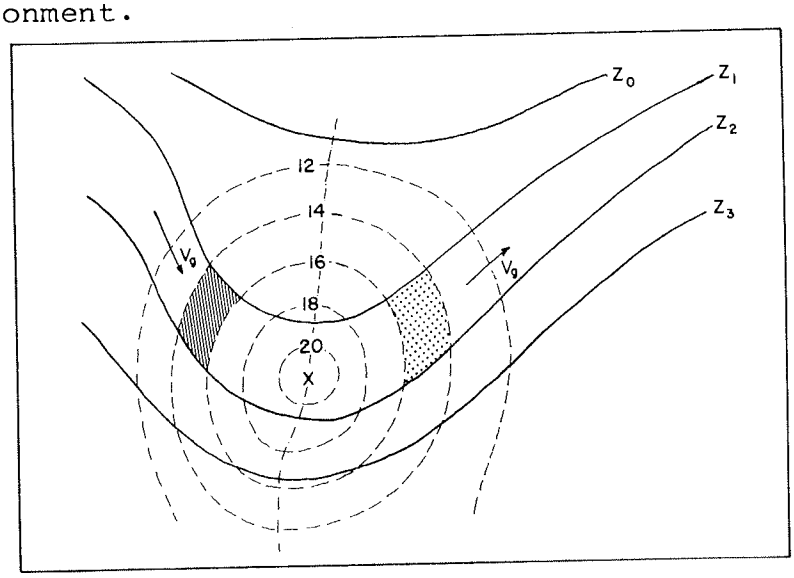


Fig. 2.2. Schematic showing vorticity advection by the geostrophic wind (V_g). Solid lines are height contours (z), dashed lines are contours of absolute vorticity (in units of 10^{-5} s^{-1}. Where the height and vorticity contours intersect, they form quadrilaterals (with curved sides). The strength of the advection is proportional to the number of such quadrilaterals per unit area. Where vorticity and height contours are parallel, no advection is occurring. The hatched quadrilateral is in a region of negative vorticity advection (NVA) by V_g, since V_g is pointing from lower to higher vorticity. The stippled quadrilateral is in a region of positive vorticity advection (PVA) by V_g.

Petterssen (1956a, p. 299ff) presents the PVA arguments as follows: at lower levels, vorticity advection is weak since the flow is very nearly parallel to the vorticity isopleths. Therefore, at those levels, vorticity changes are dominated by divergence effects. Regions of increasing vorticity must be convergent (and vice versa) at low levels. At upper levels, vorticity advection is large but local changes are small in comparison. Air passes through the vorticity pattern since wind speeds are high, so the arguments (above) apply which suggest that PVA implies

divergence. At middle levels (500 mb), divergence is small and vorticity is very nearly conserved - local changes in vorticity are dominated by advection. Historically, this is why 500 mb was chosen for early numerical forecasting models (the "Barotropic" model). Vorticity changes implied by PVA at 500 mb produce convergence below and divergence above that level - hence, vertical motion. Panofsky (1964, p.114ff) also gives an excellent description of how to infer vertical motion from vorticity concepts.

This simple physical picture is subject to many restrictions because so many assumptions are involved. Although the winds are not usually too far from geostrophic, it is often those cases of large ageostrophic departures which produce significant weather [recall the geostrophic wind is essentially non-divergent!]. Also, occasionally, the vorticity pattern may move faster than the winds, reversing the convergence/divergence patterns associated with vorticity advection. Finally, it is not at all clear that 500 mb level parcels conserve their vorticity, that divergence is the only mechanism by which parcels change their vorticity, and that 500 mb is always near the level of nondivergence.

Nevertheless, in spite of all these potential problems, PVA patterns often prove useful. The careful analyst should be aware of those situations where PVA is less likely to tell the whole story. An excellent discussion of large scale vertical motion can be found in Holton (1979, p. 136ff.). In this discussion the role of PVA in producing vertical motion is clarified. Specifically, there are two sources for vertical motion in quasigeostrophic systems. Rising motion is proportional to (a) the rate of increase with height of PVA and (b) the strength of warm thermal advection.

Note that PVA must increase with height for upward vertical motion to result. This is an essential consequence of the law of mass continuity we have described and is consistent with the physical picture presented above. If the divergence (related directly to PVA) does not increase with height, then the air is not likely to be rising, even if PVA exists at the standard 500 mb level. Hales (1979b) has recently emphasized this important point.

The contribution of warm advection to upward motion is often neglected. The physical significance of this effect can be described in a variety of ways. Consider the well-known relationship that the thickness of a layer (usually bounded by pressure surfaces) is proportional to

the mean temperature in that layer. Thus, warm advection is essentially related to thickness advection. A common situation wherein warm advection plays a role is with a warm front. The southerly flow, nearly perpendicular to the thickness contours, produces strong warm (thickness) advection, which tends to increase the thickness at a point. The vertical motion (upward) acts to cool the column by lifting and, therefore, tends to compensate for the warming. Upward motion induced by warm advection is often erroneously attributed to "overrunning".

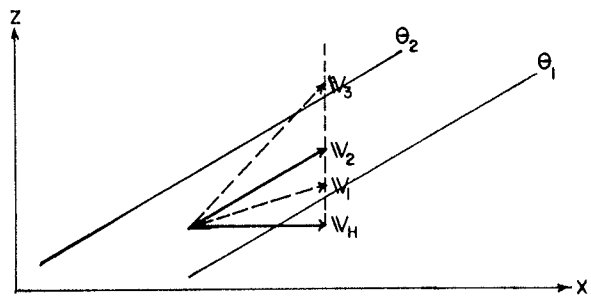


Fig. 2.3. Schematic illustration of how the 3-dimensional wind acts to displace isentropic (θ=constant) surfaces. The effect of the horizontal wind component ($\mathbb{V}_H$) is to push the θ-surfaces from left to right. Three different vertical components are illustrated; $\mathbb{V}_1$ is the typical example, which makes the θ-surfaces rise, displacing them from right to left at any given level, such that the net displacement is less than that indicated by the $\mathbb{V}_H$ contribution, but still from left to right. In the second case, $\mathbb{V}_2$ is such that the 3-dimensional flow is parallel to the θ-surface, yielding no net displacement. For the third case, which is rare, the vertical component of $\mathbb{V}_3$ is so large that the net displacement by vertical motion is larger than the $\mathbb{V}_H$ contribution, giving a net movement from right to left.

Most of the confusion about "overrunning" and the effects of warm advection result from taking a 2-dimensional, rather than a 3-dimensional view. Fig. 2.3 shows a cross section through a frontal zone, with potential temperature (θ) surfaces (isentropes). The actual winds are acting to push the θ-surfaces from left to right, by advection. However, the vertical motion also acts to lift those surfaces, which displaces them opposite to the contribution by advection. If the 3-dimensional wind happens to be exactly parallel to the θ-surfaces, there is no horizontal movement, despite a horizontal wind component across the surfaces. In general, the flow is not exactly isentropic, usually giving a net horizontal displacment less than the normal component of the horizontal wind. This also

explains why warm fronts tend to move more slowly than cold fronts. It happens that analysis on isentropic surfaces is a good way to see this on a 2-dimensional chart, subject to the limitation that the actual flow may not be exactly along isentropes. Note that in some unusual cases, the contribution by vertical motion can exceed that by advection, so the front could "back up", into the horizontal flow!

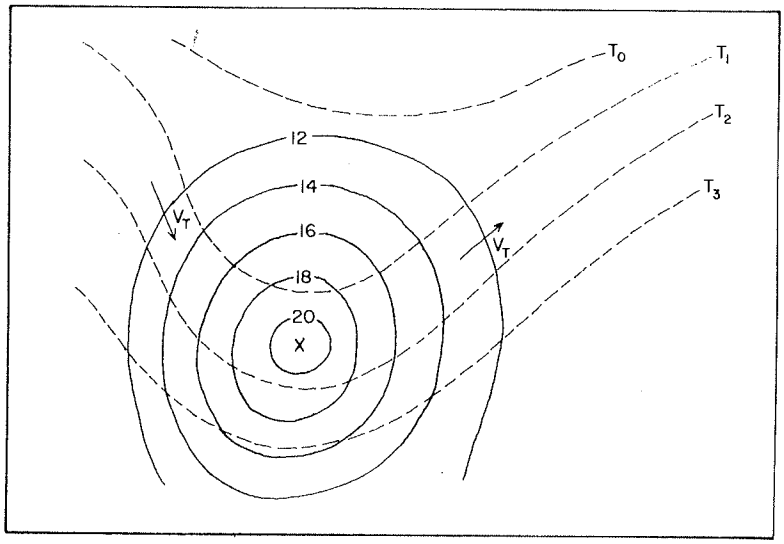


Fig. 2.4. Vorticity advection by the thermal wind (V_T). Thickness contours (T) are dashed lines, while solid lines are contours of absolute vorticity (as in Fig. 2.2). Note that thickness contours and height contours usually do not coincide, so that V_T differs from V_g.

Recently, Trenberth (1978) and Hoskins et al. (1978) have pointed out that the PVA and thickness advection effects have a tendency to cancel each other. This can also be seen in the discussion by Holton (1979, p. 139). Trenberth has proposed a solution to this dilemma by using the advection of vorticity by the thermal wind. Those familiar with the pioneering work of Sutcliffe (e.g., Sutcliffe, 1947 or Sutcliffe and Forsdyke, 1950) should recognize this approach. This requires doing the same thing that is currently done with PVA, but using thickness contours (to infer the thermal wind) rather than height contours. Sangster (personal communication) has verified the validity and value of this approach on a day-to-day basis. Sangster's estimates of 850 and 700 mb vertical motion are derived by using the vorticity and isotherms at each level. The isotherms at each level ought to be fairly good approximations to thickness contours (for a layer containing that level), so this is quite similar to vorticity advection using the thermal wind.

Since this revised method for locating areas of upward motion includes both the PVA and thermal advection terms, it has clear advantages. With AFOS, overlaying the thickness and vorticity fields is relatively simple.

There are other ways to estimate the vertical motion field, including the model output fields, which show forecast vertical motion directly. Since the model-generated vertical motion patterns are not perfect, it is in the analyst/forecaster's interest to have as many different estimates of where upward motion is (or is going to be) occurring as possible. This includes empirical rules as well as more objective methods, since no single approach applies equally well under all conditions. Most severe convection depends on larger-scale forcing to develop (and/or maintain) its severity. It is worth noting that this supportive large- scale upward motion may not always be obvious from indications in mid-troposphere (~500 mb). There is evidence (e.g., Hales, 1979b) that during the warm season, the "upper support" may only be detectable above 500 mb. Also, the forcing can be confined to levels below 500 mb (Doswell, 1977; Maddox and Doswell, 1982), as well. However, it should be recalled, that upper divergence and lower convergence are most frequently related, as we have discussed. This is an essential element in the work of Uccellini and Johnson (1979), in which the coupling of upper and lower jet streaks is stressed, and which is discussed further in III.F.3.

2. Production of Unstable Thermodynamic Stratification

Vertical motion, by itself, obviously is insufficient to develop severe thunderstorms or heavy convective rain. In fact, large-scale vertical motion produces large-scale regions of condensation. It can be argued that the major role played by large-scale upward motion is to prepare the environment for convection. One basic property of convection is that it requires an unstable thermodynamic stratification.[2] Therefore, a substantial effort in the interpretation of upper-air charts is directed also toward questions of the instability of the "air mass". Note that it is unusual for severe storms to occur in a true air mass region, i.e., one with horizontally uniform properties. Thus, it is somewhat misleading to speak of the "unstable air mass" in which severe convection develops. This is especially the case since the vertical structure associated with severe storms (to be discussed later) usually reveals different source regions for the air at different levels.

The means by which the classic severe storm sounding develops is often the result of a process of differential advection. This process, described by McNulty (1980), Whitney and Miller (1956), and Appleby (1954) among others, is simply the result of vertical differences in the horizontal advection of atmospheric properties (see Fig. 2.5). If differential advection acts to warm the lower layers relative to those above (or, equivalently, to cool the upper layers relative to those below), the result is a net decrease in the stability of the air column. Typical values in severe weather soundings suggest that differential advection can increase the lapse rate of a sounding by as much as $1^{o}C\ km^{-1}$ every 3 hr (recall the dry adiabatic lapse rate is about $10^{o}C\ km^{-1}$). As evidence of the importance of instability changes, Newton (1980) has shown that an average parcel buoyancy increase of $1^{o}C$ over the depth of the troposphere can increase the cloud maximum vertical velocity by $7\text{-}12\ m\ s^{-1}$.

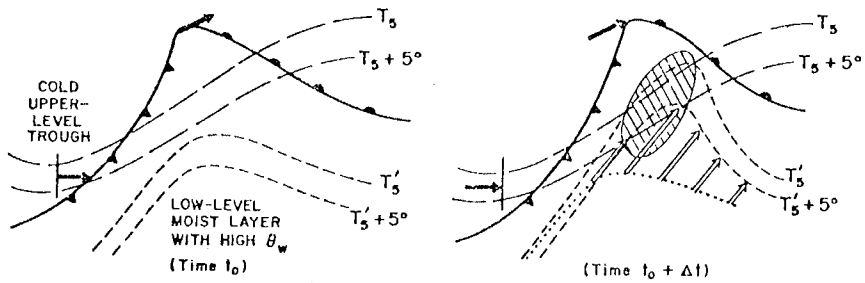


Fig. 2.5. Schematic illustration of differential advection (after Newton, 1980). Frontal symbols are conventional. Long-dashed lines are 500 mb isotherms while short-dashed lines are isotherms of low-level parcels lifted to 500 mb (proportional to θ_w). Note that the eastward progression of the 500 mb thermal trough and the northward progression of high θ_w air at low levels creates a condition of instability at time $t_o + \Delta t$ in the hatched region. That is, low-level parcels in the hatched region, when lifted to 500 mb, are warmer than their environment.

Further, if differential advection results in a net moistening of the lower layers, and/or a net drying of the middle and upper troposphere, the convective potential is also enhanced. In fact, an increase in moisture content by $1g\ kg^{-1}$ is about equivalent to increasing the temperature by $3.5^{o}C$, if all the latent heat can be released. McNulty (1980) has combined the influences of temperature and moisture by considering the differential advection of wet-bulb potential temperature (θ_w)[3], since convective instability is defined to exist when wet-bulb potential temperature decreases with height. McNulty's study was not directed beyond the short-range correlation of differential advection

with observed severe storms, so no clear-cut results concerning the relationship were found. Nevertheless, over a period of days, differential advection must play a substantial role in creating areas of unstable stratification. In most cases, the modification of stratification necessarily involves the process of differential advection. This does not imply that, once a basically unstable region has developed, differential advection is an ongoing, important process. McNulty's conclusions support this view, since he suggested that during the spring, differential advection is not effective at separating non-severe from severe storms, while it is more valuable at separating convective from non-convective regions. Because instability is confined to relatively small regions during the spring, the additional destabilization from concurrent differential advection was not significant. In summer, the opposite conclusion was drawn - i.e., differential advection is valuable for delineating areas of severe weather, but not effective for locating convective regions. As McNulty states (1980, p. 288), in summer, "Further destabilization is unnecessary for convection and appears to contribute only to severe convection development."

When discussing differential advection, it is perhaps appropriate to digress briefly and examine the concept of the thermal wind. If one examines the upper air charts, it is quite clear that the height contour pattern (and hence, the geostrophic wind) generally varies with height at any given location. The difference between the patterns at any two levels is simply the thickness between the pressure surfaces. A relationship known as the Hypsometric Law can be stated as follows: the thickness between any two pressure surfaces is proportional to the mean virtual temperature[4] in that layer (see Table 1). Thus, the thickness contours can be regarded as isotherms, as we have already mentioned.

Since the contour patterns change with height, so then does the geostrophic wind. By definition, the change in the geostrophic wind with height is the thermal wind.[5] Figure 2.6 shows how the thermal wind can be derived from the geostrohic winds at two pressure levels. Observe that we have two quantities which are related to the change of contour patterns with height: the thickness and the thermal wind. It is logical to assume that these quantities are related in some way to each other as well. This is, in fact, the case. Specifically, the thermal wind blows parallel to the thickness contours (i.e., to the layer average isotherms), with speed proportional to the thickness gradient, and with low thickness (temperature) to its left. This is totally analogous to the geostrophic wind's relationship to height

contours (recall Fig. 2.4).

TABLE 1

Pressure Ratio		Examples	Factor
P_{bot}/P_{top} =	2.0	(1000/500, 500/250, etc)	20.302
	1.70	(850/500)	15.542
	1.6667	(500/300, 250/150, etc)	14.962
	1.50	(300/200, 150/100, etc)	11.876
	1.4286	(1000/700)	10.447
	1.40	(700/500)	9.855
	1.3333	(400/300, 200/150, etc)	8.426
	1.25	(500/400, 250/200, etc)	6.536
	1.2143	(850/700)	5.687
	1.20	(300/200)	5.340
	1.1765	(1000/850)	4.760

Table 1. Factors, which when multiplied by the mean virtual temperature ($\overline{T}_v$) in a layer for which the bounding pressures have the given ratio, yields the thickness (in m) of that layer. Thus, for example if $\overline{T}_v$ = 0°C = 273.16°K in the 850-700 mb layer, then the thickness ΔZ is simply 5.687 x 273.16 = 1553.5 m.

So how does all of this apply to the subject of differential advection? Examine Fig. 2.6 and consider the wind's relationship to the thickness contours. If these contours are given their interpretation as isotherms, then it can be seen that there is a component of the winds in the layer which is blowing across the isotherms. This component, curiously enough, is the same at either level! That is, the normal component (to the isotherms) can be determined from the geostrophic wind at either level. The implication is that a change in geostrophic wind direction with height is always associated with thermal advection. As shown in Fig. 2.6, when the geostrophic wind backs with height (turns counterclockwise), cold advection is implied, while veering of the geostrophic wind with height indicates warm advection. Perhaps now the reason for this digression is clear. Thermal advection can actually be diagnosed simply by examining the change in the contour pattern with height, even if isotherms are not available. In fact, subject to the limitation that the real wind may differ markedly from geostrophic (especially at low levels), one can infer temperature advection merely by knowing the profile of winds aloft, over a single station! See also Oliver and Oliver (1945) for more details.

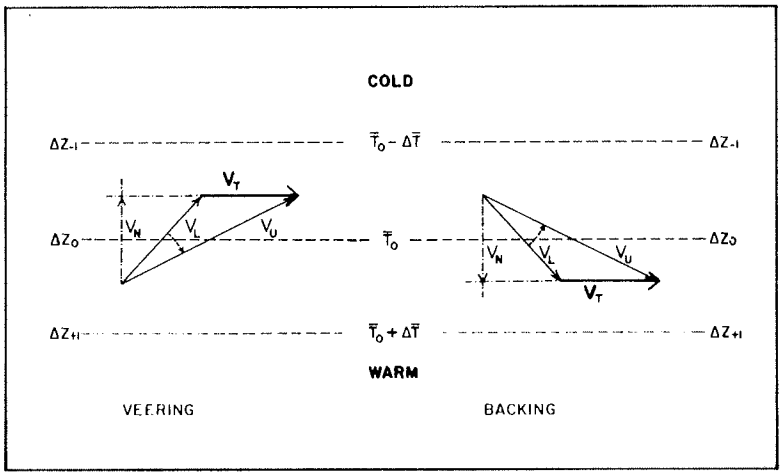


Fig. 2.6. Schematic illustration showing temperature advection as implied by the change in geostrophic wind with height. Geostrophic winds at the lower and upper levels are denoted by V_L and V_U, respectively, while the resultant thermal wind is given by V_T. Implied thickness contours are shown by dashed lines. The component of either V_L or V_U normal to the thickness lines is V_N. In the case where the geostrophic wind veers with height (turns clockwise in the Northern Hemisphere), warm advection is implied. Conversely, winds backing with height imply cold advection.

Given the geostrophic thermal advection contribution through several layers (or at several levels), one might be tempted to conclude that one could diagnose the differential thermal advection. After all, advection at any level is dominated by the geostrophic contribution. Unfortunately, this does not work. One cannot infer destabilization when the 850 mb geostrophic thermal advective change is positive and the corresponding 500 mb term is negative. In order to see why this is so, consider how the geostrophic advection changes with height. Since the geostrophic wind change is simply the thermal wind, which is parallel to the layer mean temperatures, the differential advection by the geostrophic wind must essentially vanish. Therefore, differential advection must be accomplished by the ageostrophic wind. While the ageostrophic wind may not be the largest part of the wind itself, it has two very important roles - it supplies the significant divergent part of the wind field and it provides the means to change the stratification via differential advection. An excellent discussion of how changes in stability occur can be found in Panofsky's (1964, p. 105ff) textbook.

As a final observation on differential advection, it has often been suggested that cold air advection at, say, 500 mb is an important contributor to severe weather potential. Observations do not support this on a day-to-day basis. While case studies certainly exist (e.g., Barnes, 1978) which show that cooling aloft as a result of cold advection did play a role, it is more frequently found that the environmental soundings show only weak thermal advection at 500 mb (either warm or cold) in the threat area. This is especially true during the late spring and summer (Hales, 1982).

A major contributor to the development of instability is the large-scale vertical motion itself. Regions undergoing large-scale lifting must necessarily approach an adiabatic lapse rate. The demonstration of this is available in any textbook (e.g., Hess, 1959, p. 102). By means of lifting, even an initially stable environment can become favorable for convection.

Since the classic pre-severe storm sounding (discussed in Miller, 1972) often has an inversion capping the moist layer, the lifting process may be essential for development of storms even when the atmosphere is already convectively unstable. For the typical storm environmental sounding, about 6 h of synoptic-scale lift (at~5 cm s^{-1}) is capable of eliminating the cap (i.e., about 1 km of net vertical lift). Note that the negative area for the sounding associated with the cap can be interpreted in an interesting way. This area happens to be proportional to the square of vertical motion (Petterssen, 1956b, p. 136)! That is, for negative areas, an upward vertical motion equal to the square root of twice the area is needed to cancel that amount of negative buoyancy.

When the capping inversion is too strong to be broken by the available sources of lift, no convection may occur even under conditions of extreme instability above the inversion. Thus, a coupling between "dynamics" and "thermodynamics" frequently must be present for severe storms. The capping inversion acts to enhance severe potential by confining moisture to low levels (Williams, 1960; Carlson and Ludlam, 1968) until it can be released. Although daytime heating from below may sometimes be sufficient to eliminate the inversion, the unmistakable relationship between severe thunderstorms and some source of upward motion (fronts, short-wave troughs, etc.) suggests that in most cases, the inversion is eliminated by lifting.

It should be pointed out that layer lifting in the traditional sense described by Hess occurs only for layers of small thickness. That is, the process Hess describes involves lifting the top and bottom of a layer by an equal amount. While this certainly has the effect described, one should remember that vertical velocity normally has its largest magnitude in middle levels (say, around 500 mb). Thus, layers of significant thickness will undergo stretching (or compression) which amplifies any changes in stability. Further, it is worth emphasizing that the more stable the layer is to begin with, the greater is the change in its stability as a result of lifting and stretching. In effect, the large-scale lifting process tends to drive lapse rates toward the dry adiabatic value. A layer which is already stratified nearly dry adiabatically will not undergo much change, whereas a very stable layer is altered rapidly by the lifting and stretching mechanisms.

Clearly, the source of vertical motion can be on different scales in different situations. A case like that of April 3-4, 1974 (Hoxit and Chappell, 1975) may be driven by large-scale lifting process. Doswell (1977) has shown that at times subsynoptic scale lifting may provide the means for breaking the inversion. Beebe (1958) has presented serial soundings where the inversion clearly rises and the moist layer deepens in a mesoscale area. As the scale of a translating vertical motion source decreases, the required average upward speed must increase, since it has correspondingly less time to act. That is, the time scale generally decreases with size scale. Mesoscale systems can develop vertical motions in the range of several m s^{-1}, but their life cycles can be completed in 6 h. Naturally, such detail can be unavailable to the operational forecaster, but it is clear that the existing instability (say, at 1200 GMT) in a region may not reflect accurately what the sounding will look like at the time of convection. The Lifted Index[6] (Galway, 1956) represents an adjustment of the sounding's stability parameter to account for diurnal heating. The analyst/forecaster needs to provide further adjustments based on the upper level charts, using the concept of differential advection and the possible effect of vertical motion.

3. Some Kinematic Considerations

As discussed before, many empirical rules for interpretation of upper level winds are indirect efforts to diagnose and forecast vertical motion. McNulty (1978) and Kloth and Davies-Jones (1980) have evaluated several of these ideas, as related to jet maxima. Hales (1979b) has

considered the use of anticyclonic (horizontal) shear in this context. It is pretty clear that mesoscale features exist aloft, even if conventional rawinsonde data are generally insufficient to reveal them. This insufficiency is related to the data density, to errors in the data (which tend to increase with height), to rounding winds to 5° direction intervals, and to the analyst's bias toward recognition of wind direction changes more readily than actual vector wind changes. The smaller scale details of the wind field can be inferred to some extent from the satellite images, especially when animated loops are available. The basic principle involved is that where there is cloud, there is upward vertical motion, and where there is vertical motion there is some "feature" which is forcing it (see Doswell, 1982a).

a.

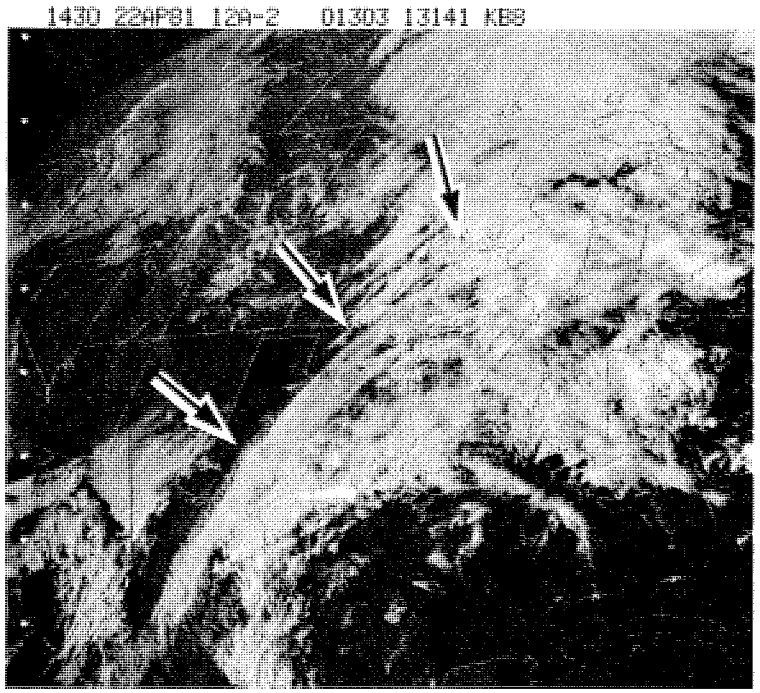
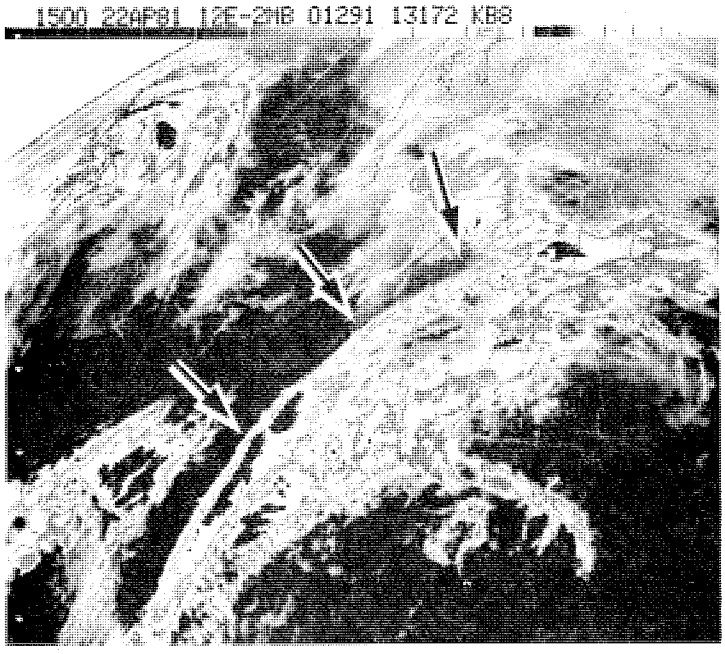


Fig. 2.7. Visible (a) and enhanced infrared (b, next page) satellite images showing anticyclonically curved band of cirrus across Texas, Oklahoma, southeastern Kansas and Missouri. Such bands are associated with upper level jet streams, with the jet axis from 1° to 5° poleward of the sharp cloud edge.

b.



Perhaps the most successful application of this principle is the location of the jet stream axis (Whitney et al, 1966; Whitney, 1977). However, the often very sharp cloud edge near the jet stream axis (e.g., Fig. 2.7) may not be the result of the deep vertical motions associated with jet stream secondary circulations (such as those described by Cahir, 1971). Rather, the "edge mechanism" appears to be an interface between shallow vertical circulations, basically confined to cirrus levels (Weldon, 1975). Details of this mechanism remain unclear.

Unfortunately, conventional data do not always relate well to cloud masses observed in satellite imagery. An interesting phenomenon which reveals the type of problems inherent in satellite interpretation is the large mesoscale convective complex (MCC) described by Maddox (1980b) (e.g., Fig. 2.8). As the MCC grows to maturity it has an increasingly obvious influence on the rawinsonde-sensed observations. The development of a diverted flow around the northern side of an MCC creates the illusion of a "short-wave trough" or "vort max" upstream, which may have no previous or subsequent history. It is an "effect" rather than a "cause", since it has a convective origin. In order to discriminate valid mesoscale features in the larger scale fields, the satellite imagery should, if possible, be supplemented with corroborative conventional data.

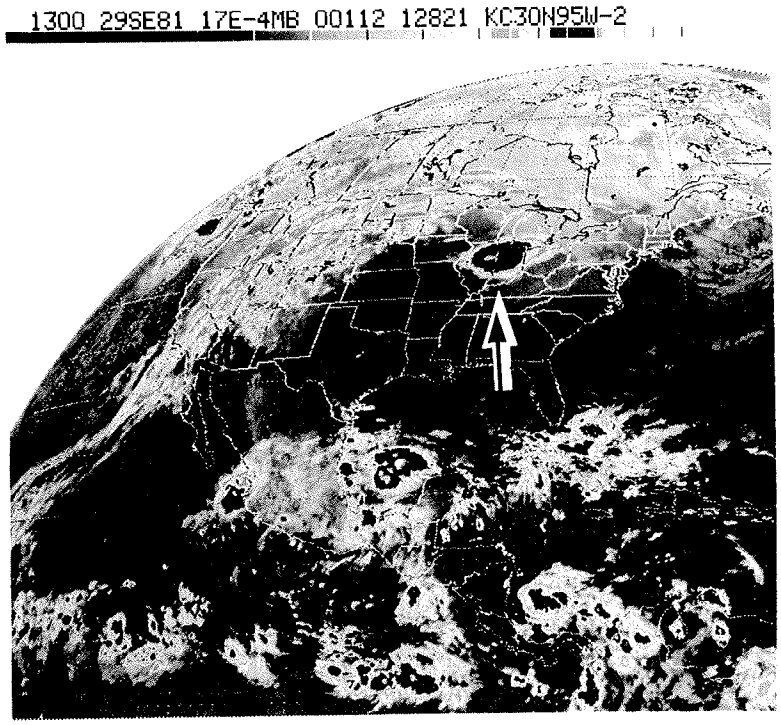


Fig. 2.8. Enhanced infrared satellite image revealing Mesoscale Convective Complex (MCC) over Illinois and Indiana.

C. Sounding Analysis and Interpretation

1. General Remarks

Part of the early morning upper-air analysis should include an examination of plotted soundings. This subject has also suffered from declining interest, along with other aspects of synoptic meteorology. It seems obvious that considerable useful information is available in the soundings. An abundance of literature (Showalter, 1953; Fawbush and Miller, 1954b; Galway, 1956; House, 1958; Prosser and Foster, 1966; Miller, 1972; Doswell and Lemon, 1979) exists which stresses the detailed vertical structure, both thermodynamic and kinematic, of the environment in which convection develops. An automated sounding analysis, such as that produced at SELS (Doswell et al., 1982), can help to decide which soundings to examine. However, such parameters as moisture depth, inversion strength, and wind directional variation are difficult to automate and can be helpful in developing a clear picture of the synoptic situation. Nothing can or should replace an examination of

the individual soundings. Such an examination can also help to evaluate and correct any erroneous data that may have crept into the constant level analyses.

Newton (1980) has presented the three types of soundings associated with thunderstorms (Fig. 2.9). Newton's Type A corresponds to Miller's (1972) Type IV tornado air mass, which is generally characteristic of High Plains severe weather situations. Newton's Type B is Miller's Type I tornado air mass, which is the classical "loaded gun" sounding of the Great Plains. Finally, Newton's Type C corresponds to Miller's Type II tornado air mass, typically identified with the Gulf Coastal regions of the southeastern United States. Newton does not explicitly describe Miller's Type III sounding and its similarity to the Type II profile (except for lower temperatures) suggests that it is a subset of the Type II (or Newton's Type C) situation.

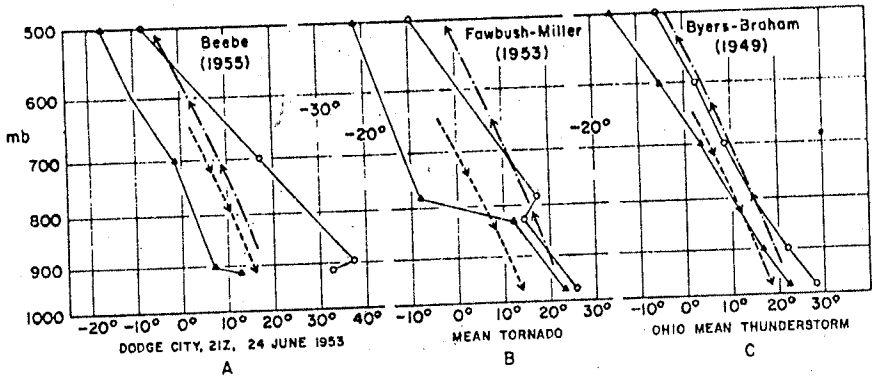


Fig. 2.9. Schematic of three distinctive soundings associated with severe convection in the U.S. (after Newton, 1980). Type A is an "inverted V" sounding typical of the High Plains or desert severe storms. Type B is the classical "loaded gun" sounding characteristic of Great Plains or central U.S. severe weather outbreaks. Type C is common over Gulf coastal states and in summer east of the Mississippi River. Dash-dotted lines correspond to moist adiabats associated with low-levels and dashed lines to moist adiabats in lower middle levels (600-700 mb).

2. Sounding Thermodynamics

First consideration of sounding analysis is an assessment, usually via a single parameter, of the stability of the thermodynamic stratification. This might be the Showalter Stability Index (Showalter, 1953), the Lifted Index (Galway, 1956), or the Totals Indices (Miller, 1972). These parameters key on the amount of buoyancy available to a lifted parcel at 500 mb, and have been in use for a considerable time. The SWEAT index developed by Miller (1972)

attempts to incorporate some kinematic properties, specifically the shear between 850 and 500 mb. The need for and value of these parameters are well-known and are straightforward.

There are other factors which can be evaluated from the soundings, some of which are not so easily automated. One important example is the depth of the moist layer. While some soundings typify the classical "loaded gun" severe weather sounding (Fawbush and Miller, 1954b) in that they have a well-defined, inversion-capped moist layer, surmounted by a substantially drier layer with a steep lapse rate, this is not always the case. The depth of the moisture has a large impact on the subsequent events. If the moisture is too shallow (say, less than 50 mb deep) there may be insufficient water vapor to support severe convection. If the moist layer is exceptionally deep (say, 200 mb or more), the likelihood of non-severe heavy rainstorms is greater. Further, as described in Schaefer (1974a), moist layer depth has a dramatic influence on dryline motion (see III.B.5).

The occasional occurrence of a very deep layer of essentially saturated conditions to, say, above 500 mb can result from convection contamination and, hence, be unrepresentative. However, it also can indicate some severe potential, especially in the southern part of the United States (Newton's type C). In many such cases, the occurrence of dry air aloft upstream from the threat area is common (Miller, 1972). This dry air typically has arisen from subsidence (but may have other origins - e.g., Carlson and Ludlam, 1968) and thus is also relatively warm, so a dry intrusion is frequently also indicative of warm advection (see III.E). The complete absence of dry air generally implies an increase in heavy rain potential and a corresponding decrease in the likelihood of severe thunderstorms.

However, moisture aloft in the absence of low-level moisture does not preclude severe weather, since high-based severe storms in such a situation are not uncommon, especially in the High Plains region of the United States (Newton's type A). A simple argument can show that such storms have a high potential for strong surface wind gusts. Further, when a shallow, surface-based moist layer is found in such soundings, tornadoes can result from High Plains thunderstorms (Doswell, 1980, Mahrt, 1977) since a few storms may be able to tap this low-level moisture by developing updrafts with surface roots.

The reader should have realized by this time that the existence of dry air, generally in the mid-troposphere, is an important factor in much severe convection. This has long been recognized (Ludlam, 1963). It appears that the enhancement of the downdraft potential, created by evaporation of cloud and precipitation into dry environmental air, plays a key role in developing the storm structures associated with severe weather (Lemon and Doswell, 1979).

Although it is not easily evaluated from a simple plotted sounding, the vertical profile of wet-bulb potential temperature (θ_w) is worth some examination. Since θ_w incorporates both temperature and moisture, its vertical distribution provides key clues about convective instability. In fact, by definition, if θ_w decreases with height in a layer, that layer is convectively unstable. As noted previously, the "loaded gun" sounding is the archetypical example of convective instability, since its moisture and temperature profiles combine to produce a minimum in θ_w in middle levels.

It has been argued that the difference between the θ_w minimum at mid-levels and the θ_w maximum at low levels (often at the surface) represents the total energy available to a severe storm (Darkow, 1968; Morgan and Beebe, 1971). This concept has been tested by Doswell and Lemon (1979). They found that, for a sample of severe thunderstorm environmental soundings before and then near severe storm occurrence, a parameter based on this difference did not seem too effective at delineating the region of most severe convection. However, they note that during the time from the sounding well before the storm to the sounding closest to storm occurrence, the minimum θ_w value actually increases slightly (about 1.5°C) and the height of the minimum rises (by about 80 mb). This can be interpreted as a reflection of the action of upward vertical motion. That is, the moist layer deepens and rises during the period before storms (Beebe, 1958).

Another factor that should be evaluated from selected soundings is the so-called negative area in the lower part of the parcel's ascent profile. If the parcel is negatively buoyant, energy must be supplied to lift the parcel through those layers. As suggested earlier, negative area can act to enhance severe potential by capping the release of energy until the optimum time (usually near the time of maximum surface heating). The Lifted Index can account for the contribution of surface heating to "cap" erosion by using a forecast maximum surface temperature. When substantial negative area remains after accounting for diurnal heating

(if applicable - at 0000 GMT, surface cooling will occur), the forecaster/analyst should try to determine whether there is a source of sufficient lift (e.g., a source of low-level convergence or some feature supplying upward motion) to eliminate the cap (recall II.B.2, above).

Also valuable in operational study of soundings is the determination of the equilibrium level for the rising air parcels. The equilibrium level (EL) is where the rising, buoyant parcel re-crosses the environmental sounding curve. It is this level, rather than the tropopause, that is physically significant. Anvil cloud material tends to accumulate here, rather than at the tropopause, since it is where rising parcels are (naturally) in equilibrium with their environment. Penetrations of the EL are indicative of strong updrafts, and the EL can be well below, near, or well above the tropopause. Naturally, depending on the characteristics of the tropopause, a storm which reaches above the tropopause is usually significant. However, when the EL is far below the tropopause, storms with tops which remain below the tropopause can still be severe (Burgess and Davies-Jones, 1979). Similarly, a storm which penetrates the tropopause may still be below the EL.

TABLE 2

	Less than 65 kt Preceding	65 kt or Greater Preceding	65 kt or Greater Next
All:	30.5 ± 30.0 (167)	41.3 ± 29.6 (45)	48.1 ± 30.9 (43)
Excl. 0's:	-5.7 ± 6.1 (62)	12.9 ± 8.8 (22)	13.2 ± 10.5 (19)
	105	23	24

$\frac{105}{167}$ = 62.9% observed gusts $50 \leq V < 65$ kts with 0 calculated (Preceding)

$\frac{23}{45}$ = 51.1% observed gusts $V \geq 65$ kt with 0 calculated (Preceding)

$\frac{24}{43}$ = 55.8% observed gusts $V \geq 65$ kt with 0 calculated (Next)

Table 2. Errors in predicted gust speeds: 01 Apr - 30 Jun 1978, for reported gusts 50 kt or greater in the two categories shown. Values are the mean difference (in knots) of (observed-predicted) ± the associated standard deviation. The number of cases is in parentheses. Values under "Preceding" refer to predictions from the sounding preceding the report by more than 6 hr; whereas "Next" refers to predictions from the sounding immediately following. Values are given for "All" predictions and also excluding cases when the prediction is "no gusts".

Positive area should also be evaluated. If a small calculator is available, the positive area can be determined from the hypsometric equation as the difference in thickness

between the observed heights and the heights using the parcel ascent curve (between the LFC and the EL). As discussed earlier, this can be used to determine the parcel theory vertical motion associated with the amount of positive area. Such a vertical motion speed is generally an overestimate (see II.II.A), but is representative of peak updraft speeds in the most severe storms.

In the past, some attempts have been made to forecast the maximum gust potential and/or the maximum hail size possible with a given sounding (Foster and Bates, 1956; Foster, 1958; Fawbush and Miller, 1954a; and Fawbush and Miller, 1953). Doswell et al., (1982) suggest that the automated estimates previously used in SELS (Prosser and Foster, 1966) do not have much skill in prediction of observed gust speeds and hail sizes. Shown in Table 2 are the average gust speed errors from that study, based on 1978 severe storm reports. Of significance is the fact that well over half of the reported gusts occurred when the predicted value from the nearest rawinsonde was for no gusts. By excluding the "no gust" forecasts, it can be seen that reported gusts under 65 knots were actually overforecast while those 65 knots or greater were consistently underforecast.

TABLE 3

	Less than 2 inches Preceding	2 inches or Greater Preceding	2 inches or Greater Next
All:	0.95 ± 0.64 (449)	2.21 ± 0.85 (70)	2.08 ± 0.85 (71)
Excl. 0's:	0.76 ± 0.62 (317)	1.94 ± 0.73 (51)	1.97 ± 0.93 (54)
	132	19	17

$\frac{132}{449}$ = 29.4% observed hail $\frac{3}{4} \leq d < 2$ inches with 0.0 calculated (Preceding)

$\frac{19}{70}$ = 27.1% observed hail $d \geq 2$ inches with 0.0 calculated (Preceding)

$\frac{17}{71}$ = 23.9% observed hail $d \geq 2$ inches with 0.0 calculated (Next)

Table 3. Errors in predicted hailstone size: 01 Apr - 30 Jun 1978, for reported hailstones 3/4 inch or greater, in the two categories shown. Values are the mean difference (in inches) of (observed-predicted) ± the associated standard deviation. The number of cases is in parentheses. See Table 2 for explanation of "Preceding" and "Next". Values are given for "All" predictions and also excluding cases when the prediction is "no hail".

In Table 3, the same sort of calculation is shown for predicted maximum hailstone size. Underforecasting is the

general rule, even when excluding the (roughly) one-fourth of reported events which occurred with a "no hail" forecast. Based on these statistics, it seems clear that relatively little skill is apparent.

Sophisticated cloud models, using soundings as input, might be able to provide better quantitative estimates (Chisholm, 1973), but they are not currently practical for operational use. Further, there are simply too many important factors in producing hail and surface wind gusts that are poorly understood, much less routinely observed.

An example of a plotted sounding is shown in Fig. 2.10 with relevant features labelled. Any textbook (e.g., Hess, 1959) provides enough understanding to plot and analyze the typical sounding. There are several aspects of this sounding worth noting. First, it is interesting to observe that the moisture cuts off just below 850 mb, so that the Showalter Index is unrepresentative of the sounding's convective instability - the Showalter Index has a value of +0.2°C, whereas the Lifted Index (based on a forecast surface temperature of 100°F) is -6.3°C! Further, it can be seen that even at a surface temperature of 100°F, a substantial negative area remains to be overcome before the convective instability can be released. Also, note the large positive area in this sounding. The size of the positive area may be a more relevant parameter than 500 mb buoyancy, since updraft speed is only crudely related to the acceleration at any (arbitrary) single level. As discussed in II.III.A.3, the updraft speed is probably a good measure of storm severity.

This sounding also shows an equilibrium level somewhat above the tropopause. Thus, a storm which slightly overshoots the tropopause in this environment is not necessarily severe. In fact, the sounding suggests that a storm which realizes most of the energy implied by the positive area shown will, in all likelihood, penetrate above 100 mb!

3. Sounding Kinematics

Perhaps one of the most widely accepted ideas about the severe thunderstorm environment is that vertical shear is a prime ingredient without which storms are unlikely to develop severe characteristics. This idea is presented quite succinctly by Ludlam (1963): "When there is little or no wind shear the updraft is upright and the precipitation falls through and impedes it."

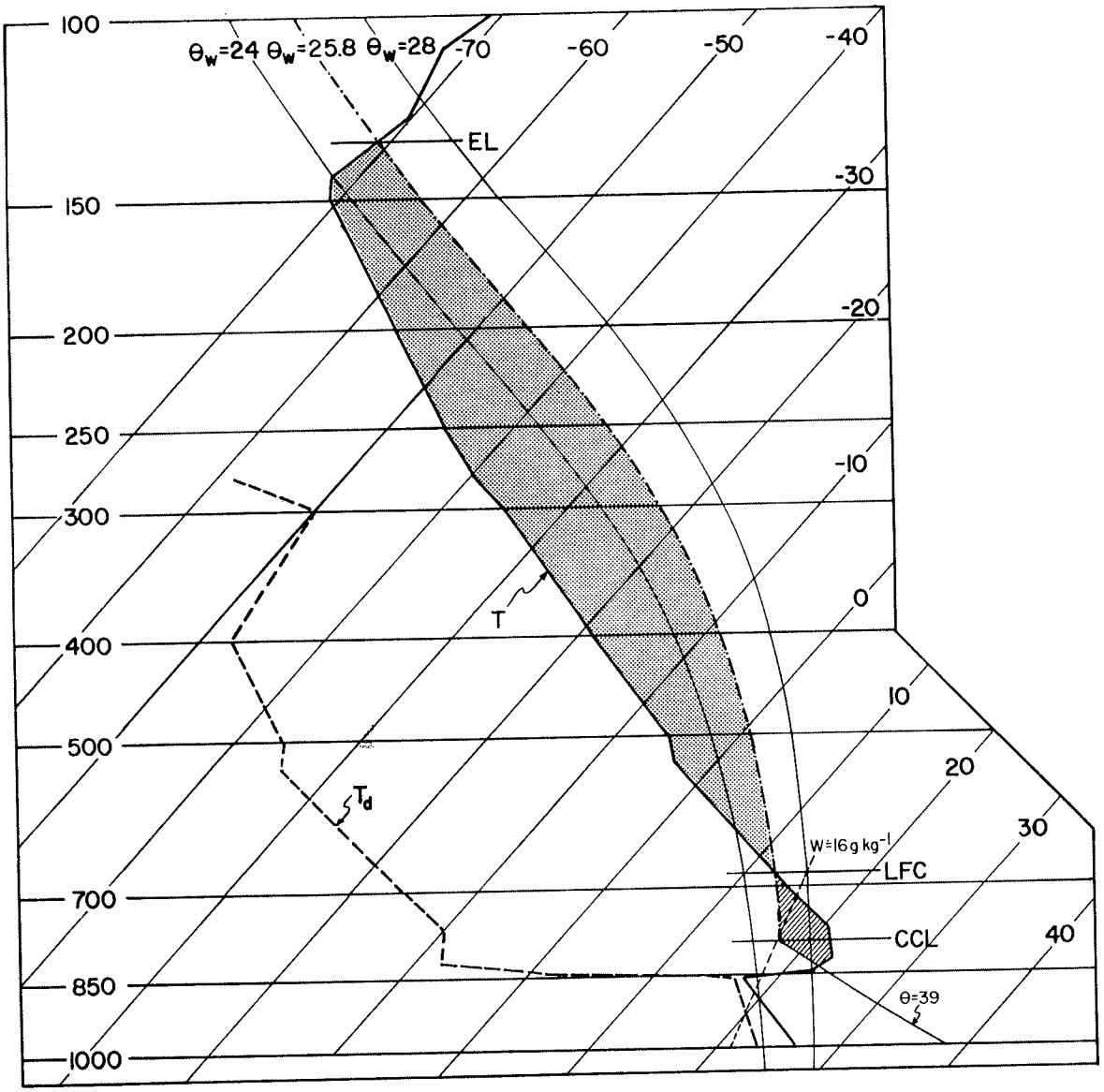


Fig. 2.10. Sample plotted sounding (Oklahoma City, Oklahoma, 1200 GMT, 8 June 1974) on a skew T-log p thermodynamic diagram. Heavy solid line is the temperature (T) observation during balloon ascent, heavy dashed line is for dewpoint (T_d). The thin solid line labelled θ=39 (°C) is the dry adiabat through the forecast maximum surface temperature, while the thin dashed line labelled w=16g kg^{-1} is the mean mixing ratio in the lowest 100 mb. These intersect at the Convective Condensation Level (CCL).[7] From the CCL, the parcel ascends along the pseudo-adiabat labelled θ_w=25.8 (°C), denoted by a dash-dot line. Pseudo-adiabats labelled θ_w=24 and θ_w=28 are the thin solid lines included for reference. The ascending parcel is initially negatively buoyant and the negative area is depicted by hatching. As the parcel rises through the Level of Free Convection (LFC) it becomes positively buoyant, with the positive area stippled. After the parcel rises through the tropopause (about 150 mb) it once again crosses the environmental ascent curve, at the Equilibrium Level (EL).

The common presence of substantial shear in severe storm environments actually has presented a minor paradox. Certain theory suggests that shear can be detrimental to convection, while observations indicate, in general terms, the opposite tendency. This has been reconciled by the suggestion that while shear certainly inhibits weak convection, a sufficiently strong updraft can overcome this tendency and can, in fact, be enhanced by it. The details of this apparent cooperation between shear and updraft have never been completely understood although a variety of mechanisms have been proposed (e.g., Newton and Newton, 1959; Alberty, 1969; Charba and Sasaki, 1971; Rotunno and Klemp, 1982). Model results (Weisman and Klemp, 1982) suggest that supercell storms arise only for a restricted range of shear, with any given amount of instability. The operational utility of these results remains unproven, but they are indicative of the delicate balance required to produce severe thunderstorms. Such details may be below the resolution of operation data sets.

Many believe that the essential feature which allows this interaction to benefit convection is the establishment of a tilted updraft (Fig. 2.11). By this means, precipitation formed within the updraft can fall out of the updraft, rather than having to pass through it. Once again, a potential problem arises since the updraft must somehow tilt upshear, or else the precipitation falls into air that the storm is ingesting, which cools (and thereby decreases the buoyancy) the inflow. This is generally resolved by noting the common severe storm hodograph shows not only shearing, but veering with height.[8] Since parcels rising in the updraft should tend to conserve their horizontal momentum, this is one means of developing the appropriate tilt. Ludlam (1963) also shows (his Figure 21) how upshear updraft tilt can arise from the combined effects of storm motion and the finite time required by successive updraft parcels to rise to a given level.

There are some problems with this picture of how the storm structure arises from its environment. These are also discussed in II.III.A.5. Documentation of many supercell storms suggests that their updrafts are essentially vertical through great depths in the storm. Further, McNulty (1978) has found that "crossover" (the veering of winds with height) is not an essential feature in all severe weather situations. Rather, he found that severe thunderstorms can occur in environments with relatively little directional shear. Crossover may be more important in the process of differential advection than as a means of locating severe

convection directly. Finally, Doswell and Lemon (1979) have found that supercell storms (see II.III.A.5.b) can be found in environments which have a rather wide range of cloud-bearing layer shear values.

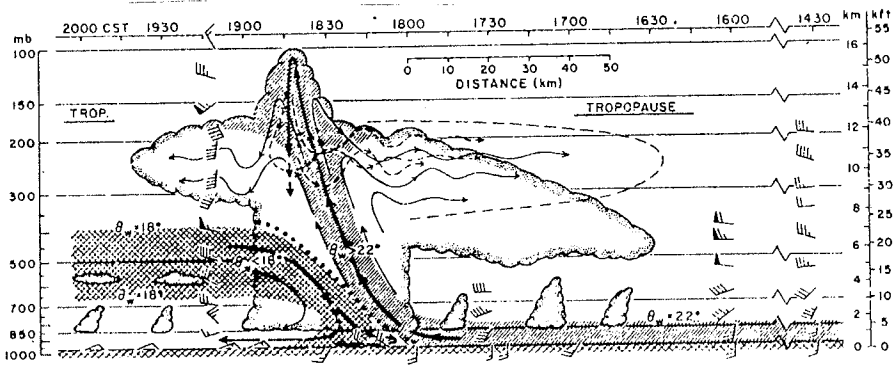


Fig. 2.11. Schematic cross section through squall line thunderstorm of 21 May 1961 (after Newton, 1966). Note the upshear tilt of the updraft. Although cumuliform clouds are shown in the downdraft region to the rear of the storm, there should not be cloud material in the downdraft region; rather, precipitation is descending and becoming less intense as one moves away from the updraft. Vertical scale is exaggerated for clarity.

Marwitz (1972) has stressed that the subcloud layer veering of wind with height is important in differentiating supercell environments from those involving multicellular storms. Doswell and Lemon have supported this concept by suggesting that subcloud layer average shear is better correlated with their supercell sample. This concept is further supported by recent thunderstorm model studies (Weisman and Klemp, 1982) which reveal that hodographs having a veering subcloud wind profile, especially within the lowest 3 km (as well as an appropriate thermal instability structure) seem to be most successful for model simulations of supercell thunderstorms. Such a hodograph is shown in Fig. 2.12.

It is not yet entirely clear that severe thunderstorms are dependent on a given type of hodograph, at least in terms of their large scale setting, as sensed by operational rawinsondes. As developed in II.II.A.3 and II.III.A.4, it seems more likely that the relative flows are more physically significant in producing a particular storm structure. If this is the case, then the explanation of the climatological fact that severe thunderstorms occur most frequently in sheared environments becomes more clear. It is in the sheared environment that a given convective element can develop the appropriate relative flow most easily. Shear, per se, may not be a necessary condition.

This situation creates a forecast problem, since a major element in knowing the relative flow structure is the storm motion, which is not generally known, a priori. From a forecast viewpoint it is probably best to examine hodographs with respect to shear structure first. However, one should also be aware of features in the environment which could result in storm motions favorable for developing appropriate relative flow. For example, Weaver (1979) has pointed out that a persistent stationary source of low-level convergence can dominate the effects of advection, yielding a storm which has relatively little movement. A very slow-moving storm, or one which moves substantially differently from the flow in which it is embedded, can have strong relative winds even when the flow is not highly sheared and/or with only modest wind speeds.

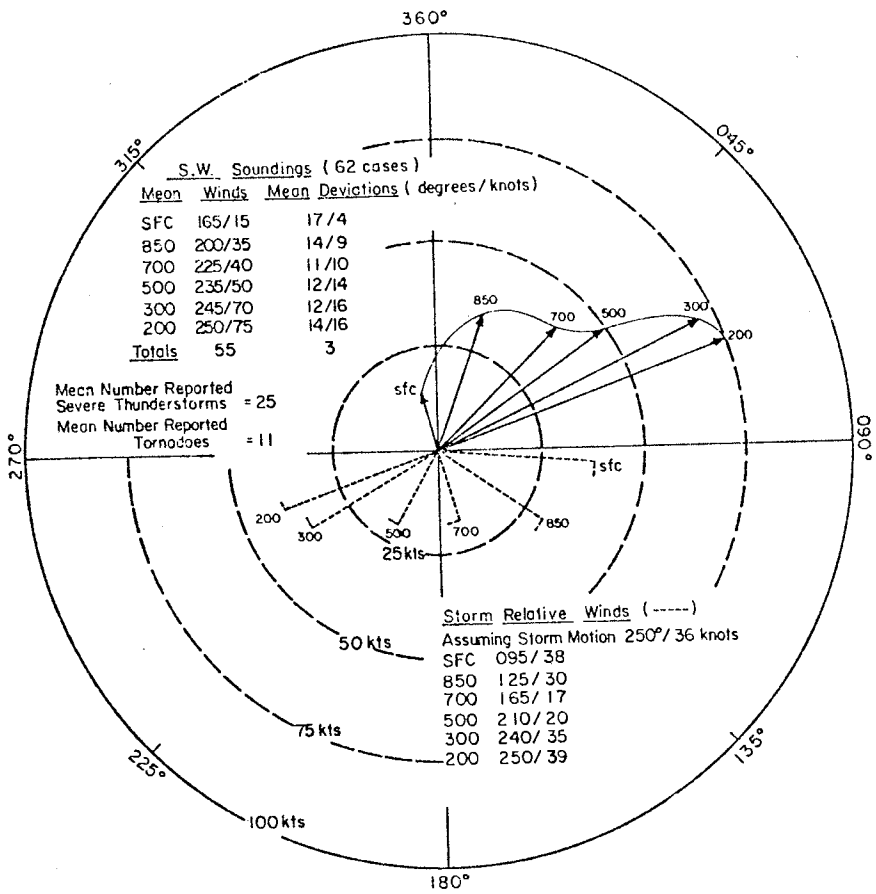


Fig. 2.12. Mean proximity hodograph for 62 cases of severe thunderstorms under southwesterly flow aloft (after Maddox, 1976).

There are certainly features in the hodograph which can be useful in identifying environments which favor, or oppose

development of severe convection (e.g., see Darkow and McCann, 1977). As Maddox et al. (1979) describe, the typical non-severe convective rainstorm exists in an unstably stratified environment with very weak winds through a great depth. This agrees with conventional ideas about shear. Unless storms in a weakly-sheared environment move in ways much different than the mean flow, they cannot develop strong relative flows.

At times, strong shear at very high levels can overlie regions of weak shear. This is not an environment conducive to severe convection, although the average shear in the cloud-bearing layer might be fairly high. The tops of convection may well be literally sheared off and persistent updrafts are unlikely. This is supported theoretically by Schaefer and Livingston (1982), since they find that moderate "shear of the shear" is favorable, whereas excessive changes in shear are detrimental to convection.

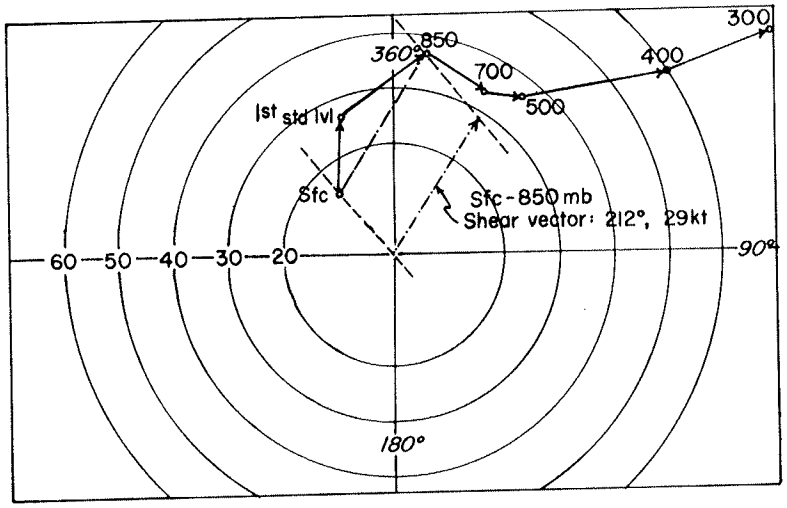


Fig. 2.13. Schematic showing how the hodograph can be used to calculate the shear vector (the dashed vector) graphically between the surface and 850 mb. By translating this vector to the origin, one can see its magnitude and direction. Then, knowing the distance between the top and bottom of the layer, one can easily find the magnitudes of the shear.

The analyst should certainly be prepared to examine the hodographs in situations with relatively weak winds. If the winds are only moderate in speed, directional shear can result in large relative windspeeds, especially since the storms may not move very quickly under those conditions (e.g., Doswell, 1977). Further, in moderate upper-level winds, the low-level jet may assume greater importance. It is not uncommon for the strongest wind in the sounding below, say, 300 mb to be associated with a low-level jet.

This sort of structure can, and does, produce very severe storms even though the cloud-bearing layer shear is only moderate. This is consistent with the enhanced significance of the subcloud layer shear (see the discussion of numerical model results in II.III.D), and with the physical significance of relative winds.

Note that the analyst can evaluate numerical shear values quickly and easily. AFOS can be programmed to calculate shear over various layers, at the discretion of the analyst. This can also be done with a programmable hand calculator. Plotting the hodograph can allow the graphical calculation of shear, as shown in Fig. 2.13. Doswell and Lemon (1979) have summarized their findings on shear values, showing that the average value for supercell storms is about $3 \times 10^{-3}\ s^{-1}$ in the cloud-bearing layer, and confirming the values of Marwitz (1972) for supercell storms. For subcloud shears, Doswell and Lemon find an average value of $7 \times 10^{-3}\ s^{-1}$. As discussed, the analyst should be aware of the broad range of values which can be found in supercell environments.

D. The Composite Chart

The final product of the morning analyses should be the so-called Composite Chart. Guidelines for this have been developed (e.g., Miller, 1972) and there is some value in having a consistently structured composite chart. However, the relevant parameters on any given day may not be useful on some other day. Owing to their obviously central role in severe convection, moisture, vertical motion, and instability parameters should always be included on the composite analysis. Choices for additional parameters should be made by the analyst/forecaster, depending on the situation. For example, instability at analysis time may be rather weak but a low-level jet stream is rapidly increasing moisture in the threat area. On some other occasion, instability may already be substantial so the advection via the low-level jet is not important, whereas an upper level jet streak may suggest a region of vertical motion necessary to break a capping inversion.

Thus, the weather situation should, in part, dictate the parameter choices. However, there exists a necessarily small set of standard features which represent a reliable starting point. For example, this might include the basic features analyzed on the surface chart (fronts, drylines, etc.), low-level moisture fields, static stability, and jet stream axes at high and low levels.

The prognosis should also influence parameter choices. This can be derived largely from numerical guidance, since broad trends are the basic controlling factors. However, one should not be limited to mere reproduction of the model forecasts. Since the analysis should be used, in part, to assess the quality of numerical guidance, a forecast composite is somewhat dependent on the analysis. The decision about what is necessary in constructing the composite analysis and forecast is a feedback process between current and anticipated conditions.

The old adage about expecting the severe weather where the Composite Chart is most illegible has some basis in fact, although only to a limited extent. The forecaster benefits most by the process of preparing the Composite Chart, not by looking at the end result. The necessity for determining the spatial relationships among features (at the various levels in the vertical) ought to be self-evident. Preparation of the Composite Chart accomplishes this. A by-product is that the Composite Chart can be used as a quick reference throughout the day's activities and for briefing the next shift. A mesoanalyst working without a Composite Chart is severely handicapped, and this step should be routine during severe weather analysis.

CHAPTER II FOOTNOTES

[1] p. II-7: It is commonly stated that vertical motion is related to the Laplacian of the thermal advection. This is not true! In effect, the quasigeostrophic equation for vertical motion says that the Laplacian of the vertical motion is proportional to the Laplacian of thermal advection - the Laplacian operators effectively "cancel", giving the relationship as stated by Holton (1979, p. 136ff).

[2] p. II-10: There are numerous definitions by which we express the stability of the stratification (see e.g., Hess, 1959, p. 95ff). The two of most importance here are conditional and convective stability. Conditional instability simply refers to the lapse rate of temperature. If the environmental temperature decreases at a rate between dry and moist adiabatic, the stratification is said to be conditionally unstable. The "condition" in question is whether or not rising parcels can reach condensation. For dry ascent, such an environment is stable, since the dry parcel cools more rapidly than the environment. For moist ascent, the opposite is true. Convective stability is somewhat more subtle. It accounts for the change of moisture with height, as well as the temperature lapse rate. In physical terms, if the bottom of a layer being lifted reaches saturation first it will then cool at the moist adiabatic rate. Should the upper portion continue to rise without condensation, it cools at the much larger dry adiabatic rate. Thus, the upper portion is cooling more rapidly than the bottom. Therefore, within that layer the lapse rate is rapidly increasing. Such a situation results from environments where moisture content decreases rapidly with height. Clearly, there must be enough low-level moisture to allow moist ascent - if the air dries out too quickly off the surface, rising parcels will mix in too much dry air to maintain moist ascent.

[3] p. II-11: These notes make extensive references to wet-bulb potential temperature. One can easily argue that equivalent potential temperature (θ_e) is easier to calculate and equally useful. No attempt will be made to rationalize the choice of θ_w over θ_e - it simply depends on what one is accustomed to. They are essentially the same - there is a (non-linear) one-to-one correspondence (see Hess, 1959, p. 103). However, θ_w is no longer more difficult to compute, if one has a computer and the algorithms given in Doswell *et al*. (1982).

[4] P. II-12: The virtual temperature (T_v) accounts for changes in density as a result of water vapor being present. As discussed by Saucier (1955), T_v is always greater than the actual temperature. The difference (T_v - T), therefore, is always positive. If the moisture content is given by the mixing ratio (r), then to a good approximation, $T_v = T + (r/6)$, where r is given in g kg^{-1}. Since mixing ratios above 500 mb are normally smaller than 1 g kg^{-1}, the difference is negligible at those upper levels. The correction can be substantial in lower levels.

[5] P. II-12: It may be noted that since the winds above, say, 850 mb are pretty close to geostrophic, the change in the real wind often may be a good approximation to the thermal wind.

[6] P. II-16: Note that all uses of the term "Lifted Index" in these notes are based on the original version developed by Galway (1956). Other definitions exist, some of which bear little resemblance to the original - these are still used for facsimile (or AFOS) analysis and forecast charts.

[7] P. II-26: The definition of CCL used here differs somewhat from that of Huschke (1959, p. 134). As used here, the CCL is the condensation level for the well-mixed boundary layer (the lowest 100 mb) at the time associated with the forecast maximum surface temperature. Thus, it is the intersection of the boundary layer mean mixing ratio line with the dry adiabat from the forecast surface temperature.

[8] P. II-27: Recall in the discussion of the thermal wind in II.B.2 above, we found that veering means warm advection. Also remember that warm advection implies upward vertical motion! This line of reasoning must be used with caution, since veering with height is much more common than upward vertical motion. The contributions to vertical motion from differential vorticity advection and from non-quasigeostrophic effects cannot be disregarded, and may often be the dominant factors. Also, low-level flow is less likely to be in geostrophic balance, so quasigeostrophic arguments do not apply.

III. Surface Data Analysis

A. General Remarks

With the exception of remote sensing (radar and satellite imagery), the only source of data suitable for "mesoanalysis" is at the surface. In the central United States, station separation averages about 125 km, with routine data collection every hour. Under certain conditions, special observations are available between the hourly observations. The station density has become increasingly diurnally dependent, with a substantial number of stations closing at night. Also, not every station reports at hourly intervals.

With these data, it is possible to accomplish what is referred to loosely as mesoanalysis in these notes. Under most definitions of mesoscale, these observations are still not dense enough to perform a truly mesoscale analysis. The author has previously chosen to refer to the scale definable by the routine observations as "subsynoptic" scale (Doswell, 1976). Others have referred to this as "meso-alpha" scale analysis (Orlanski, 1975; Maddox, 1979a). Of course, the name is irrelevant, but the intent here is to emphasize analysis on the smallest scale allowed by the data density of routine surface observations. As we shall see, this process can be enhanced and the effective scale limitations reduced somewhat by using remotely sensed data.

Before plunging into the analysis itself, some things need to be emphasized concerning the data. It is an aphorism that mesoanalysts never discard any observations, making every datum part of the analysis. Like most aphorisms, this contains a sizable element of truth. However, there are clearly limitations to this.

Anyone who has done surface analysis comes eventually to realize that some stations have persistent biases in their observations. Assuming that the instruments themselves are not faulty, the main cause of persistent biases is the unique nature of the station location. Station elevations influence the pressure observations, even after the "corrections" to sea level. The surrounding terrain can cause the winds to favor a particular direction. Although every effort is made to get "uniform" instrument exposure, it is simply not possible. Furthermore, the general trend for observation sites at airports creates special problems, since some airports are in urban environments while others are more rural. It is the responsibility of the analyst to determine and correct for as many of these local biases as possible. This should not be confined to one's own station,

but should include all reporting sites within the confines of the analysis region. Generally, mesoanalysis will be confined to a limited area, usually encompassing 3-5 states, called a "sectional" chart. This is done to allow the plotting and analysis at hourly intervals. It would be desirable to document these biases and have the documentation available for training new local analysts, as necessary.

The problem of occasional "errors" in the data is somewhat more difficult to deal with. Some percentage (unknown) of these are actually valid observations of the variables, but are unrepresentative in some sense. A station experiencing a thunderstorm will frequently, as we shall see, have surface conditions that appear anomalous in comparison with its neighbors. It is precisely this sort of observation we wish to retain and use to construct an operational mesoanalysis. Not all such mesoscale features are possible to identify with such ease. Further, even when the phenomenon itself is recognized, its quantitative influence on the station data is not always clear. This is especially important for objective analysis (see IV.C). When doubts about the validity of an observation arise, it is probably wise to err on the side of retention of the datum. One additional test that may be valuable is to examine how the analysis looks with the datum accepted -- if the resulting pattern does not fit any recognizable mesoscale structure, it is likely to be erroneous. Clearly, in this context, it is to the analyst's advantage to be familiar with convective mesoscale systems. Naturally, it is also beneficial to consider independent data sources such as satellite images to help determine a datum's validity. A final tool to use in this process is continuity in time -- if a feature persists through several observations, it is probably valid.

In the typical situation, surface observations come in as a "sequence" (or on AFOS, in batches). In order to do an analysis, these need to be plotted on a map. In the interest of saving time, a given forecast office has only to plot a sectional map, an example of which is given in Fig. 3.1. Data are plotted according to the station model shown in Fig. 3.2. Using AFOS, such a chart can be machine plotted rather quickly. It is unlikely that a really detailed "mesoanalysis" (recall that this is actually a subsynoptic scale analysis) can be done hourly for an area much larger than that shown. Even national centers like NSSFC must focus attention on a confined region in order to accomplish a detailed hourly analysis.

Note that the altimeter setting is plotted,[1] rather than the so-called "sea level pressure." This is done for

several reasons (Magor, 1958). The primary reason is that there is a 30 to 40 percent increase in the number of data points. Almost every reporting station transmits an altimeter setting, even with the off-hourly "special" observations. This, by itself, is a substantial advantage. Further, the altimeter setting is related to the station pressure in an identical manner at each observation point.

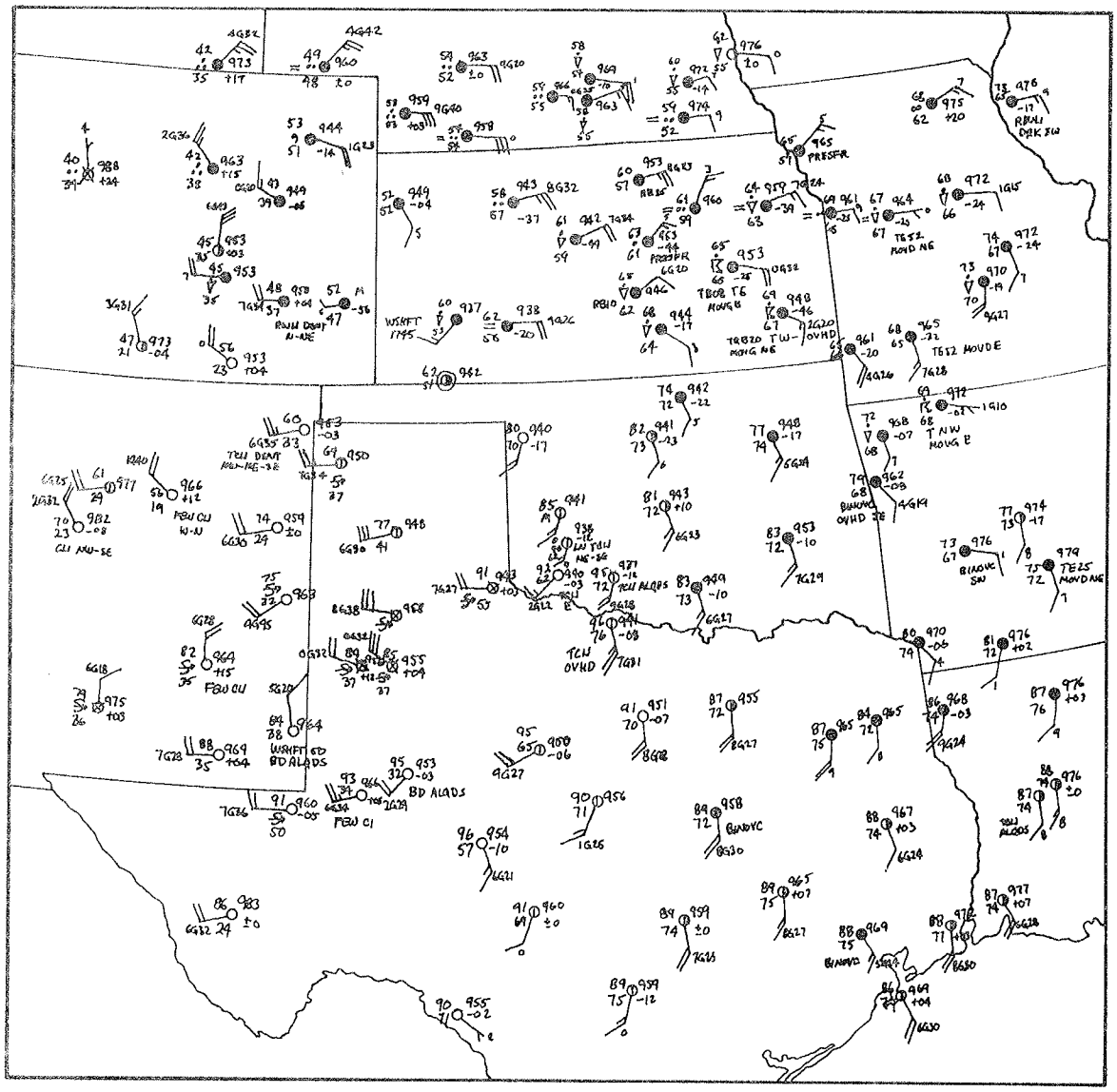


Fig. 3.1. Sectional plot of surface observations at 1800 GMT, 8 June 1974. Station model for plotted data given in Fig. 3.2.

Also note that significant remarks are plotted. These can frequently provide important clues to the analyst, and are a valuable complement to the remote sensors like radar

and satellite cameras. If possible under the time constraints, these additive remarks should be included in the station plot, perhaps abbreviated (e.g., "PRR" for "PRESRR").

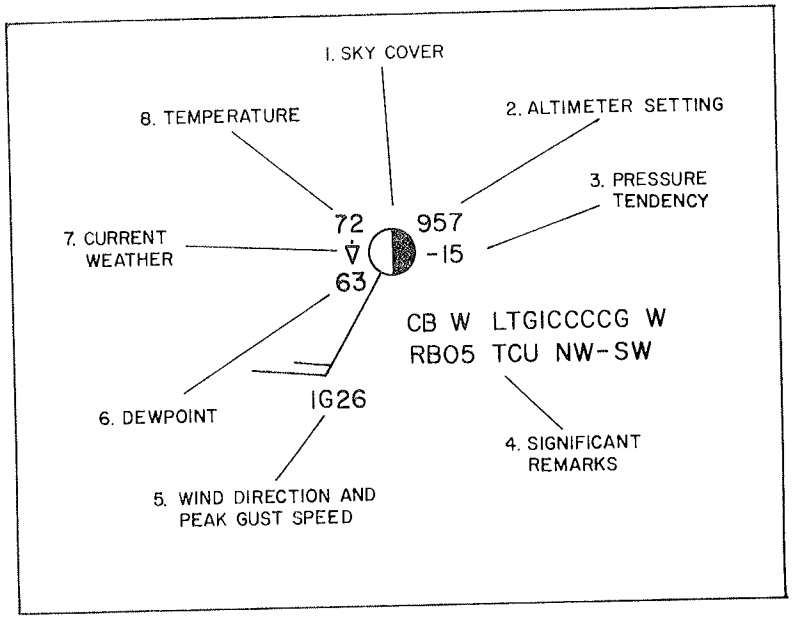


Fig. 3.2. Station model for plotted surface observations. Note that the second digit of the wind direction is indicated - in this example, a wind from 210° has a "1" shown. Wind speeds are in kts, temperature in °F, altimeter setting in hundredths of an inch of mercury (leading digit suppressed - in this example, 29.57 in Hg = 957) and 3-h pressure tendencies in tenths of a millibar (down 1.5 mb over past 3 h = -15).

B. Surface Discontinuities

Much emphasis has been put on discontinuities in the surface fields, and rightly so. The identification and prognosis of boundaries is generally a key element in producing a forecast. In terms of what the atmosphere does, not much "weather" goes on in regions of uniformly distributed weather elements. Generally speaking, what we perceive as weather is the result of atmospheric processes acting to relieve some form of non-uniformity. Weather continues until the non-uniformity has been eased to the point where the process can no longer be sustained. This statement applies to all scales of weather phenomena, from global circulation patterns down to microscale activity.

While these notes place a substantial emphasis on proper analysis of surface boundaries, the analyst/forecaster is cautioned not to place so much effort into analysis of boundaries that the basic physical reasoning process suffers.

There are many factors to consider, and one should avoid getting "hung-up" in a complex weather situation in trying to fine-tune the surface analysis. A surface analysis is, after all, a working chart, not a work of art.

Probably the most commonly recognized surface discontinuities are air-mass boundaries. This broad class of boundaries includes true fronts, drylines, sea and land breeze fronts, and thunderstorm outflow boundaries. These notes do not dwell on the details of each of these, and once again the reader is urged to examine the references. Rather, we wish to point out some features to look for in helping to locate these boundaries properly on the sectional mesoanalysis. As in the case of upper air analysis, the end result of surface analysis is generally to locate and forecast where upward motion is found to coincide with regions of convective instability. The lift associated with surface convergence tends to be concentrated in the vicinity of these boundaries.

First of all, consider the true front. By definition, a front is a zone where atmospheric density varies substantially. Basically, the frontal zone separates two distinct air masses. The width of the frontal zone can vary greatly, and only approaches a true discontinuity in an idealized, limiting case. Nevertheless, we choose to draw "the front" as a line on the weather chart. Typical large-scale surface frontal gradients are in the range of $10^{\circ}K$ per 100 km. The question might arise as to where in the frontal zone do we place the hypothetical line called "the front"? The convention is to place the front on the warm-air side of the frontal zone. If we consider the isotherm analysis of Fig. 3.3, derived from the data shown in Fig. 3.1, we would then have a possible frontal structure as indicated. It should be pointed out that if 20 meteorologists started with the data of Fig. 3.1, we would probably have 20 different analyses of the fronts and other surface boundaries. This not necessarily bad. A careful examination of the data in Fig. 3.1 suggests that the frontal positions are quite clear in some areas and not so clear in others. This situation is most common. Real surface data simply do not fit the idealized patterns which are usually chosen for presentation in textbooks. This is not a fault in textbooks, since their aim is generally to increase the reader's basic physical understanding, not to teach one how to analyze real data.

It is also possible to define air-mass boundaries (not necessarily fronts!) quite effectively by calculating and plotting the observations of wet-bulb potential temperature

(θ_w). Since air mass density is influenced by pressure, temperature and moisture content, the θ_w field is quite well-suited for this purpose. See Fig. 3.4 for a θ_w analysis associated with the data from Fig. 3.1. Note that virtual temperature (recall II.B.2, above) is directly proportional to density, so that a T_v analysis gives as clear a picture of frontal locations as is possible.

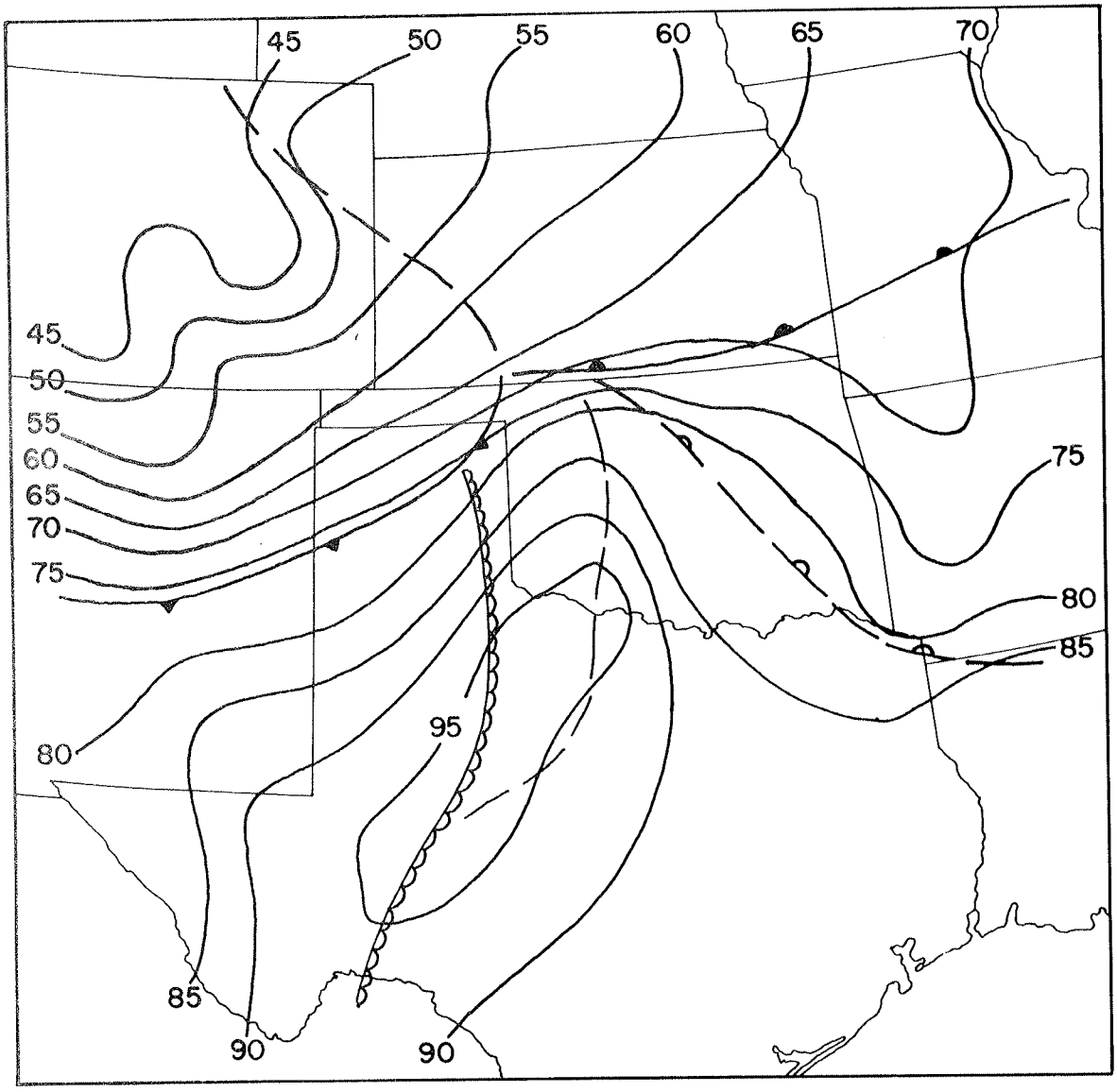


Fig. 3.3. Isotherms (°F) and one possible frontal analysis for data in Fig. 3.2.

A well-recognized method for locating fronts is to position the front in the zone of strong cyclonic wind shear which usually accompanies a change of air mass. That such a

wind shift is associated with frontal zones is a subject well-covered in textbooks (see, e.g., Hess, 1959; p. 230ff), and is a natural consequence of the strong density gradients involved. Furthermore, such a wind shift is implied also by

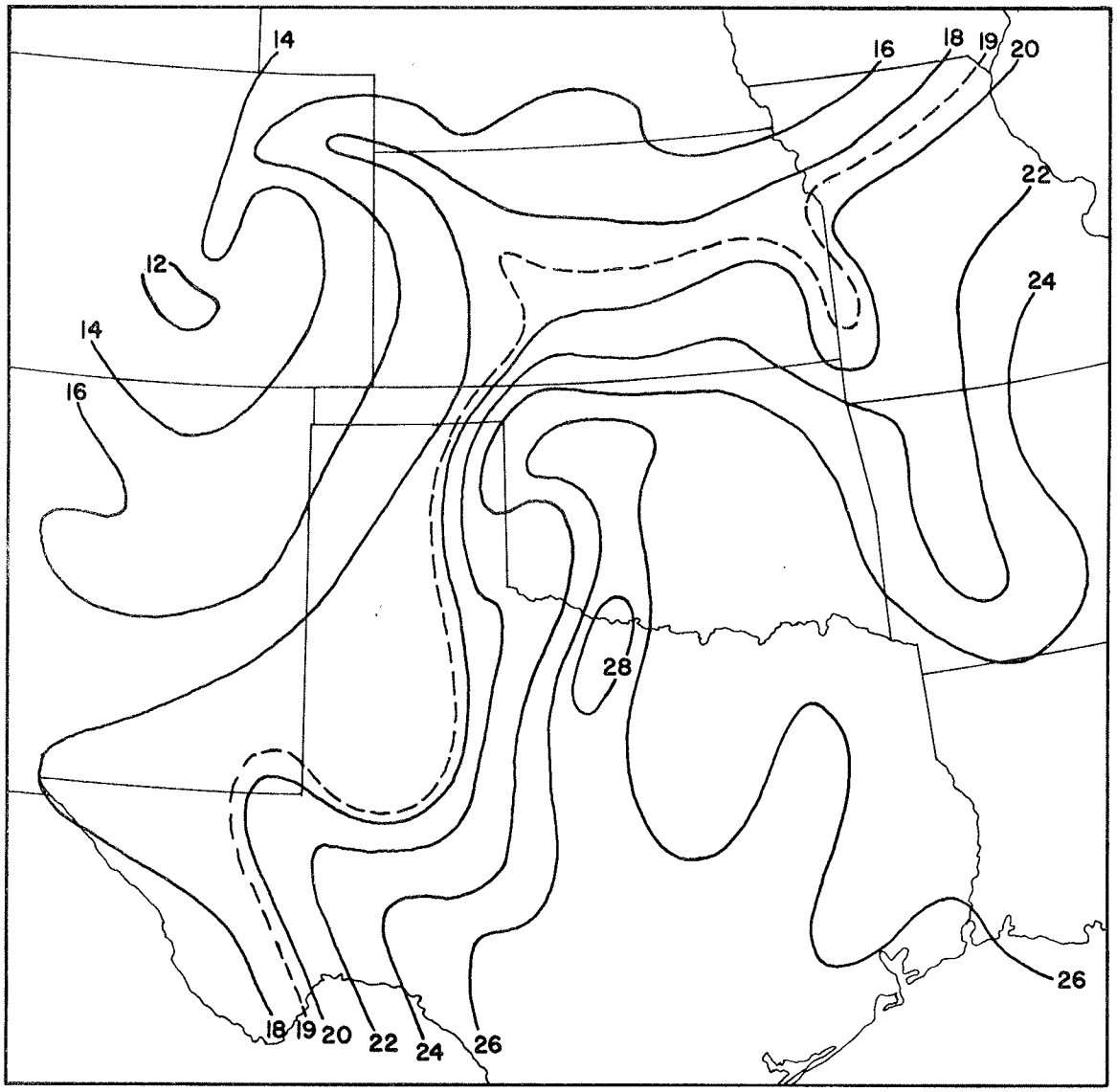


Fig. 3.4. Isotherms of θ_w (°C) for data in Fig. 3.1.

the pressure trough in which fronts tend to occur. Since a front is located generally in a pressure trough, pressure tendencies can be used to locate the frontal boundary (see III.D.1). These relationships are all theoretically justifiable and show a remarkable agreement with the typical real weather situation. However, the analyst should certainly be aware that all wind shift lines and pressure

troughs are not frontal in nature. Further, not all fronts have easily recognized cyclonic wind shear across them, and occasionally liberties must be taken with the pressure pattern to put the front in a pressure trough.

This brings to mind the often-discussed issue of whether or not to "kink" the isobars across a front. At times, in actual practice, it is more clearly justified than others. Naturally, as we have already mentioned, a front is only a transition zone and not a true discontinuity. Therefore, it would seem reasonable to suggest that the "kink" is only a theoretical artifact and not a reflection of the true pressure distribution. Nevertheless, this issue is actually more an aesthetic or pedantic one. In short, it really doesn't matter except to purists (on both sides of the issue). Rather, one should concentrate on the proper pressure analysis all over the chart.

1. Cold Fronts

By definition, a cold front is a frontal boundary along which warm air is being replaced by cold (Fig. 3.5). In the northern hemisphere then, the basic flow behind the front typically has a northerly component, since cold air generally has its source region to the north. When the cold air begins its southward movement away from its source, the contrast between the cold air and the air it replaces is

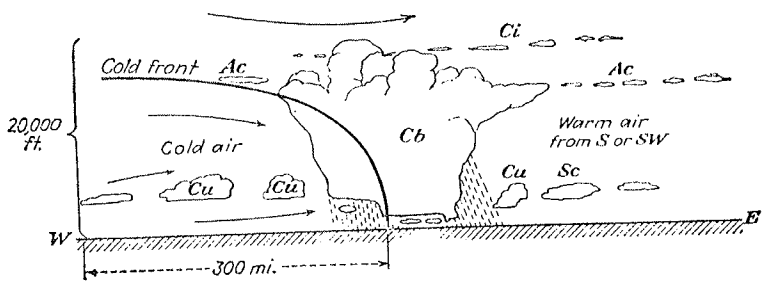


Fig. 3.5. Schematic cross section through cold front, showing cloud and precipitation (after Byers, 1959).

often large, and the front is quite easy to locate. By moving cold air to the south, the atmosphere is trying to equalize the temperatures over the earth. As the cold, usually dry, air moves southward over warmer ground, it is gradually heated from below. Since the warm air created by contact with the surface naturally tends to rise, this heating is quickly spread to great depths. Thus, it takes a relatively long time to note the surface modification of the air mass, since the heating is so quickly dispersed

vertically. The opposite is true when the front is moving over colder surfaces. Situations where a cold front moves over colder ground are relatively rare in the United States, except perhaps when maritime polar air from the west displaces modified artic air masses. In such situations the cold front may be hard to find at the surface, since a surface inversion is created which tends to mask the normal features associated with a front (see Saucier, 1955, p. 297).

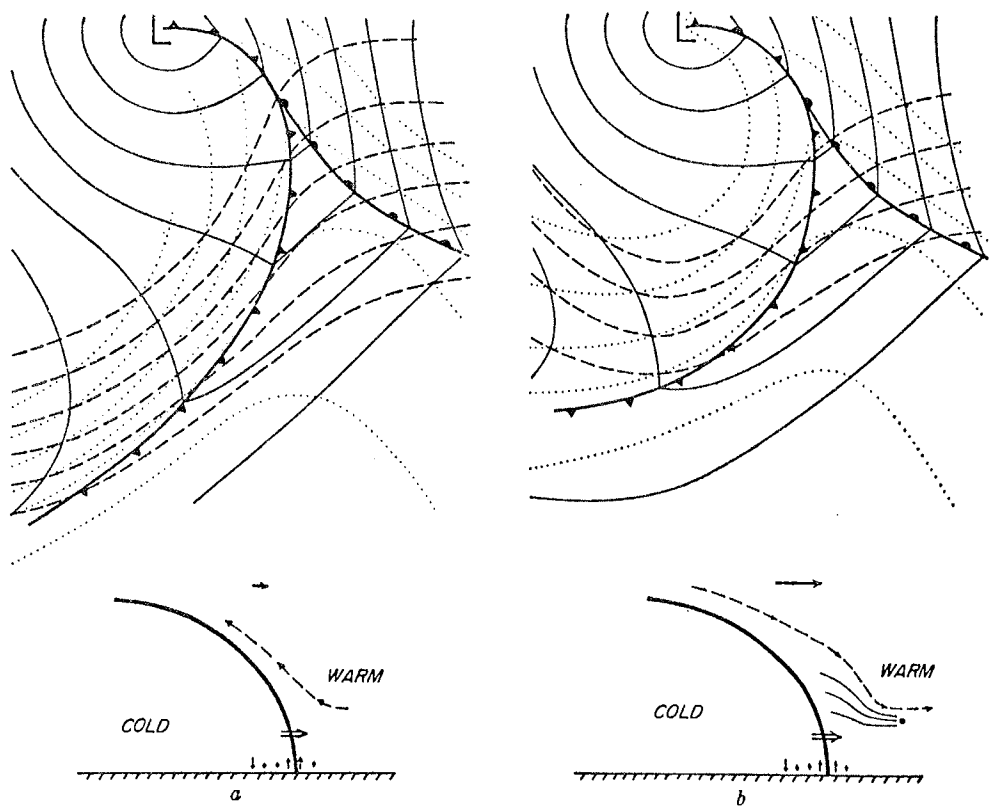


Fig. 3.6. Schematic of two types of cold fronts (after Saucier, 1955). In upper figures, solid lines show contours of the 1000 mb surface (with frontal analysis), dashed lines are contours of an upper level isobaric surface (say, 700 mb), and dotted lines depict the thickness contours between them. In (a) the front is moving faster than the normal wind component aloft. Therefore, relative to the front, warm air is lifted "upslope", producing precipitation behind the frontal zone. In (b), the front is moving more slowly than the normal component aloft. Thus, there is "downslope" movement and subsidence of the warm air over the front, giving clear air behind the surface boundary. Any weather associated with type (b) is likely to occur ahead of the front, perhaps by several hundred km.

While the basic large-scale flows imply surface convergence along the frontal zone, care must be exercised in using this generalization. The obvious picture of the cold front acting as a wedge, lifting the air ahead of it is probably not a bad concept, but the presence of lift along the frontal boundary depends critically on the relative, normal components of the winds on either side of the boundary (Fig. 3.6), as described by Saucier (1955, p. 292ff). One really needs to examine the wind field in the vicinity of the front carefully, as well as the speed of frontal movement, to complete the picture. Note that the structure of the pressure and temperature fields along a cold front tend to increase the cyclonic shear across the front near the surface (Petterssen and Austin, 1942). Therefore, cold fronts are usually characterized by abrupt wind shifts.

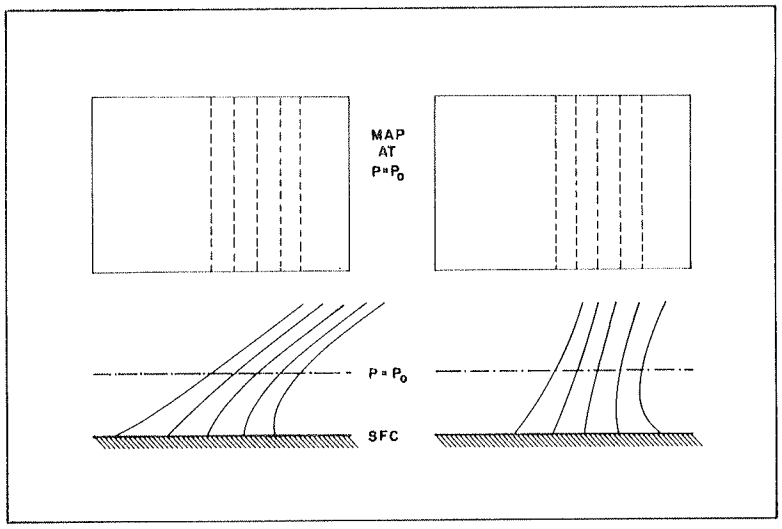


Fig. 3.7. Schematic showing how fronts with different slopes can give the same horizontal temperature field on a given isobaric surface.

By simple density considerations, the stronger the density contrast, the shallower the cold air mass tends to be. Therefore, the overall frontal slope is less when the cold air mass is markedly colder than the air it displaces. As a result, the airmass boundary aloft may not be easily identifiable much above 850 mb, and the horizontal displacement of the boundary with height can be large. When the air masses are only weakly different, the frontal zone can be more nearly vertical. Note in Fig. 3.7, that the resulting isotherms on an isobaric surface which intersects the boundary may have similar gradients in both cases, or the 'weak' front can actually show a stronger horizontal gradient on a given isobaric surface!

A final word on cold fronts concerns the use of dewpoint gradients to locate the boundary. As we shall see, a dewpoint difference is not necessarily indicative of a frontal boundary. Further, if maritime air is displacing continental air, the dewpoint difference may be negligible or, indeed, may be such that the cold air is also wetter. As with any rule of thumb, the use of dewpoints to locate fronts must be done with caution, using other information like winds, temperatures, and upper-air data, as well as common sense. This further suggests that one should not anticipate the post-frontal weather based on classical models. In many cases, cold-frontal passage marks the onset of cloudiness, not a clearing line.

2. Warm Fronts

As one might expect, along a warm front cold air is being replaced by warm (Fig. 3.8). In the northern hemisphere, this suggests that the basic flow behind the front generally has a southerly component. Unlike cold fronts, when a warm air mass begins moving northward, it typically originates in a region where the air mass contrast is weak and the front may well be difficult to locate at the surface. Since the air is often warmer than the surface over which it moves, the presence of a surface inversion can make the boundary difficult to find early in the day, until surface heating can break the inversion.

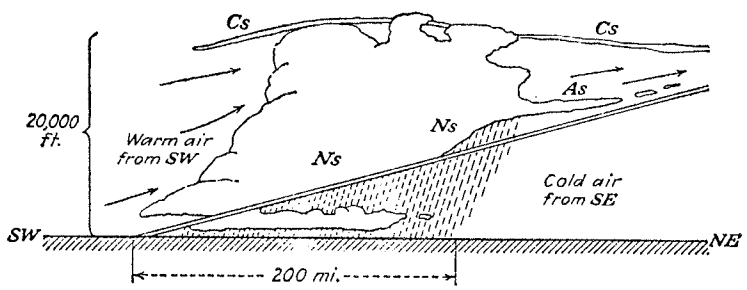


Fig. 3.8. As in Fig. 3.5, except for a warm front (after Byers, 1959).

As with the cold front, there is a general picture of the flow pattern associated with a warm front which has wide applicability. That is, it is generally conceived in terms of the warm air gliding up over the retreating cold air. The same problems exist with this concept as with that associated with the cold front, since the true picture depends on the relative normal wind components and how they vary with height. The concept of "overrunning" can be valuable

when combined with a clear understanding of how vertical motion arises. However, "overrunning" is a term often over-used and inappropriately applied.

By finding the 850 mb warm front, it is often possible to return to the surface data and find clues as to the location of the warm front that might have been too subtle to see at first glance. Since the thermal contrast at low levels across a warm front tends to be weaker than that associated with cold fronts, the picture shown in Fig. 3.7 suggests that a possible aid to warm frontal location is by first checking the isotherm pattern at 850 mb. Experience tends to bear this out. As Petterssen and Austin (1942) have shown, the temperature and pressure fields along a warm front act to decrease the cyclonic shear across the boundary, so the surface wind shift is often very subtle. A convenient rule of thumb which can be helpful in locating warm fronts is the observation that isobars in the warm sector of an extratropical cyclone are usually more or less straight lines. Isobars with substantial curvature are generally on the cold side of the warm frontal boundary.

As with cold fronts, the dewpoints can be useful, provided the analyst is aware of the source region for the advancing air. In the central United States, most warm air masses have a trajectory over the Gulf of Mexico's warm waters and are, therefore, usually more moist than the air they displace. However, the descent of moisture from precipitation into the cold air mass in case of "overrunning" can act to smear out or displace the dewpoint gradients. Of course, some warm air masses are continental in origin or may be associated with subsidence, implying that they might be drier than the displaced air. As with cold fronts, the analyst/forecaster should avoid uncritical use of classical models of warm front weather sequences.

3. Stationary Fronts

Although the obvious definition for a stationary front is one which has neither air mass advancing, it must be modified somewhat, since it is rare that the boundary is truly stationary along its entire length. The term frequently applied is "quasistationary" which generally requires that the movement be slow (say 10 knots or less) or erratic. It is not uncommon for a "stationary" front to move slowly across a state over a 24 hour period.

Fronts become quasistationary when the flow normal to the boundary becomes negligible. This can arise either when the front has a weak flow field or when the winds become

essentially parallel to it. In the former situation, the thermal contrast across the front is also usually weak and the frontal zone is quite diffuse. In such a case, it may be said that the front has "washed out" and the front dropped from the analysis. It is noteworthy that the old frontal zone may still be characterized by a pressure trough (see III.C.2) and significant weather may be found roughly colocated with it.

Stationary fronts which retain a recognizable contrast are quite significant, since "overrunning" may be a possible weather threat. This is essentially a situation where some mechanism creating vertical motion moves over the cold air and has a source of warm, moist air (generally to the south of the boundary) to sustain significant convective weather. In these cases, depending on how far north of the boundary the vertical motion is occurring, there may not be any easily recognized surface feature prior to the weather events. Satellite imagery is very useful for locating and forecasting regions of upward motion with roots above the surface. Also, numerical model forecasts may give an indication of some upper-level feature which is moving over the surface cold air dome.

Most of what relates to locating warm and cold fronts can also be applied to stationary fronts. With respect to winds at the surface, it is not uncommon for substantial flow normal to a front to occur on the warm air side of the boundary without significant frontal movement. This is the typical case in "overrunning" situations. Examination of the winds aloft often reveals that the winds become essentially parallel to the front a short distance off the surface, which hardly fits this simple picture. In such cases, some source for vertical motion over the front can usually be found (recall the discussion in II.B.1).

When a vertical motion field passes near the surface boundary, it is common for cyclogenesis to occur, and an extratropical cyclone may develop along the boundary. This process is the classical sequence of development for such storms and is thoroughly discussed in the references.

Alternatively, a similar development can take place, but on a smaller scale, in which a weak "frontal wave" is formed (Fig. 3.9), which never goes on to develop into a synoptic-scale weather system. Such systems are only hundreds of km in diameter, as opposed to thousands of km for extratropical cyclones. This process of weak wave formation can occur several days in succession along a quasistationary boundary, producing a situation where heavy

precipitation and severe weather of various types (depending on the season) occur repeatedly over a limited area. The dynamics of such weather systems are not well understood, but careful analysis can be of value to the forecaster, since these systems tend to concentrate low-level convergence (values of order 10^{-4} s^{-1}, implying upward motion of order 10 cm s^{-1}) into localized areas (Tegtmeier, 1974; Doswell, 1976).

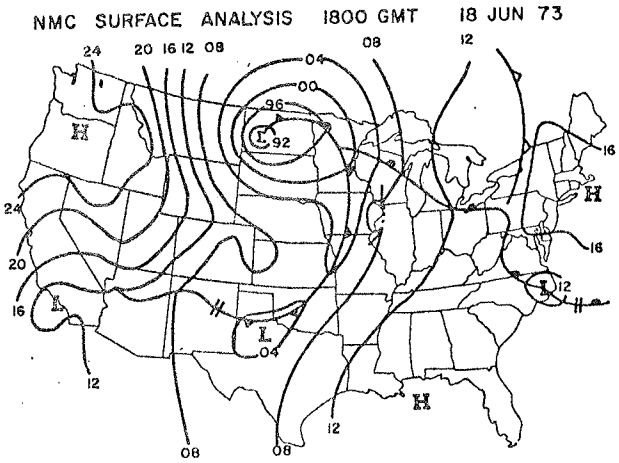


Fig. 3.9. Surface analysis showing example of a weak frontal wave in Texas and Oklahoma.

4. Occluded Fronts

In the standard picture of the extratropical cyclone, the cold air moves southeastward faster than the warm air moves northeastward. This leads to the cold air overtaking the warm front and the initiation of the occlusion process. In terms of the extratropical cyclone's life cycle, intensification of the large-scale storm occurs when it has energy available from the air mass contrast. By moving warm air upward and northward, as well as cold air downward and southward, the storm acts to diminish the available energy used for intensification. By the time occlusion begins, this air movement has usually exhausted the cyclone's potential for further development. Occlusion proceeds because the flow pattern established during cyclogenesis does not just cease after its amplification stops. Rather, mixing of the contrasting air masses continues essentially on its own inertia. When a cyclone is thoroughly occluded, it has succeeded in mixing the air masses so well that near the center of the low it becomes quite difficult to locate any significant air mass boundaries. However, wind shift lines and pressure troughs may continue to rotate around the low like "spokes in a wheel" (see

Kreitzberg and Brown, 1970 for more details about the features in an occluded system).

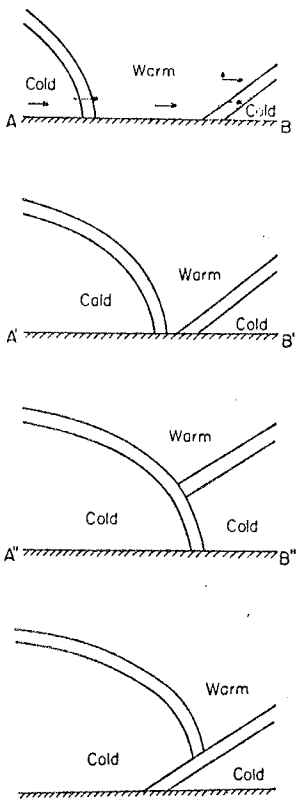


Fig. 3.10. Schematic example showing formation of an occluded front, with cold front overtaking warm front (after Saucier, 1955). In the example on the bottom, the cold air ahead of the warm front is colder than that behind the cold front, giving rise to a warm-type occlusion. This is relatively rare in the U.S. In the case just above, the cold air behind the cold front is colder than that ahead of the warm front, yielding a cold-type occlusion.

Traditionally, discussions of occluded fronts go to great pains to differentiate between "warm type" and "cold type" occlusions (see Fig. 3.10; also Saucier, 1955, p. 268 ff). The cold type is probably the most common in the United States. As Saucier (1955, p. 271) points out, the upper warm front in a cold occlusion is not usually significant insofar as surface weather is concerned. Since the wind shifts and pressure trough need not be coincident with the surface location of an occluded front, it is debatable whether or not occluded fronts per se are really a significant aspect of surface analysis. Generally, the surface trough (in which the occluded front is usually inserted) actually reflects the axis of the deepest warm air (a thermal ridge), rather than a front in the strict sense. In fact, the region near the center of an occluded cyclone can be so thoroughly mixed that the analyst might better spend time on other aspects of surface analysis than the confusing problem of locating an occluded front.

As a final word with regard to occluded systems, the analyst needs to be alert to the possibility of intrusions of "fresh" air masses into the circulation, and to structures conducive to frontogenesis. Although classical pictures of occluded systems indicate little likelihood for additional significant convective weather, there are two

situations that bear watching. First, if the circulation (which may persist for days after occlusion begins) can tap another air mass, a new "boundary" can be created which may result in an abrubt shift in the location of significant weather with respect to the circulation. Second, air mass modification can occur in the clear zone following a closed system's frontal passage. If surface heating is sufficient, a rapid destabilization can occur in this clear air (often near the core of the vertically stacked circulation), resulting in convection and severe weather on a local scale. Such developments can also be associated with secondary shortwave troughs rotating through the large scale occluded system (see III.C.2). These secondary systems often have frontogenetic circulations and enough vertical motion to contribute to destabilization.

5. Drylines

The dryline is a subject which has received relatively little formal attention in textbooks, but forecasters in the Great Plains must frequently deal with it in relation to convective weather. Schaefer (1973a,1974a,b) has given the most detailed accounts of the structure and origins of the dryline as a synoptic scale feature, while Rhea (1966) has discussed the occurrence of thunderstorms in relation to the dryline. Since this feature is not well-recognized outside the regions where it frequently occurs, it is often mistakenly analyzed as a "Pacific front" - i.e., the leading edge of an air mass with a Pacific source region.

Research has emphasized the regional character of the dryline - its occurrence at the surface is generally confined to the Great Plains, west of the Mississippi River and south of the Dakotas. This is generally recognized to be the result of the sloping terrain. As moisture returns (from the Gulf of Mexico) to the plains following the passage of an anticyclone, the rising terrain limits its westward penetration and shunts it northward (Schaefer, 1974b). Thus, a natural tendency exists for the development of a north-south boundary separating dry from moist air. The moisture gradients observed with mesoscale networks can be enormous (Fujita, 1958; McGuire, 1962), with mixing ratio changes of 5 g kg^{-1} over a distance of 1 km. The operational data do not provide the detail to resolve such intense gradients, of course. However, time series observations at specific stations often reveal dewpoint temperature drops of 30°F (~16°C) in a matter of minutes.

The dynamics of the dryline flow regime are dominated by boundary layer processes, as suggested by Schaefer's work

(1974a,b). To summarize briefly, the development of a well-mixed boundary layer during the morning acts to disperse moisture vertically. If the moisture is shallow, this can be seen as a "movement" of the dryline past the station, at which time the surface dewpoint drops while the temperature rises. Such movement often is related very poorly to the wind field, under quiescent conditions (i.e., when the dryline is not involved in a synoptic-scale circulation). With the re-establishment of the inversion during the evening, the dryline "backs up" and its motion then is related much more clearly to the winds in the moist air as it returns.

It should be pointed out that the dryline is not a front, in the sense of a density discontinuity. During the morning hours, the air on the dry side of the boundary is quite cool, since dry air (usually cloudless) enhances radiational cooling. However, by early afternoon, the dry air normally is warmer than that on the moist side and the resulting surface virtual temperatures (or, equivalently, the densities) across the dryline are essentially equal on both sides. The morning thermal contrast is not a real frontal characteristic, but it can be quite deceptive to an unwary analyst.

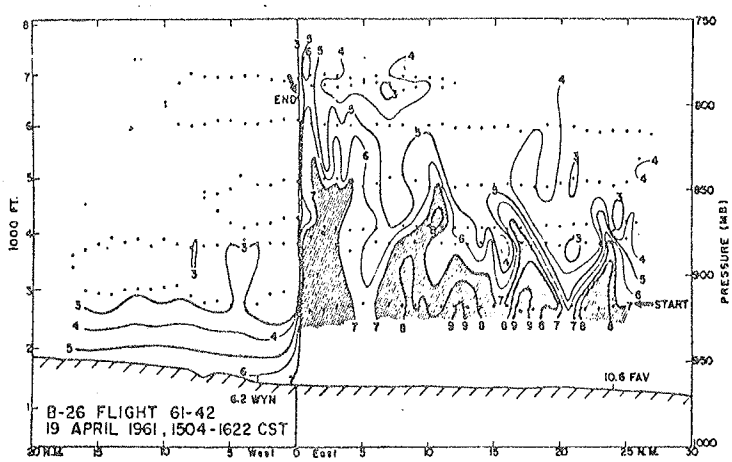


Fig. 3.11. Aircraft traverse through dryline (after Staff, NSSP, 1963) showing mixing ratio isopleths (g kg^{-1}). Note the very intense moisture gradient, the nearly vertical boundary, and the evidence of wave-like perturbations on the wet side of the dryline.

Further evidence of the non-frontal nature of the dryline is the finding that the interface between dry and moist air is nearly vertical off the surface, and quasihorizontal to the east (Fig. 3.11; recall the discussion in III.B.1).

As Rhea (1966) and Schaefer (1975) point out, the dryline is typically located in a zone of small scale convergence. This convergence is reflected typically in a narrow line of cumulus clouds which is colocated with the dryline. On a somewhat larger scale, the convergence of the routinely observed surface wind field often actually reaches its maximum somewhat behind the dryline, while the moisture convergence maximum is right along the boundary (Doswell, 1976). The velocity convergence maximum behind the dryline usually lies within a region of strong downstream speed decreases. Although the dryline itself may be located in the area of rapid directional changes, the convergence is usually dominated by the speed changes upstream. Since moisture convergence also depends on the moisture field, the combined effect usually is to put the strongest moisture convergence near the dryline axis. As an aside, the reader should exercise caution in "eyeballing" convergence patterns from the plotted wind data.

Recall that frontal boundaries are defined to lie on the warm side of the gradient, by convention. No such convention exists for the much less well-documented dryline. As suggested by Fig. 3.11, the moisture gradient can be so strong as to be indistinguishable from a true discontinuity on a normal surface map used for analysis. Of course, conventional surface data are nowhere near being dense enough to sense this intense gradient - so the gradient is seen to be smeared out. Schaefer's work follows, insofar as possible, the convention that the dryline is on or near the $45^{o}F$ ($\sim 7^{o}C$) isodrosotherm. This also corresponds roughly to a mixing ratio of 9 g kg^{-1}. Adjustments can be made on the basis of veering winds, as the dryline moves past an observation site. It should be emphasized here that, especially during the morning hours when the dryline is usually not moving rapidly, winds can be very deceptive for dryline location. The morning inversion has de-coupled the surface winds from the significant free atmosphere flow, thus making local effects more important.

Although most of Schaefer's studies concern the dryline under quiescent conditions, they are of wider applicability, subject to some modification. The forecaster should be aware of moisture depth as a clue to dryline motion - if the moisture is shallow, it is likely that the solar heating can break the morning inversion quickly, thus suggesting an early dryline "passage". Deeper moisture will slow down the apparent movement of the dryline during the day.

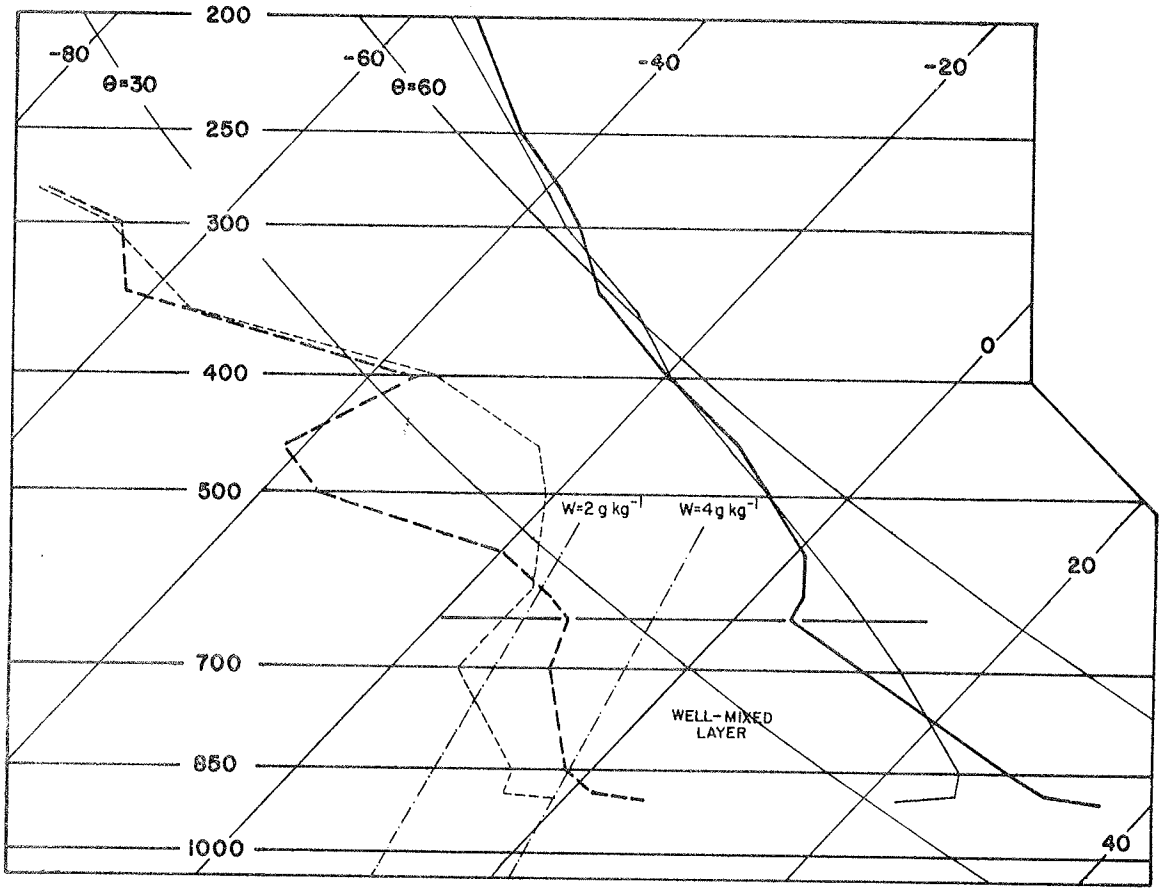


Fig. 3.12. Sample soundings through dry air on west side of dryline (Midland, Texas on 8 June 1974). Thin lines show temperature (solid) and dewpoint (dashed) curves at 1200 GMT. Thick lines show similar curves at the following 0000 GMT sounding time. Dash-dot lines show representative mixing ratio lines on the chart. Note shallow, surface superadiabatic "contact" layer at 0000 GMT.

When large synoptic-scale systems are present over the plains, the dryline is often a major factor in the day's severe weather forecast problem. It is well-known that a deep surface-based layer of nearly dry-adiabatic lapse rates (as shown in Fig. 3.12)[2] is present on the dry side of the dryline (Carlson and Ludlam, 1968). This very dry, unstable air mass is not likely to produce any clouds or weather, but it does play a role by allowing strong mid-tropospheric winds (occasionally as high as 25 m s^{-1} [~50 knots] or more) to penetrate to low levels. With the elimination of the surface inversion, momentum from higher levels is free to be transported all the way to the surface. This high-speed flow (usually from the southwest) can generate dust storms

in the dry air, and influences the motion of the dryline, as well as the resulting moisture convergence along the boundary. The appearance of a strong "push" (high winds) at the surface in the dry air is a reliable indicator that the dryline will become active with severe thunderstorms. While thunderstorms of a non-severe character may occur on the dryline without such conditions, the dryline is not usually active unless the strong winds appear in the dry air.

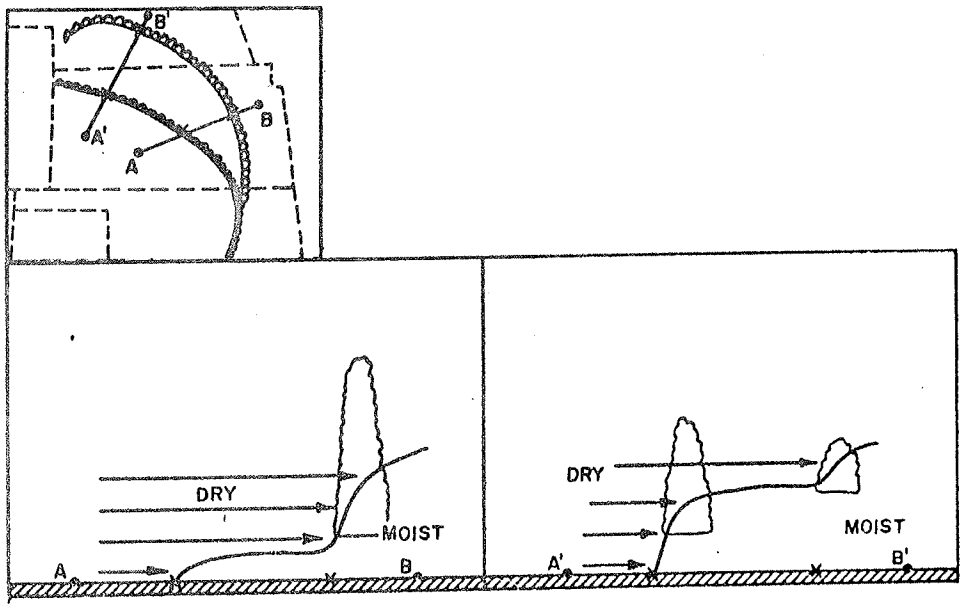


Fig. 3.13. Schematic cross sections along lines AB and A'B' showing different depths of moist layers in different locations along the dryline (after Siebers et al., 1975). Clouds are indicative of convective intensity.

Note that the structure of the dryline can be exceedingly complex, as suggested by the references. Often, the convection is most active on a second boundary, ahead of the "true" surface dryline (as revealed by the air with dewpoints less than 45°F). This may be a reflection of an upper dryline (Fig. 3.13), with shallower moisture between this boundary and the true dryline (Siebers, et al., 1975). Further, the surface dryline rarely intrudes east of the Mississippi, but Miller (1972) refers to upper level "dry prods" as a significant ingredient for severe weather throughout the country. Such features are sometimes reflected in the surface data as a subtle break in the dewpoint temperature gradients. At other times, the surface data may not reveal the "upper dryline" at all.

Much work remains to be done in illuminating the structure and behavior of drylines in conditions supporting

severe convection (see McCarthy and Koch, 1982), but the careful analyst needs to be aware of the dryline and its interaction with large-scale flows. It is a common feature of extratropical cyclones in the Great Plains, with the dryline typically intersecting the front at or near the low center, forming what is often referred to as a "triple point" (Fig. 3.14). A common mistake in analysis, as mentioned, is to put a cold front in the trough containing the dryline and to ignore the real thermal contrast to the northwest. This error can be avoided by examining the 850 mb chart to locate the air mass discontinuity, by continuity checks on the location of the cold front, and by careful examination of the wind fields at or near the surface. In the dry air, there is less tendency for a northerly component to the flow than in the cold air behind a true cold front. Further, the dewpoints in the cold air mass are often relatively high.

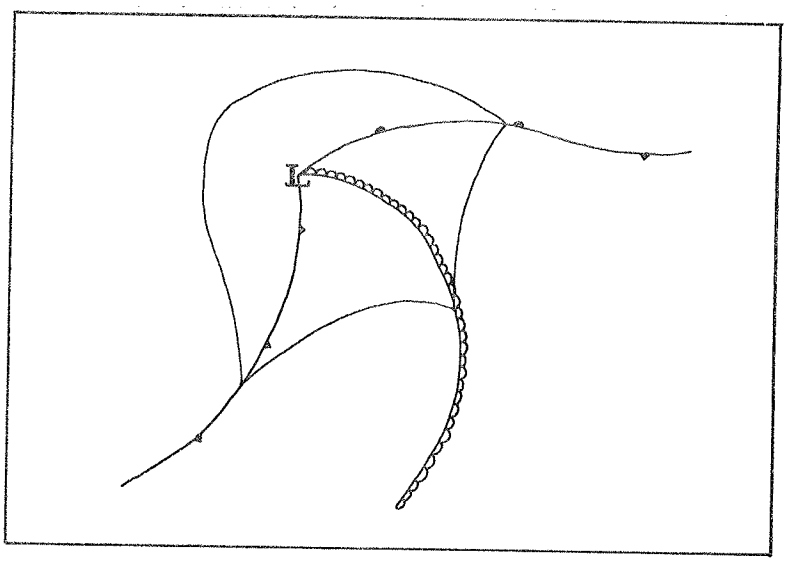


Fig. 3.14. Schematic example of triple point intersection of fronts and dryline in a developing frontal wave.

6. Land/Sea Breeze Fronts

Unlike the dryline phenomenon, the thermally driven circulation associated with the boundary between land and water is well-recognized. Also in contrast with the dryline, this feature is a true mesoscale feature, with its influence generally confined to the near-shoreline zone, say within 30 km either side of the shoreline (O'Brien and Pillsbury, 1974). The theory of such a flow has been fairly thoroughly developed (Haurwitz, 1947; Defant, 1951; Estoque, 1961; Pielke, 1973).

Generally speaking, the land/sea breeze is not a factor in most severe weather. This is a direct result of the fact that most severe weather of concern to SELS occurs well inland, away from large bodies of water. Further, the circulations involve wind speeds of only a few m s^{-1}, over limited areas, which limits the overall significance of the flow. However, the mesoanalyst needs to be cognizant of this phenomenon since there certainly have been occasions where it has been a significant influence (e.g., Fujita and Caracena, 1977; Lyons, 1966). Note that the Great Lakes and Lake Okeechobee in Florida are large inland bodies of water which show clear land/sea breeze circulations (see Neumann and Mahrer, 1975).

In physical terms, the resistance that water has to changes in temperature, compared to the adjacent land areas, is the basic factor leading to the circulation. During the warm season, a water surface typically is cooler than the land during the day and warmer at night. Allowing for a time lag in the heating and for circulation development, the surface flow reaches its peak during the morning and again in the afternoon (see e.g., Hsu, 1969 or Neuman and Mahrer, 1974).

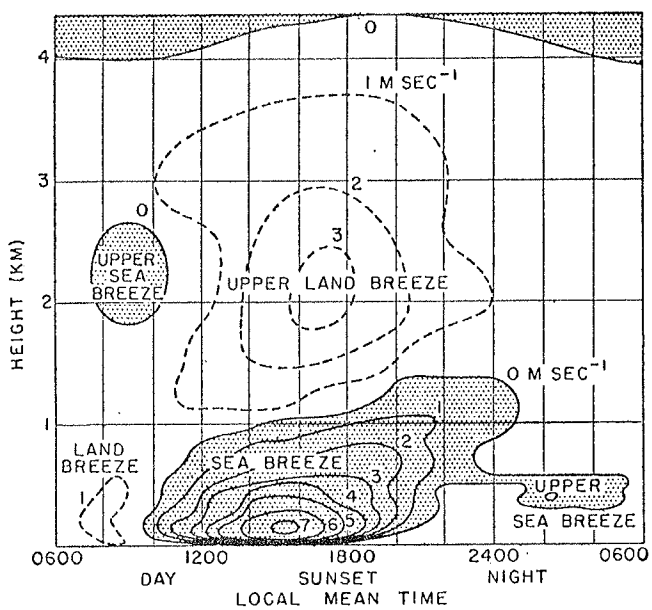


Fig. 3.15. Isotachs of land/sea breeze regime as a function of time (after van Bemmelen, 1922).

Basically, as the land surface heats during the day, it warms the air above it relative to that over water, giving the air over land a greater tendency to rise. This creates a horizontal flow off the water (the sea breeze) and

compensating subsidence above the water. The cooler air flowing off the water penetrates inland in the form of a small-scale front, with the amount of penetration depending partly on the strength of the flow, which in turn depends on the land-sea thermal contrast. This basic flow is modified by friction and Coriolis forces, resulting in the development of along-shore components to the wind (Dutton, 1976, p. 375ff). A compensating flow is required aloft to complete the circulation (Fig. 3.15), with its height and strength depending on the overall stability of the air mass. If the overall flow is capped by a persistent inversion, the compensating horizontal current aloft will be stronger. During the night, the situation is reversed, with a period of transition separating the daytime "sea breeze" from the nighttime "land breeze".

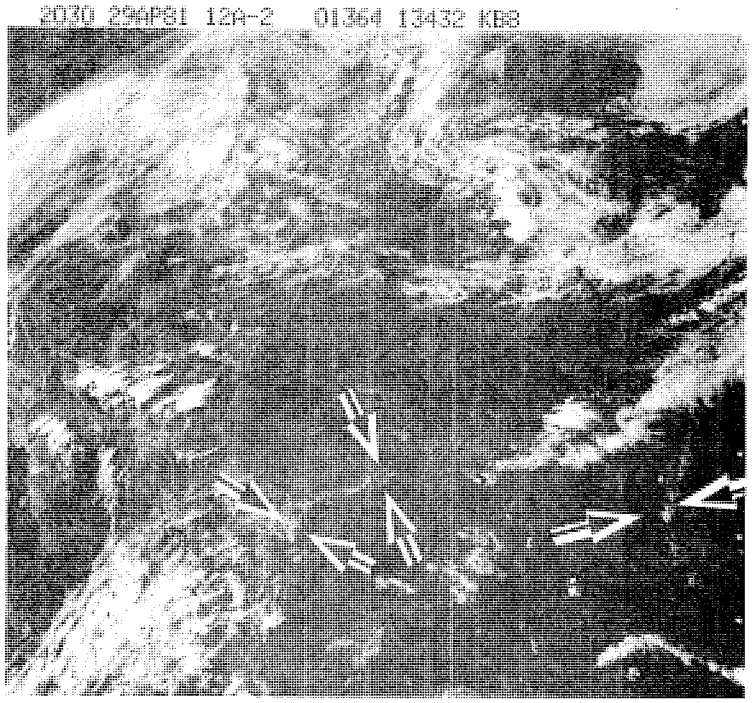


Fig. 3.16. Visible satellite image showing sea breeze-induced cloud lines along the Texas and Louisiana gulf coasts, and around the Florida peninsula, as well (arrows).

Although local non-severe thunderstorm activity often is influenced by convergence along the sea breeze front (Neumann, 1951), the analyst is often more concerned about possible interactions of other mesoscale features with the front. By itself, the sea-breeze circulation usually is not strong enough to produce severe convection. In fact, the

front itself may be masked by larger scale features which can dominate the wind and thermal effect circulation. Often, the land/sea breeze front is offset by prevailing flows, which displace the boundary downstream.

Owing to its mesoscale nature, the frontal boundary is often difficult to analyze from conventional data and the best tool for locating the sea breeze front is usually the satellite image (e.g., see Fig. 3.16). Clearly, this is a small-scale frontal boundary in the true sense of the term, so many of the clues for locating synoptic scale fronts can be used here. The analyst is cautioned not to expect the winds to blow directly off the water surface, especially late in the day when the Coriolis force has had time to develop along-shore wind components.

7. Thunderstorm Outflow Boundaries

Since thunderstorms are almost totally limited to environments with unstable thermodynamic stratification (mechanically forced convection does occur - e.g., Carbone and Serafin, 1980), one interpretation of thunderstorms is that they exist to exchange warm, moist air at low levels with cold, dry air at upper levels. While this is an oversimplification, observations generally support this basic view. Thus, the downdrafts which accompany thunderstorms bring down to the surface a mass of relatively cold air (compared with the air rising in the updraft). The greater density of this cold air forcing the downdraft also serves to keep it at lower levels, just as in the case of the advancing cold air behind a cold front. Therefore, upon reaching the surface, the cold air is forced to spread laterally, forming a small scale cold front, usually called a <u>gust front</u>. Temperature contrasts across the gust front can be in the range of $10^{o}K\ km^{-1}$.

In many ways, the mesoscale outflow boundary resembles a true cold front. There have been numerous studies of this feature (Charba, 1974; Goff, 1976; Mitchell and Hovermale, 1977) which emphasize the details of its structure. From the viewpoint of the mesoanalyst, most of these details are not adequately resolved by conventional surface data, but they can be important in trying to interpret the weather situation. Figures 3.17 and 3.18, taken from Charba (1974), show many details of the outflow's character, including the fact that the initial wind shift and the beginning of the pressure rise can be some distance ahead of the temperature break and the really strong winds. Further, the results of Goff (1976) show there is some tendency for a low-level backflow, which can create observed surface flows which

appear to be blowing through the gust front. In such cases, the cool temperatures and high pressures are the main clue to the gust front's passage.

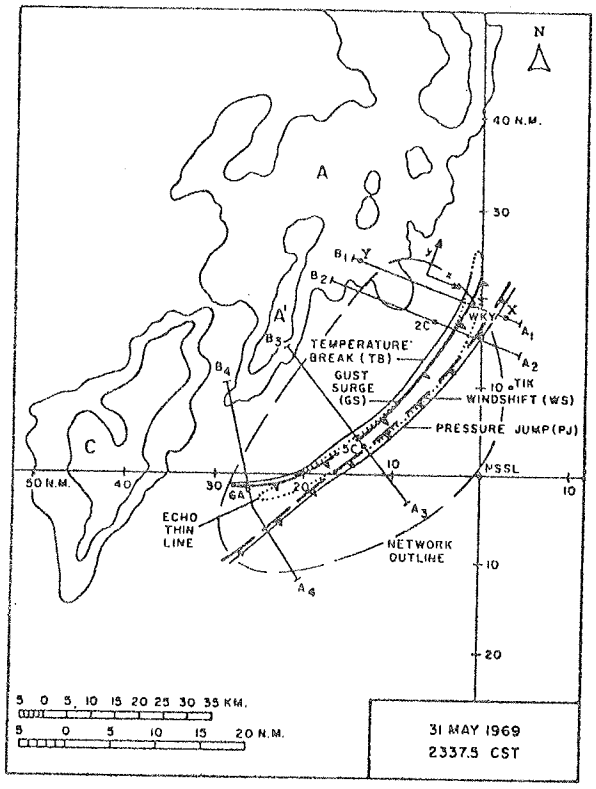


Fig. 3.17. Detailed structure of thunderstorm outflow, showing spatial relationship of outflow boundary to radar echo at a specific time (after Charba, 1974). Note the complex nature of the outflow boundary with the windshift line and pressure jump leading the temperature break and gust surge.

Obviously, the general effects associated with gust front passage at an observing site are abrupt changes in wind direction and speed, usually accompanied by rapid cooling and rising surface pressure (Fujita, 1959). A typical sequence of events observed at a station during gust front passage is shown in Fig. 3.19. Since the gust front usually is associated with a rainy downdraft (see the discussion of storm types and structure, II.III.A.5), precipitation begins after gust front passage. When the outflow has just begun, the precipitation may be quite close to the gust front. Later in the storm's life cycle, this first surge of outflow can move well out ahead of the precipitation area. Because the outflow spreads out in all directions (although predominately in the direction of storm motion), some areas may experience gust front passage and

never receive any precipitation.

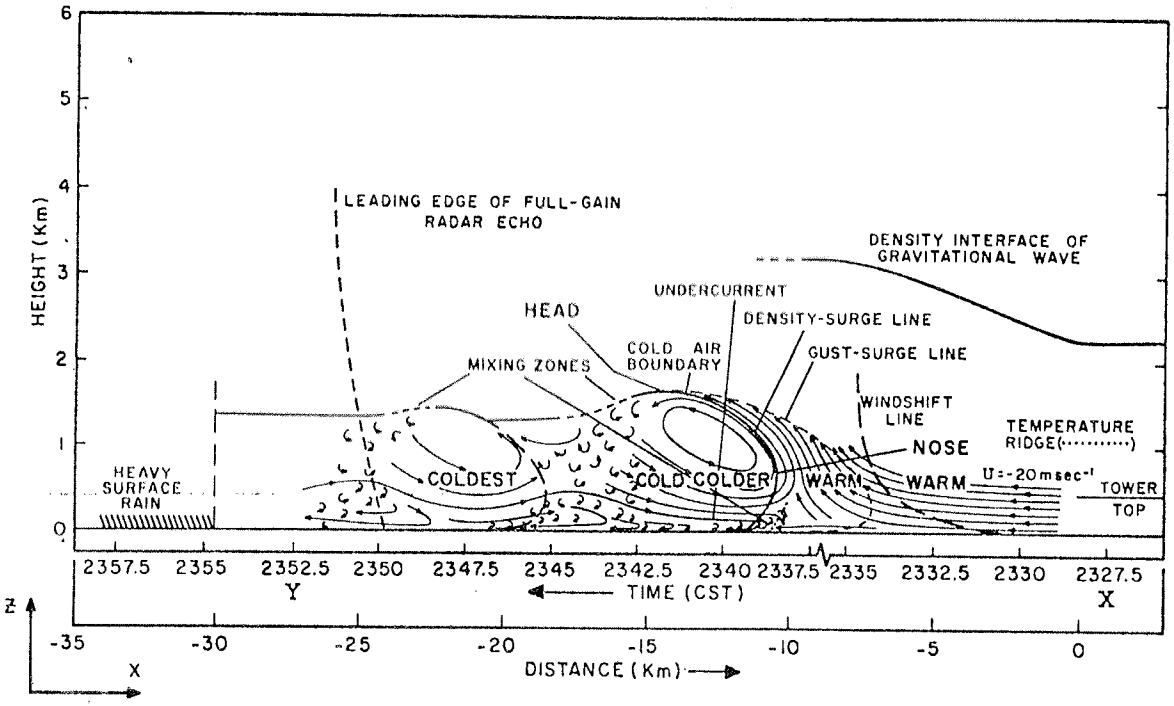


Fig. 3.18. Schematic cross section through thunderstorm outflow, showing relative depths and vertical structure of features (after Charba, 1974).

Most convective situations produce more than a single thunderstorm cell, so several surges of cold air may be produced in succession. Further, storms separated in space and time can produce separate regions of outflow which eventually merge into a larger area of basically similar rain-cooled air. With time, this produces a subsynoptic scale "bubble" (or "mesohigh"), characterized by higher pressures and cooler temperatures, easily resolved by the conventional surface network. The leading edge of this bubble is the combined gust fronts from the storms which produced it and is often analyzed as a "squall line". As pointed out by Fujita (1955), this feature can mask the true fronts in the area, making the analysis quite difficult. This is true particularly when the storms occur near a front and have not moved far enough away from it to make the distinction clear. Occasionally, the outflow boundary can dominate the true front, and the squall line effectively becomes the frontal boundary. This seems to occur most often in the case of "southward burst"-type squall lines, as described by Porter et al. (1955).

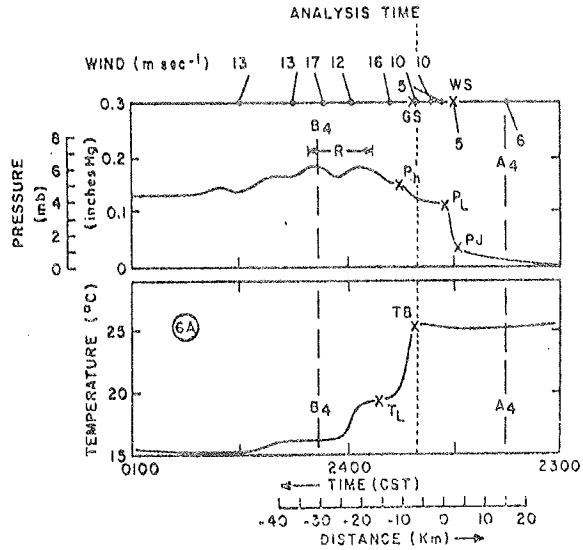


Fig. 3.19. Example of time sequence of events during gust front passage at a point (after Charba, 1974). See Fig. 3.17 for explanation of abbreviations.

Outflow boundaries are generally convergent, so new convection frequently develops along them in response. These boundaries can have convergence values of 10^{-3} s^{-1} associated with them (implying vertical motions of 1 m s^{-1} at a height of 1 km). It is important to analyze and forecast these boundaries properly. It is also of value to be able, if possible, to distinguish the gust fronts from the true fronts, since new activity can develop on the front after the first line of convection has moved away. Wind and pressure patterns behind the squall line's gust front have been well-documented by the pioneering work of Fujita (1955), Fujita and Brown (1958), Zipser (1977), and others. After the storm's passage, pressure may fall below the pre-storm value behind the mesohigh, forming what Fujita has called a "wake depression". Often the winds return to a southerly direction, the clouds break up, and temperatures and dewpoints recover. Naturally, this sequence will not occur when the true front retains its identity and passes before recovery can occur. The recovery to near pre-storm conditions following gust front passage is the primary clue which allows the analyst to distinguish a gust front from the true front.

Many of the details concerning behavior of the gust front and the flow behind it are connected intimately to the storm(s) generating the outflows. The basic configuration, shown by Fig. 3.20 is widely applicable, but some features in a given situation may differ as a result of the

interactions between storms. As described by Lemon (1976), Barnes (1978), and Lemon and Doswell (1979), the gust fronts from previous storms can have a significant role in the organization of severe "supercell" storms. In fact, Lemon and Doswell describe in some detail how a single supercell can develop two downdrafts, with the resulting gust front interaction possibly having a major role in tornadogenesis. Such a process generally is below the resolution limits of surface data. As the supercell collapses, it is a common event for the dissipating storm to become imbedded in a more or less solid line of storms, with the gust fronts merging, as already described. The details of storm type occasionally can be inferred from radar and satellite data, and this topic will be covered elsewhere (II.III.A.1).

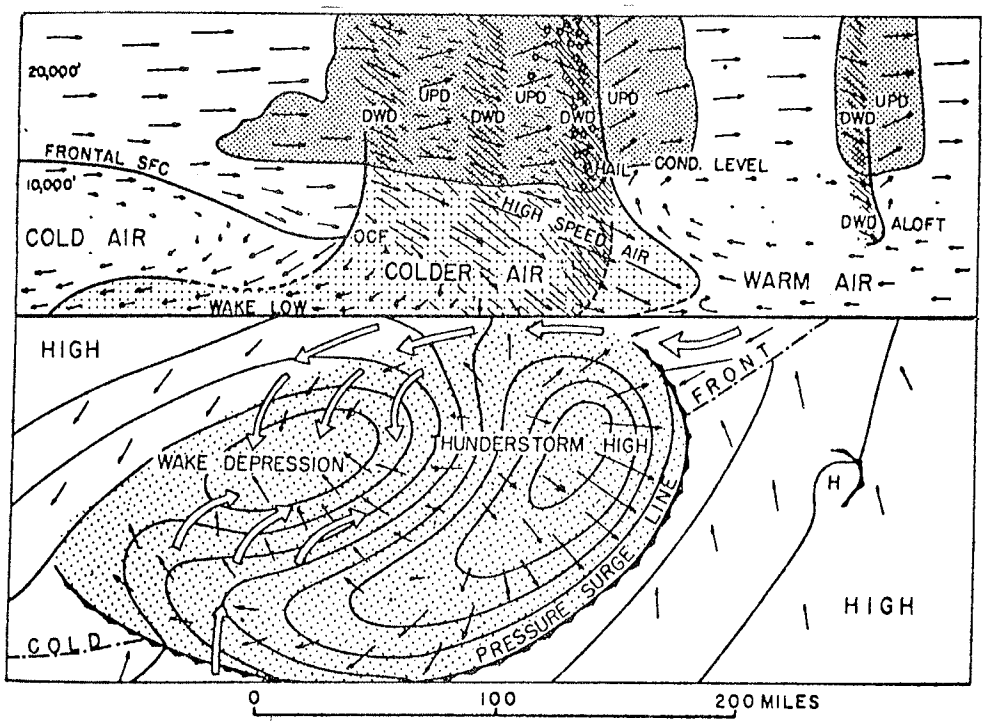


Fig. 3.20. Model of thunderstorm-induced mesoscale weather system (after Fujita, 1955). UPD denotes updraft, while DWD stands for downdraft.

Available surface data can aid the analyst in his assessment of possible future behavior of gust front convection in much the same way as it can with true fronts. That is, as described recently by Maddox et al. (1980), the surface convergence and vorticity fields associated with interacting surface boundaries have a clear relationship to subsequent activity. The analyst should monitor the situation hourly to detect important clues as to how the

interaction(s) are proceeding. Although the time-to-space conversion technique of mesoanalysis[3], used by Fujita (e.g., 1955) and Barnes (1973), among others, is not advocated in these notes, there is no question that time series observations (Maddox et al., 1980; Moller, 1979) are of great value in helping relate the observations to what is known about how convective storms behave.

A reasonable approach for the analyst is to plot the hourly observations (and any received "specials") along a time line, for selected stations. This makes a convenient reference tool when constructing spatial charts, and should not require too much time to accomplish and to update on an hourly basis. The key idea here is to compare the sequence of observations against the spatial analysis, to check on the consistency of the analysis with the analyst's view of the events.

One problem which has become widespread needs to be mentioned here. There is a tendency for the terms "instability line","gust front", and "squall line" to be used interchangeably. The appearance of thunderstorm lines on the surface (and radar) chart has received considerable attention (Newton, 1950; Fulks, 1951; House, 1959). In the author's opinion, this has resulted in an overemphasis on "squall lines". Confusion exists between what are referred to in this document as squall lines (see II.III.A.5.d) and other pre-existing, non-frontal, non-convective linear features in the analysis. Clearly, there is a marked tendency for storms to form in lines (with the spacing between storms quite variable in space and time) which is not entirely understood. The relationships, if any, among these phenomena (i.e., non-convective linear analysis features, solid lines of radar ehoes, thunderstorms in lines, connected gust fronts, etc.) is not at all obvious. Considerable research remains to be done and some of the present confusion is apparently related to a failure to distinguish between phenomena of different scales.

Intersecting solid lines of radar echoes are not a common event. If one can assume a rough equivalence between instability lines and squall lines, one might suspect that the so-called line echo wave pattern or "LEWP" (see Nolen, 1959; and II.III.A.5.c.ii for more details) configuration of radar echoes is related to this "intersecting lines" phenomenon. However, recent studies (Fujita, 1978) suggest that the LEWP and the so-called "Bow Echo" are basically the same phenomenon and are the result of downdraft accelerations, not circulation around a mesolow. This is not to say that mesolows are not associated with LEWPs. We

shall return to mesolows later. However, the subsynoptic scale aspects of the mesolow phenomenon may have been overemphasized in the recent past, since the picture described by Magor (1959; his Fig. 1) is not particularly common, as sensed by conventional surface data.

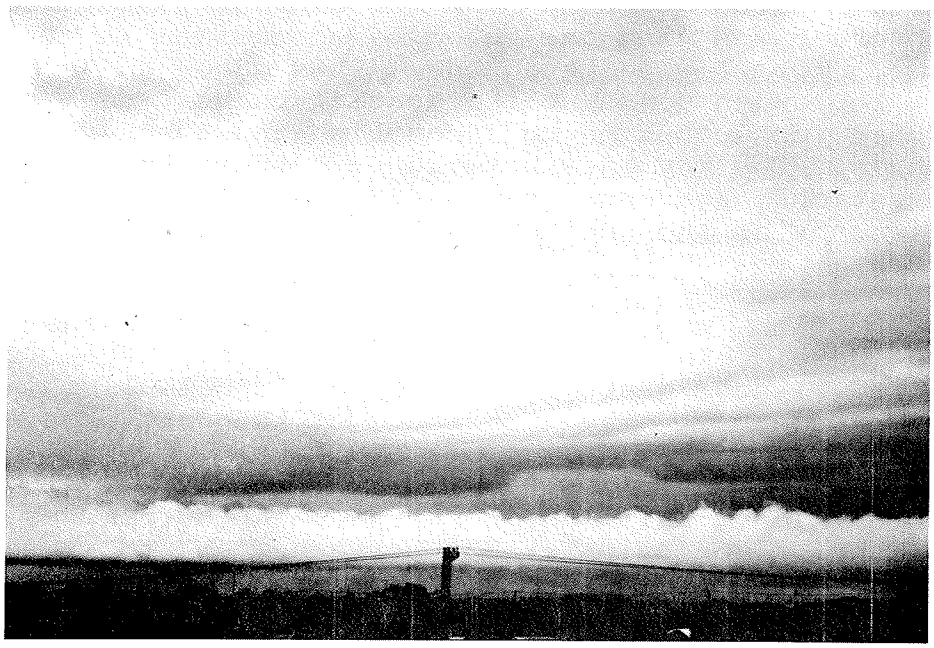


Fig. 3.21. Example of a roll-type outflow cloud, the detached, horizontal tube-like cloud near the bottom of the photograph. Occasionally, this tube may appear to be rotating slowly about a horizontal axis, with the forward edge rising and the rear side sinking.

Remarks on surface observations often provide useful information and one facet of these remarks has a direct bearing on the mesoanalysis. The leading edge of a cold outflow is often marked by a "shelf" or "roll" cloud, the latter term appearing most frequently in additive remarks. Actual roll-type clouds look like Fig. 3.21, while the more common shelf cloud appears in Fig. 3.22. The presence of either type of outflow cloud is usually a clear indication of a gust front [or on rare occasions, a true cold front (Livingston, 1972)]. When remarks include the observation of such a cloud, the location of the gust front can be made quite accurately, at least in the vicinity of the observation. If the radar echoes (precipitation) are quite distant from the shelf cloud observation, then it is clear that the outflow boundary has moved well away from the downdraft which initiated it.

In the area of thunderstorm outflow boundary identification and tracking, satellite images make an important contribution to mesoanalysis. It cannot be overemphasized that concurrent satellite images (preferably in the form of animated loops) need to be integrated thoroughly into the process of mesoanalysis. It is by this means that the positioning of surface boundaries can be refined and clarified. At the same time, the significance of cloud features can be assessed by comparison with the surface (and upper air) observations.

Fig. 3.22. Example of a shelf-type outflow cloud, in this case showing a somewhat terraced appearance. If motions can be seen, they will be upward along the upper portion of the wedge-shaped shelf cloud. Note that the shelf is attached to cloud base above it.

The stabilizing influence of the outflow air can be seen readily in images of "air mass"-type thunderstorms (i.e., the small, short-lived, isolated convective cell). The air which has been cycled through the storm is typically cool, subsiding, and stably stratified. This suppresses clouds in the wake of the storm, forming a cloud-free area (Fig. 3.23), perhaps surmounted by an anvil remnant. It is common that such a feature is too small to be resolved by the surface network and yet it may have an influence on a local weather forecast.

The outflow boundaries from more significant convection may live on after moving away from the storm. This typically takes the form of the "arc" cloud, which Purdom (1973) has described in detail. The arc cloud usually coincides with the gust front and results from cumulus and towering cumulus which concentrate in the zone of convergence along the boundary. On some occasions, the arc may be associated with a pressure jump line (Shreffler and Binkowski, 1981).

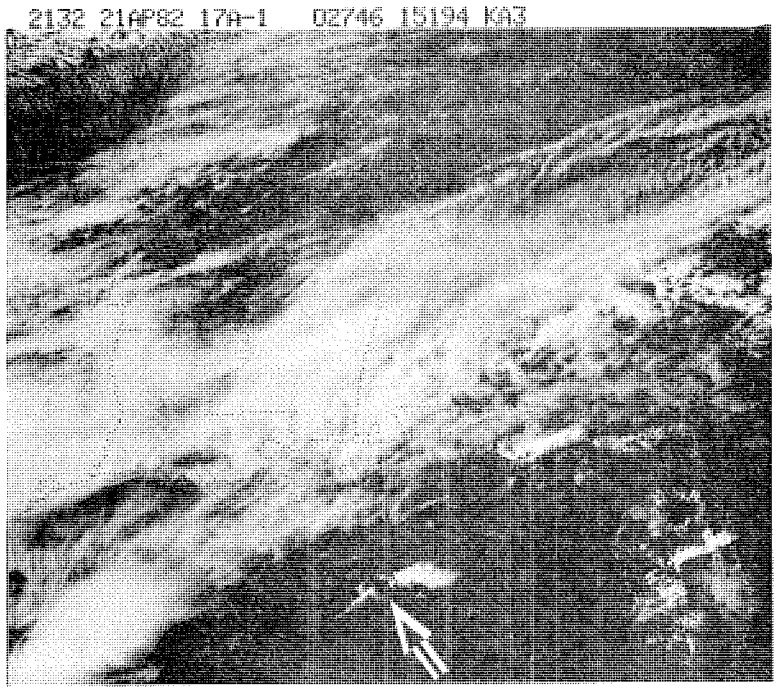


Fig. 3.23. Visible satellite image showing a small, nearly circular outflow boundary in the Gulf of Mexico (arrow). Note the ring of enhanced cumulus surrounding the area of nearly cloud free air. Remnants of the thunderstorm anvil can be seen on the northeast side of the ring.

There are two aspects of arc clouds which need clarification during the act of mesoanalysis. The first question is which, if any, arc clouds are likely to become reactivitated during their lifetimes? Since gust fronts (and their associated arc clouds) occur with virtually all significant (i.e., those producing downdrafts) convective storms, while only some persist and re-develop thunderstorms, this is an important problem. The second question is where are new developments likely to occur, given the observation that storms do not always develop uniformly along the outflow boundary? Purdom (1973, 1979) has made considerable progress in helping to resolve the second question. He has indicated that wherever outflow boundaries intersect other

surface boundaries (including fronts, sea breeze fronts, other outflow boundaries, etc.), these are preferred locations for development. This observation is an obvious argument for the value of a careful hourly mesoanalysis of conventional data, since surface features are not always well-delineated by satellite imagery alone. Purdom's most recent studies also suggest that the cloud types and amounts ahead of the arc cloud can be useful in a qualitative assessment of the stability of the air mass which will be influenced by the arc. He asserts that arc clouds propagating into clear skies are unlikely to produce strong convection. These ideas are not yet developed fully, since exceptions occur and their day-to-day value remains to be demonstrated. Nevertheless, they suggest ways in which the satellite data can be integrated with the conventional, and represent a rapidly evolving line of research.

When storms mature, their anvils frequently obscure any low-level features like gust fronts. It is during this phase of storm development that radar is a natural tool for supplementation of conventional observations. There are three types of radar data that may be available: onsite, remote, and facsimile (or AFOS) displays. Onsite radar can be controlled, within its operational guidelines, to give the best possible depiction of the storm configurations for purposes of the mesoanalysis. An identification of storm types (see II.III.A.5,6) can be valuable, but a broad-scale picture of the precipitation distribution is most valuable for mesoanalysis. That is, the analyst needs to know the location of any thunderstorm lines, isolated storm cells, radar fine lines (which often reveal fronts and/or gust fronts), and the heaviest precipitation cores. These provide details which may be important when the storm anvils cover the area of interest and, of course, at other times as well. "Spearhead" or "bow" echo configurations (Fujita, 1978) may reveal accelerating gust fronts, between conventional observation sites and times. Although radar data are considered most important for warning decisions and discussed in that context in III.IV.8, they cannot be ignored by the mesoanalyst.

When radar is not located at the station where the analysis is done, the alternatives are remote displays of one (or more) radar(s) in the area, and the facsimile (or AFOS) display of the national radar data. These displays are not as flexible, but can still provide enough information to get a rough idea of the precipitation distribution. Although storm type identification is more difficult under these circumstances, enough information can often be obtained to be of help to the analyst (Wilson and Kessler, 1963).

As shown clearly by some of the earlier studies in mesoanalysis (e.g., Fujita and Brown, 1958), after the mesosystem has evolved into a mature storm complex, the area influenced by outflow can be dramatically larger (perhaps 20 to 100 times) than the area actually covered by precipitation echoes (see Fig. 3.24) at any given moment. However, as also seen in their study, the area eventually receiving precipitation is usually a substantial fraction of the outflow "bubble" (Fig. 3.25).

One can readily see that outflow boundaries tend to have a reasonable continuity from hour to hour. When conventional data are meshed with radar and satellite information, the analysis and prognosis of the boundaries are relatively straightforward, provided the analyst is aware of the basic types of structures seen.

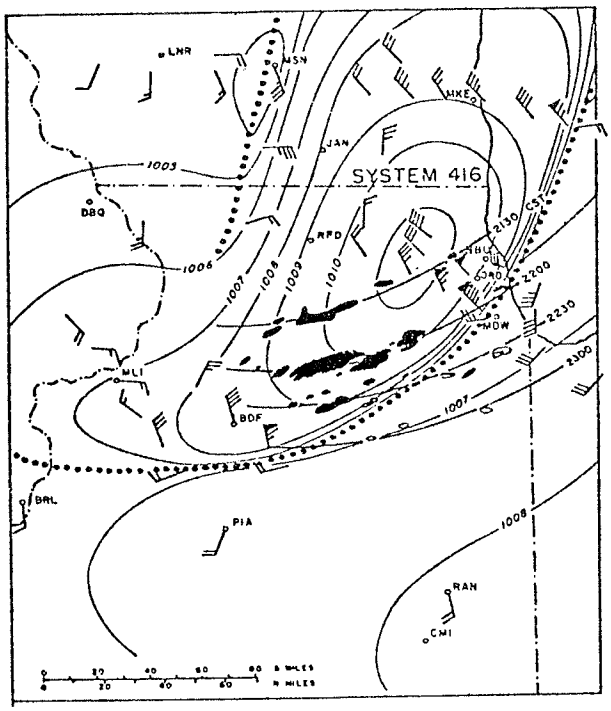


Fig. 3.24. Relationship between radar echoes (black areas) to overall mesosystem produced by the storms associated with the echoes (after Fujita and Brown, 1958).

As a final word on outflow boundaries, the reader needs to be aware of the flash flood potential in situations involving this phenomenon (also see III.F and II.III.C). Maddox et al. (1979) have noted that many of the flash floods which they studied (34%) were associated with storms developing along an outflow boundary. Therefore, the

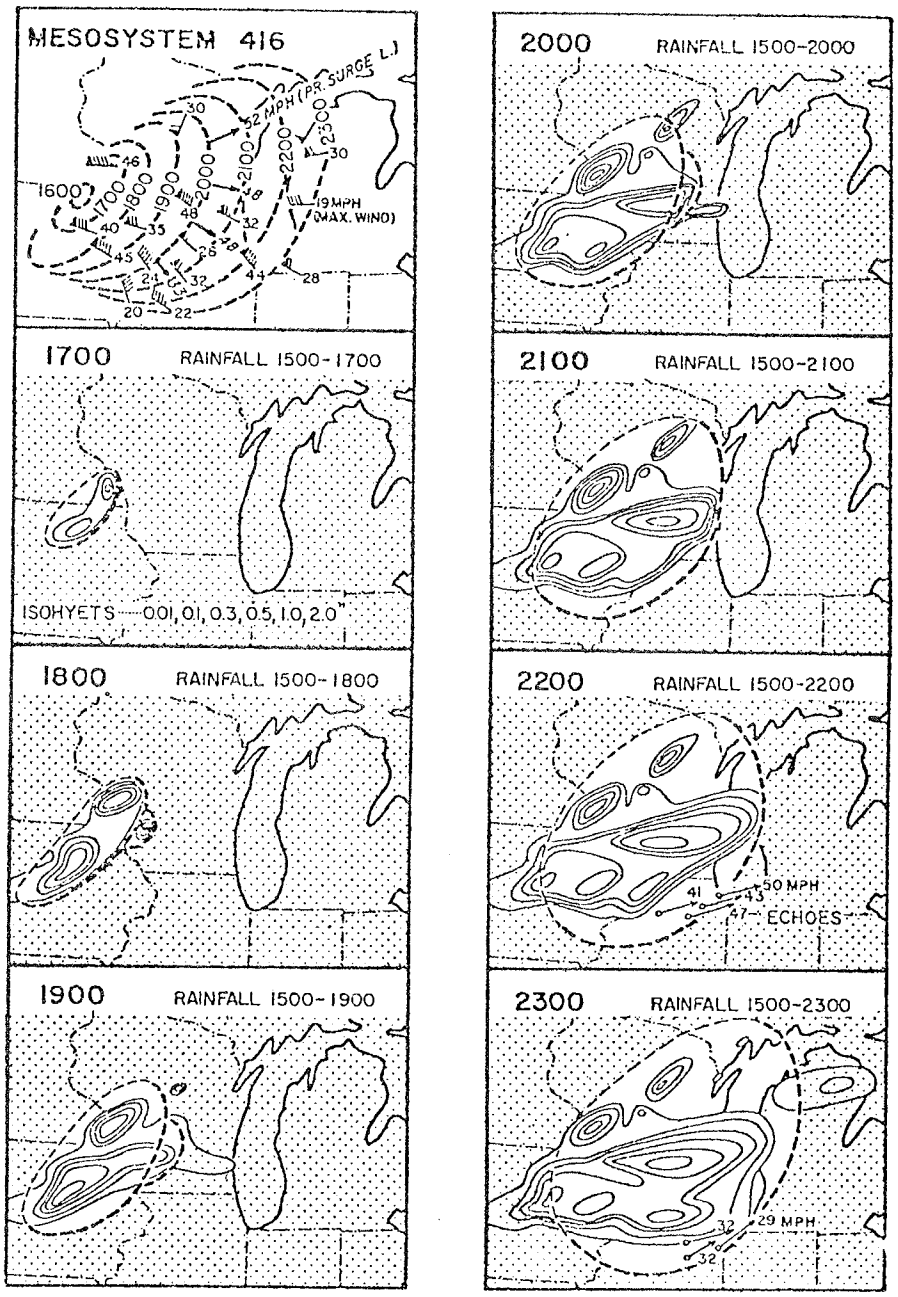


Fig. 3.25. Time averaged precipitation produced in several periods during the life of the mesosystem in Fig. 3.24 (after Fujita and Brown, 1958). Reported wind gusts are plotted in the upper left.

analyst needs to be concerned with factors relating to flash flood potential. A significant factor in the frequent occurrence of flash floods during the nighttime hours is the low-level jet phenomenon. Although the low-level jet will be discussed more fully later, the analyst should note the

axis of the low level flow and where the speed maxima are along that axis. Any situation wherein strong low-level flow impinges on an outflow boundary has flash flood potential, when the undisturbed low-level flow is moist and unstable.

C. Boundaries Not Involving Air Masses

1. Wind Shift Lines

The careful analysis of surface data frequently reveals organized windshift lines which are apparently non-frontal in character. Since, by definition, no clear change of air mass is involved, it is often difficult to explain or understand their origins. Nevertheless, they do occur and, since they may give rise to surface convergence, they can be involved in severe weather events. At times, the wind shift may be traceable to an old outflow boundary from thunderstorms occurring, say, the previous day. Modification of the outflow air may have erased any apparent temperature and dewpoint differences. The same thing can happen with old quasistationary fronts.

A commonly observed non-frontal wind shift line is one which occurs ahead of a cold front. This situation can make the analysis difficult since fronts themselves often are analyzed (mistakenly) along the wind shift line. This wind shift may be the focus for development of the pre-frontal "squall lines". At other times, the front itself is the main source of convective development (see III.B.1). Analysts should always be alert to the formation of these pre-frontal wind shifts. While the mechanism for such a phenomenon is not yet completely understood [frequently-mentioned candidates are the "pressure jump" proposed by Tepper (1950) and described by Shreffler and Binkowski (1981), and Clarke et al. (1981); and the gravity wave (Uccellini, 1975).], analysts need to be aware of this phenomenon, and take care to distinguish the true front from this pre-frontal wind shift line. Note that if the wind shift is not present until after storms have begun, one might suspect in such a case that the wind shift is merely the gust front.

Since the nature of windshift lines tends to be somewhat obscure, it is difficult to generalize about them. Certainly, the analyst needs to monitor them during an hourly analysis routine, in order to check on their continuity. If they show some tendency to dissipate during the diurnal cycle, the suggestion is that the windshift is related to topography in some fashion, which indicates that

such a windshift is probably not significant. If, on the other hand, the windshift is located in a clearly defined pressure trough, even though no obvious temperature/dewpoint differences are present, the indication is that this may be an important surface boundary.

2. Pressure Troughs

Non-frontal pressure troughs are a common phenomenon. For reasons discussed in III.D.2, they can have an important influence on winds. Perhaps the most commonly observed non-frontal pressure trough is that which develops in the lee of the Rocky Mountains, usually in the wake of a large polar anticyclone. While there are many textbook discussions of "Lee Side Trough" development, (see Panofsky, 1964, p. 118, for a simple summary; Palmen and Newton, 1969, p. 344ff, for an excellent detailed treatment) these need not concern the analyst.

Basically, trough development requires at least moderate westerly components across the mountains. By the creation of a lee side region of lower pressure, a southerly wind regime is established over the Plains which acts to return moisture and warmer temperatures driven out by prior anticyclone passage. This process is a necessary precursor to the establishment of drylines (see III.B.5) and to "Lee Cyclogenesis". The problem of lee cyclogenesis per se is beyond the scope of this report and the reader is encouraged to examine the numerous references on the topic (e.g., Bleck, 1977; Tibaldi et al., 1980; Chung et al., 1976; Kasahara, 1966; Hage, 1961; Klein, 1957; and Bolin, 1950).

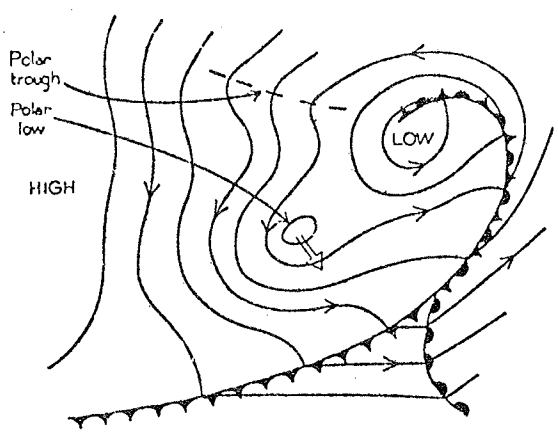


Fig. 3.26. Schematic diagram of polar low and polar trough.

Another common type of non-frontal pressure trough appears in the polar airstream behind a cold front, when the

parent cyclone is well into the process of occlusion. These troughs are apparently similar to what are sometimes known as "polar lows" (see Reed, 1979; Harold and Browning, 1969; Rasmussen, 1977). Several can appear in the northwesterly flow behind the low pressure center (Fig. 3.26), and may be supported by windshifts as well. It is common that small-scale "comma cloud" formations are associated with these secondary pressure troughs, and they can be significant convective weather producers. As described earlier (see III.B.4 above), when a pool of cold air aloft is situated near the center of the occluded system and sufficient heating at the surface occurs, these secondary troughs can result in severe thunderstorm activity (including tornadoes, at times). The co-existence of secondary troughs accompanied by clouds (associated with upward motion) indicates that significant deformation (see Doswell, 1982a,b) and frontogenesis may be occurring. Thus, an initially non-frontal trough may develop into a secondary frontal system of significant proportions. In fact, the work of Johns (1982a,b) suggests that many of the northwest flow severe weather cases have origins in these secondary troughs. Such developments can occur repeatedly as the larger, overall occluded cyclone gradually decays.

D. Pressure Change Analysis

1. Applications to Synoptic Analysis

It is not an exaggeration to say that <u>an analysis of the changes in surface pressure is probably more valuable than the pressure pattern itself</u>. At SELS, automated plots of pressure changes over several different periods are available and analyzed routinely. Synoptic analysis makes most use of the 3- and 12-h changes, with which it is possible to obtain a pretty clear picture of the movement of the large-scale pressure systems. The relative strengths, speed and direction of movement, and the tendency for deepening or filling can also be evaluated. Naturally, some of the diurnal and semidiurnal variations are included in the observed pressure changes, but these are easily accounted for, at least in principle.

By having a history of the 12-h changes, at 6-h intervals, the broad pattern of synoptic-scale features can be easily grasped. Further, significant large-scale events often are seen first by rapid changes in the isallobaric fields. While the theory involved in pressure changes is more than adequately covered by textbooks (e.g., Saucier, 1955; Hess, 1959 [p. 219ff]; Petterssen, 1956a [Ch. 19]), some simple concepts are useful to review.

A well-known equation, logically enough known as the pressure tendency equation (see Panofsky, 1964, p. 124ff), governs the change in pressure at any fixed point. Since pressure is simply the weight of the atmosphere per unit area above the point, it is obvious that pressure changes reflect the vertically averaged effect of processes occurring aloft. The tendency equation states that the change of pressure at a point is the combined result of: the vertically averaged horizontal divergence, the vertically averaged horizontal advection of mass, and the vertical motion (above the level of interest).

It is unfortunate that our ability to measure all three of these influences is not adequate for a direct applicaton of the tendency equation. Since surface pressure change is the result of small differences among these three, poorly measured variables, which usually act partially to compensate for one another, the tendency equation is only of academic (or instructional) value.

However, the atmosphere has no difficulty solving the equation for itself and we see the result of imbalances among the three effects as pressure change patterns which can reveal features of interest to the analyst. For example, turning our attention to the 3-h pressure changes, the location of fronts may be clarified by the isallobaric pattern. The large temperature changes associated with frontal passage may not be clearly revealed at the surface (e.g., as in mountainous terrain) because of local or topographic influences, but the pressure tendencies are less likely to be effected because they reflect changes through a deeper layer.

2. The Isallobaric Acceleration

Just as the pressure field can be used, via the tendency for winds to be in near-geostrophic balance, to infer the basic wind flow pattern, the pattern of isallobars can be a valuable indicator of accelerations in the wind field. At times, the contribution of isallobaric accelerations is the major contribution to departures from geostrophic balance (Brunt and Douglas, 1928 as referenced by Saucier, 1955 [p. 242]). Be aware, however, that this is not always the case, and other contributions to the ageostrophic wind can be important. See the discussions in Saucier (1955, p. 240ff) or in Petterssen (1956a, Ch. 4). As Saucier states, "... it is improper to attribute existing ageostrophic winds only to the isallobaric pattern...."

It is important to remember that the isallobaric "wind" is not a real wind (the geostrophic wind is not a real wind either!) - it is part of the total acceleration which acts on air parcels (Hess, 1959, p. 225ff). Specifically, it is that part created by local changes in the pressure field. Like any acceleration, it changes the velocity, but those changes do not take immediate affect. Rather, they operate over time to produce ageostrophic winds. If isallobaric accelerations are the only effect producing ageostrophic winds, the ageostrophic component will oscillate (as shown by Hess) about a mean value which is the isallobaric wind, with a period of about 17 h (see III.F.3, below). It takes several hours for this oscillation to develop and stabilize.

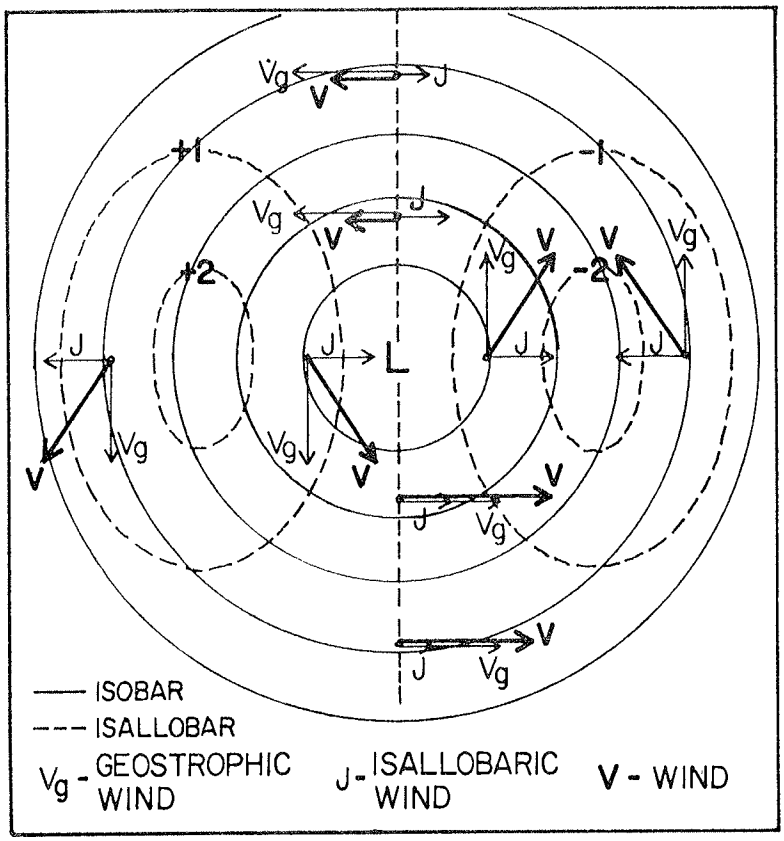


Fig. 3.27. Illustration of isallobaric wind effect for a translating low pressure center (adapted from Petterssen, 1956a).

What all this means is that parcels moving through a region of pressure changes must be acted upon for a period of time sufficient to produce a significant velocity change. This can occur when parcels are moving (at least initially)

slowly. Alternatively, the region of pressure change can be large enough (in the direction of parcel motion) that a fast-moving parcel remains under its influence for extended periods. This clearly implies a scale dependence of the isallobaric contribution to parcel velocity. Very small-scale, brief isallobaric patterns do not imply immediate ageostrophic contributions equal to the calculated isallobaric "wind".

The acceleration induced by a changing pressure field increases with the gradient of the isallobars, just as the geostrophic wind increases with the pressure gradient. For example, an isallobaric gradient of 1 mb per 3 hr over a distance of 100 km (a large value) yields an isallobaric "wind" of about 10 m s^{-1}. Unlike the geostrophic wind, the isallobaric wind is directed perpendicular to the isallobars (the geostrophic wind being directed parallel to the isobars, of course), and toward falling tendencies. This is illustrated in Fig. 3.27, showing how the isallobaric acceleration acts to turn the geostrophic wind toward pressure fall centers.

The results of isallobaric accelerations can easily be seen in the analysis process. One of the reasons for the strong ageostrophic flow nearly perpendicular to the isobars behind a cold front is the large contribution from the pressure tendencies. It is easy to visualize the effect, since it is quite clear intuitively that flow should tend to rush into an area of falling pressure and out of a region of rising pressure.

An important contribution of the isallobaric acceleration to severe weather situations is backing of the winds ahead of an approaching trough (or into a deepening stationary trough, like the "Lee Side Trough"). This backing of the flow can be seen aloft, as well as at the surface, and may serve to enhance low-level moisture influx and convergence. Note that this application of the isallobaric wind concept is on a relatively large scale, so that it is reasonable to expect isallobaric accelerations to produce ageostrophic winds which resemble the isallobaric wind.

3. Mesoscale Isallobaric Analysis

For purposes of mesoanalysis (as we have used the term), the short term pressure changes are of maximum interest. In what follows, it should become clear that their main value for mesoscale analysis is to detect mesoscale pressure systems. While these pressure systems seem to be associated frequently with severe thunderstorms,

they are not completely understood. As suggested above, the analyst should not attempt to interpret the isallobaric analysis in terms of the isallobaric "wind", on this time and space scale.

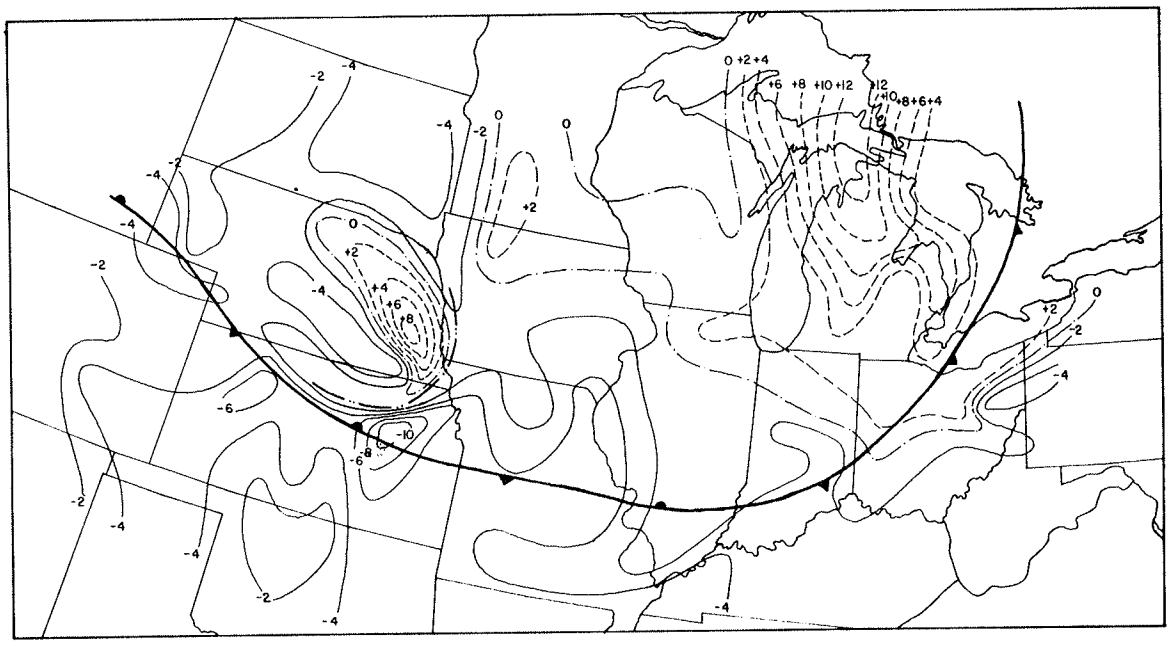


Fig. 3.28. Isallobaric analysis (hundredths of an inch of Hg, per 2 hr), at 1800 GMT, 31 May 1980.

Two-hourly pressure changes are routinely analyzed at SELS and it should be clear that in order to make full use of these change fields, they need to be produced and analyzed hourly during those periods of greatest threat of severe convection. Owing to its greater spatial coverage, one should use the altimeter setting (recall III.A above) for this purpose. An appropriate contouring interval for isallobars is somewhat dependent on the field's extreme values, but a generally acceptable pattern can be obtained with a 0.02 inches of mercury (about 0.68 mb) per 2 h interval. As mentioned several times previously, diurnal and semidiurnal effects are present.

Assuming the isallobaric analysis is done, the analyst will likely have a complex picture to try to sort out. Some observations will be missing, some reported changes will represent errors, and some will represent real atmospheric phenomena. Two examples showing 2-h altimeter setting changes are presented in Figs. 3.28 and 3.29. The example of Fig. 3.28 is one which provides an insight into the large-scale changes, as well as the smaller scale. Note the

broad area of rises, behind the front in the Great Lakes. Most obvious, perhaps, is the small-scale system in southeastern Nebraska and northeastern Kansas. Also, there is an area of concentrated falls in Kansas ahead of the rises associated with a "bubble" high in Nebraska. Considerable severe weather occurred in Nebraska and Kansas in association with this mesosystem.

Lest one be led to assume that detectable 2-h pressure changes always accompany severe weather, consider Fig. 3.29. This is a good example of how substantial severe weather can occur (in this case in central Oklahoma and southern Kansas) without any clear-cut 2-h pressure change features. Moller (1979) has examined 1-h altimeter changes for this case and has been able to detect very small-scale, fast moving pressure change couplets associated with the Oklahoma severe weather. If one uses 2-h changes routinely, it may not be possible to detect such small systems.

Of the real phenomena, part of the pattern should reflect the large-scale pattern seen in the three- and twelve-hourly pressure change analysis. Embedded in this broad scale field occasionally will be non-convective small scale phenomena which typically do not have the strength of convective isallobaric patterns. These may be real atmospheric phenomena (gravity waves come to mind as an example), but they generally do not represent a significant factor in mesoanalysis. Although this suggests that the analyst could safely ignore these, there exist examples (Uccellini, 1975; Eom, 1975), where it has been argued that they had an exceedingly important role in the convection. The key concept in separating the wheat from the chaff here is continuity in space and time. Very small-scale, transient phenomena are indistinguishable from noise, even if their origins are in real atmospheric processes. Only those which affect several observation points in space and/or time are likely to have some influence on mesoscale weather.

Further, the appearance of intermediate-scale pressure change structures can be a valuable check on the importance of satellite imagery. If a more or less shapeless cloud mass (as opposed to a "comma" cloud) is accompanied by a localized region of concentrated surface pressure falls, there is obviously reason to believe that it is linked to some dynamic "feature". Should this travelling feature be moving toward an area of convective instability, this corroboration of the significance of the cloud mass is of great value to the forecaster/analyst. Recall the discussion in III.D.5 about the effects of pressure falls on the wind field.

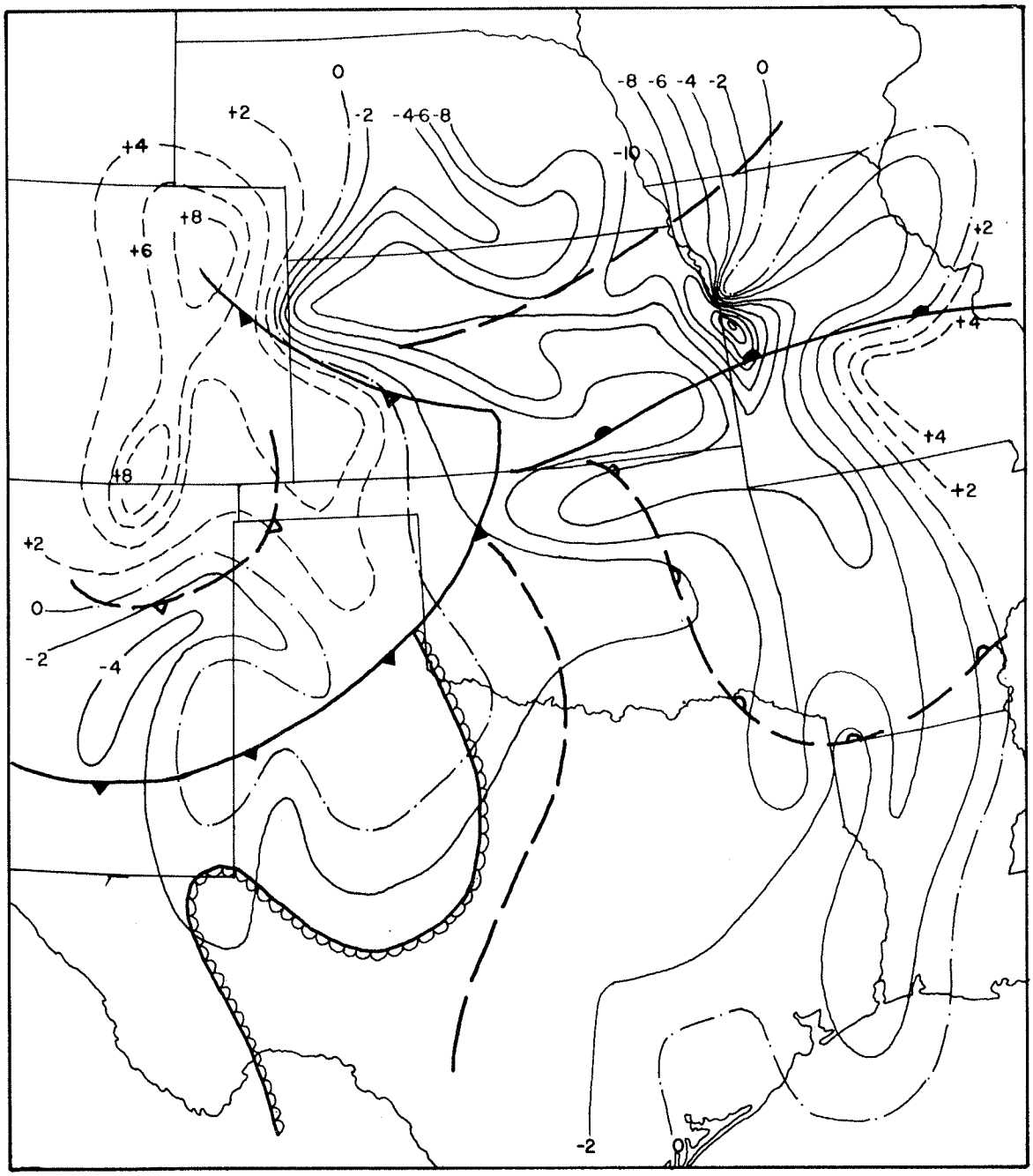


Fig. 3.29. As in Fig. 3.28 except at 2100 GMT, 8 June 1974.

Many of the high-amplitude mesoscale pressure changes associated with convection manifest themselves only after convection has developed. As one might expect, an indication of the development of a substantial

downdraft/outflow is rapidly rising pressures. This pressure rise is initially confined to a small area and is usually only observed at one station, if at all. With time, the area covered by substantial pressure rises expands, moving with the convection. Naturally, as this "bubble" moves, pressures in its wake may then begin to fall rapidly. Thus, we have a rise/fall couplet developed as a result of the outflow, with the fall trailing the rises, just as the wake low follows the mesohigh. This is common even with relatively non-severe storms.

In the severe weather situation, there usually exists a zone of concentrated pressure falls ahead of the mesohigh-associated rises. It is not entirely clear that this is indicative of an undetected mesolow. Hoxit et al. (1976) have proposed a mechanism wherein the mesoscale vertical motion fields induced by the convection (far away from the convective drafts themselves) can lead to falling pressures ahead of the storms. Whether or not this is a valid suggestion in the majority of situations remains to be demonstrated, but it does allow for the creation of a pressure fall center without requiring a travelling mesolow to be its source. Moller (1979) has observed mesoscale pressure falls, which he found to precede most severe thunderstorm outbreaks in the Southern Plains. Magor (1971) has also noted the influence of isallobaric accelerations associated with such pressure falls, which act to advect heat and moisture into area ahead of the advancing convection.

Regardless of the origins of the falling pressures ahead of the area of rises associated with outflow, the mesoanalyst should be alert for the fall/rise couplet. If a wake low is trailing the bubble high, a fall/rise/fall triplet may result. Such a feature can also evolve into a double couplet structure, should new convection develop in the wake of the first.

Although the actual mesolow has received little attention in these notes, there is no doubt that such phenomena occur. The reader should be careful to distinguish between the "mesocyclone" (and accompanying low pressure area), which is directly related to the convection (see II.III.A.5.b.(2)), and a pre-existing (or at least larger-scale) mesolow which apparently acts to enhance convection but is not directly tied to the storm's draft. Once we reach the scale of the mesolow (which is from several tens of km to a few hundreds of km in diameter), we have moved into an area of relatively little understanding. It has been suggested that at least some subsynoptic circulations have their origins in weak frontal disturbances

(Doswell, 1976). These systems are at the upper end of the size range for what loosely can be grouped together under the heading of mesolows. As we decrease the size of the mesolow, correspondingly less is known about their structure and origins. As Magor's work suggests, knowledge of mesolow existence is based at times solely on localized pressure falls. It is logical to suggest mesolow presence at intersecting boundaries, each of which generally lies in a trough of lower pressure. This is enhanced by the already described tendency for severe convection to occur in the vicinity of intersecting boundaries. However, Magor (1959) readily admits that "the meso-low [at the intersection of instability lines] would be suspect rather than directly observed."

Certainly the literature includes some well-defined examples of mesolows, which have been detected when they fortuitously passed through a data-rich region (Brooks, 1949; Mogil, 1975; Hales, 1980; Magor, 1958). From this one might be tempted to suggest their existence even when data are not available to define them. However, such inferences are essentially unscientific. In a sense, it is irrelevant whether or not one actually has a mesolow present in a given situation. The mesolow is most important insofar as it can enhance the potential for severe weather. It is probable (but not yet proved) that it does so primarily by means of its ageostrophic accelerations of the flow. The isallobaric contribution to ageostrophic accelerations can frequently be diagnosed from available data. It should be recalled that the isallobaric contribution is not necessarily the most important one. Naturally, when a mesolow can be detected reliably, it is probably a significant feature.

Finally, the short time scales of most subsynoptic pressure systems do not usually allow the winds time to react appreciably to Coriolis acceleration. Thus, one should anticipate that the wind flows in mesoscale pressure systems will not often be in near-geostrophic balance. Further, it is not uncommon for the circulation center of subsynoptic cyclones to be removed some distance from the center of subsynoptic low pressure systems. The analyst should constantly be aware of the distinction between the actual and geostrophic flow, since departures from geostrophy are related to flow field accelerations (see IV.B).

E. Thermal Analysis

It has long been recognized that severe thunderstorms, and especially tornadic storms, do not occur randomly within the warm, moist air in a large-scale extra-

tropical cyclone's warm sector. Rather, there are moisture and temperature patterns which have proven to be reliable as indicators and locators of severe storm potential. For example, when a thermal ridge axis is present upstream (generally, to the west) from the axis of maximum moisture (dewpoint temperature), a favorable configuration is present. Convection usually begins on the west side of the moist axis, between the thermal and moisture ridges. This is probably the result of upward motion (owing to warm advection) encountering abundant moisture. An example of this configuration is shown in Fig. 3.30. This is generally valid both for the lower levels in upper-air analyses, say, at 850 mb (Miller, 1972) and also at the surface (Moller, 1979). Considerable attention has been focused on the surface thermal ridge in the past (Kuhn et al., 1957; Darkow et al., 1958; Whiting, 1957), but recently this awareness has waned somewhat (probably inappropriately).

The general picture of a thermal axis upstream from the moisture axis in a severe weather situation is certainly consistent with dryline structures we have seen earlier (III.B.5). Since the dryline is a persistent feature in the Southern Plains, Moller's (1979) results for Southern Plains tornado outbreaks should not be surprising. That the basic idea applies in a much greater area than that influenced by the surface dryline may be somewhat unexpected. Recall that considerable empirical and theoretical evidence suggests that most severe storms are accompanied by the intrusion of dry air aloft. If a dryline exists aloft, it seems clear that there is no essential difference between such situations and those involving a surface dryline. Note that the thermal, moisture, and windflow structures associated with bubble highs can create a mesoscale version of the pattern we are discussing (recall Fig. 3.20). The common occurrence of severe weather on these old thunderstorm-created boundaries no doubt results from the favorable thermodynamic structures left behind by the convection (Maddox, et al., 1980), as a return to pre-storm conditions proceeds.

Temperature and dewpoint analyses are useful for locating air mass boundaries. The analyst should perform them before attempting to delineate the features on the surface map. The details of the surface temperature pattern are crucial to proper location of thunderstorm outflow boundaries, which may not have clearly defined wind and pressure fields (i.e., the wind and pressure structure may not resemble the classical patterns shown previously). As with larger scale analyses, the θ_w field can be very useful for this purpose. Recall that the boundary between air

masses is, by convention, put on the warm side of the zone of the strong gradient.

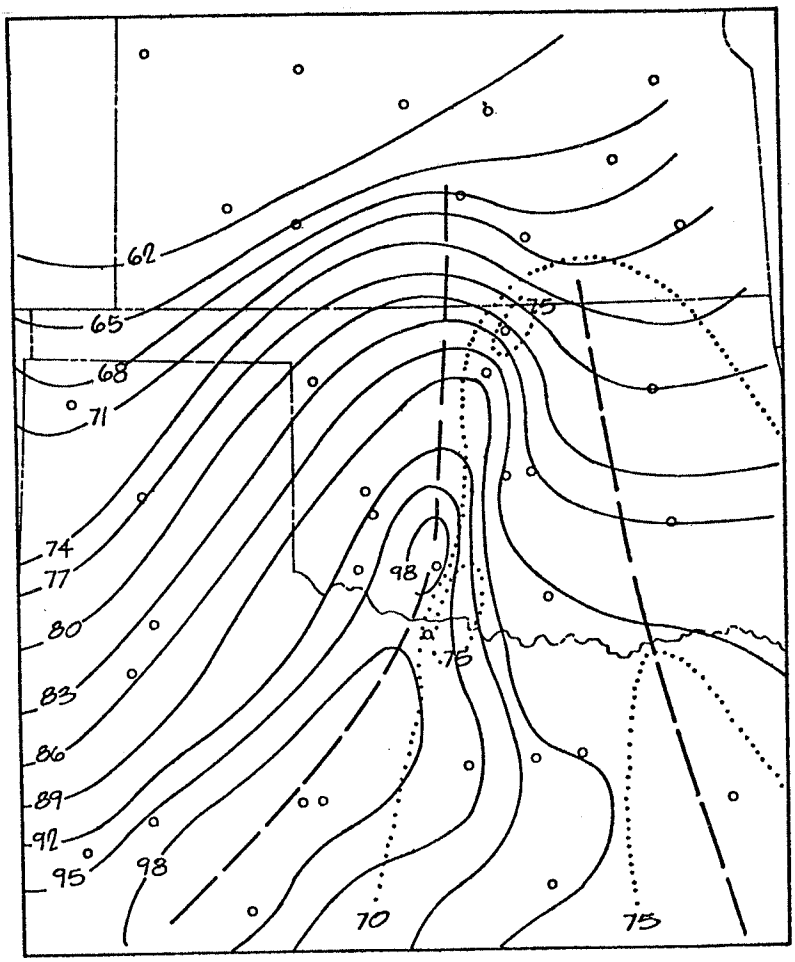


Fig. 3.30. Surface temperature and dewpoint (solid and dotted lines, respectively; °F) analysis at 1900 GMT, 8 June 1974 (after Moller, 1979). Thermal and moisture axes are shown as dashed lines, while open circles locate surface observation sites.

In any case, a careful analysis of the surface temperature (and dewpoint) pattern[4], keeping in mind the above concepts, is in order. As with pressure data, the short term (one- to two-hour) change patterns of temperature and dewpoint can be valuable in locating the thermal ridge (and moisture boundaries), as well as in making a short-range forecast of its movement. Since thunderstorms typically develop between the thermal ridge and the moisture axes, an accurate location and forecast of these features can be critical. In many cases, isotherms at 5°F intervals will suffice to locate the appropriate axes and boundaries -- in

the warm season, it may be valuable to do analysis at $2^{o}F$ intervals. At the same time, areas of thermal (and moisture) advection can be diagnosed, noting that advection is an obviously important factor in local temperature (and moisture) changes. As discussed in III.B, thermal and moisture changes (apart from diurnal effects) do not usually occur independently of the pressure and windflow patterns, so this aspect of analysis should be a routine part of surface mesoanalysis.

F. Terrain Effects

Much of the climatology of severe weather can be directly related to topographic features. It is commonly asserted that the overall pattern of high severe weather frequency in the central United States is a result of the combined effects of several large-scale topographic features (Carlson and Ludlam, 1968). Without casting doubt on this basic idea, one should also be aware of limitations in the climatological record (Kelly et al., 1978; Galway, 1977; Doswell, 1980; Snider, 1977). Simply because a region lacks a large number of reported severe weather events is not sufficient evidence to infer that severe weather is rare in that region. Similarly, the occurrence of severe thunderstorms without much reported tornado activity is not necessarily clear proof of the absence of tornadoes.

The central plains region of the United States has the world's highest observed concentration of strong-to-violent tornadic activity. The existence of a source region for warm, moist air near a north-south mountain chain and the lack of significant east-west topographic barriers seems ideal for the creation of convectively unstable environments. The mountains act to wring out much of the moisture in the upper-level westerlies and the general subsidence of the flow to the lee of the barrier further reduces the humidity of the westerly flow.

Dynamically, the preferred lee side location for cyclogenesis creates a strong tendency for low-level moisture to be drawn from the moisture-rich air mass over the Gulf of Mexico into any developing circulations. Further, the cool air to the north has no restrictions to its southward advance, thereby creating a strong baroclinic zone along which circulations can intensify rapidly by drawing on the potential energy which is the result of the thermal contrast.

It is precisely these factors which are classically seen as creating a favorable environment for severe thunder-

storms to develop. However, this is not the whole story of the terrain's influence. There can be small scale terrain-related phenomena which act to enhance the severe weather potential on a local scale. These are of special interest to the analyst/forecaster.

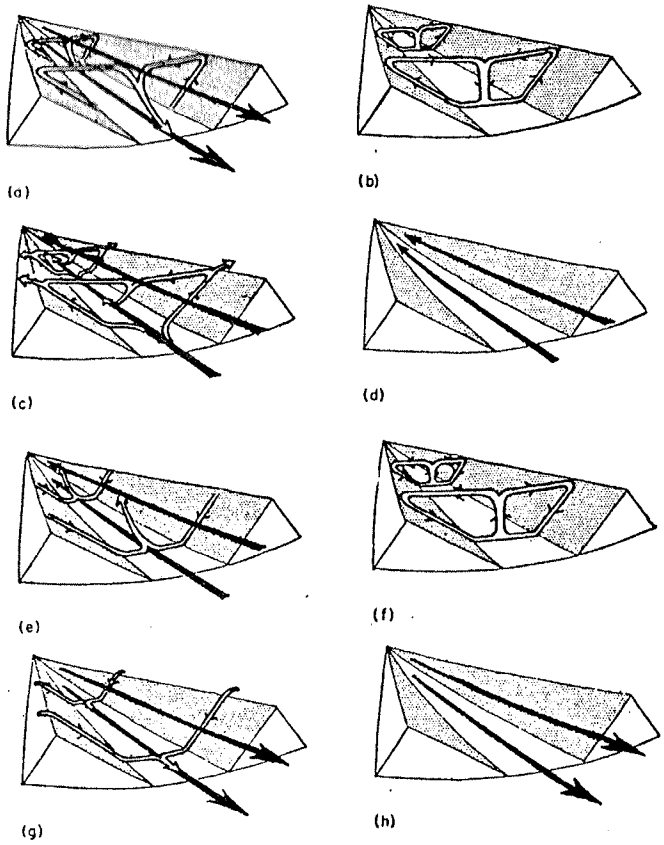


Fig. 3.31. Schematic illustration of the normal diurnal variations of the air currents in a valley, beginning at sunrise (a), and at about 3 h intervals through early morning (b-h) [after Defant, 1951].

1. Mountain/Valley Circulations

Physically, the mountain/valley circulation originates in a manner similar to the land/sea breeze. Since the eastern slopes of mountains face the sun more directly during the early morning (when the sun is at a low angle), they are heated more quickly. A rising plume of warmed air on the sun-facing side of the mountain results, with upslope flow developing to replace the rising air. During the afternoon, the sun is shaded soonest on that eastern face, so the opposite process eventually results in downslope flow.

As described by Defant (1951), this basic picture is complicated by several factors. Figure 3.31 shows the air currents in a valley adjacent to a region of plains. The extremely complex terrain features in any mountainous zone can complicate this basic pattern, but the essence remains fairly simple. A period of experience over perhaps a two-year time span should familiarize the analyst with most of the persistent mountain/valley circulations in the local area. There is bound to be some dependence on seasonal changes and, like land/sea breezes, influences from the prevailing flow regime.

2. Upslope Flow

The trend for flow up- and downslope, as we have just seen, is partially a diurnal effect. As shown by Dirks (1969) in a numerical simulation, this basic circulation can effect the flow over a large portion of the Rocky Mountain chain's lee slopes. In addition to the upslope flow directly over the slopes, a compensating downward current was found by Dirks to exist from near the foot of the slopes, out over the plains for about 40 km. From this point out to about 300 km, a weak upward motion regime could be found, wherever the plains are also slightly sloped. This shall be discussed somewhat more fully in the next section. What is important for the analyst to realize is that the collective effect of the rather abrupt transition from mountainous terrain to gently sloping plains can be seen on a rather large scale. Naturally, upslope flow produces a net upward vertical motion which is clearly associated with the frequent occurrence of thunderstorms along the mountain ranges west of the plains, even in synoptically quiescent conditions. These thunderstorms can be purely local in nature, or may propagate into the plains under the right synoptic conditions (George, 1979).

It should be observed that the strength of the contribution to upward motion by upslope flow is scale-dependent. Although the fine-scale detail of the terrain can create areas of locally large upward motion, these details are smoothed out as we shift our point of view upscale and the vertical motion magnitudes are correspondingly reduced. On the large scale, a terrain-induced vertical motion of several cm s^{-1} is associated with horizontal slopes in the range of several km per 1,000 km (for a horizontal wind speed of order 10 m s^{-1}). Similarly, for a mesoscale vertical motion value of several m s^{-1}, the corresponding terrain gradients are several km per 10 km, which is extremely fine scale topographic detail. See Schaefer (1973b) for some

mathematical details of this aspect of low-level divergence in relation to terrain.

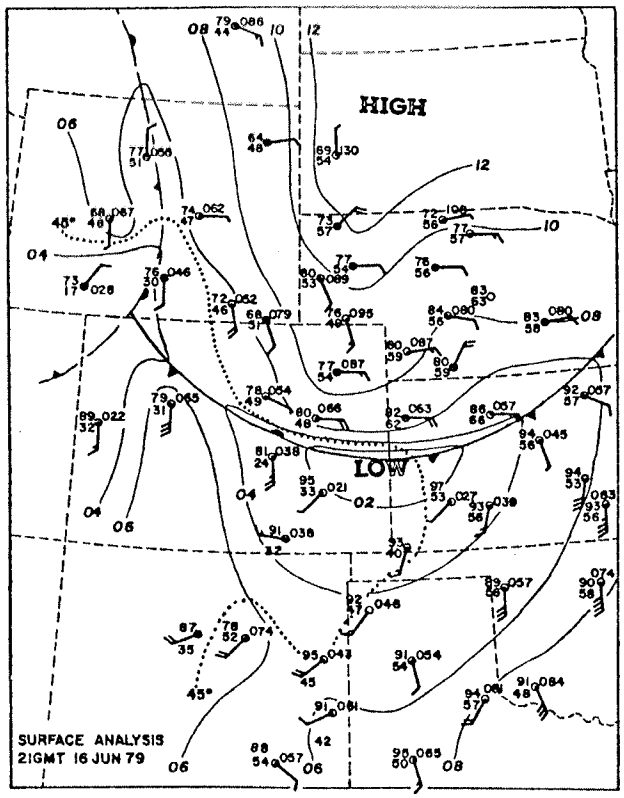


Fig. 3.32. Example of upslope flow in the cold side of a surface front, leading to a High Plains severe weather episode (after Doswell, 1980).

Under the appropriate circumstances, the synoptic-scale pattern can enhance the diurnal trend for upslope flow. Doswell (1980) has offered the idea that the majority of High Plains severe thunderstorms occur under just such conditions. Following anticyclone passage to the east, into the Great Lakes for example, an easterly low-level upslope flow regime can be established which augments the diurnal upslope tendency and the influx of low-level moisture behind a normally-present quasistationary front (Fig. 3.32 shows an example). Given appropriate upper-level flow and adequate instability, severe thunderstorms frequently result. This surface pattern can also be responsible for flash-flood producing convection (with little or no severe weather) under weak flow regimes aloft (Maddox et al., 1978).

It is important to the analyst that conditions of upslope flow be recognized whenever the air being forced to

rise over the terrain has adequate moisture. Such conditions are not confined to the immediate lee of the Rockies, as we shall see.

3. The Low-Level Jet

There are actually two somewhat different phenomena involved in the well-recognized "low-level jet".

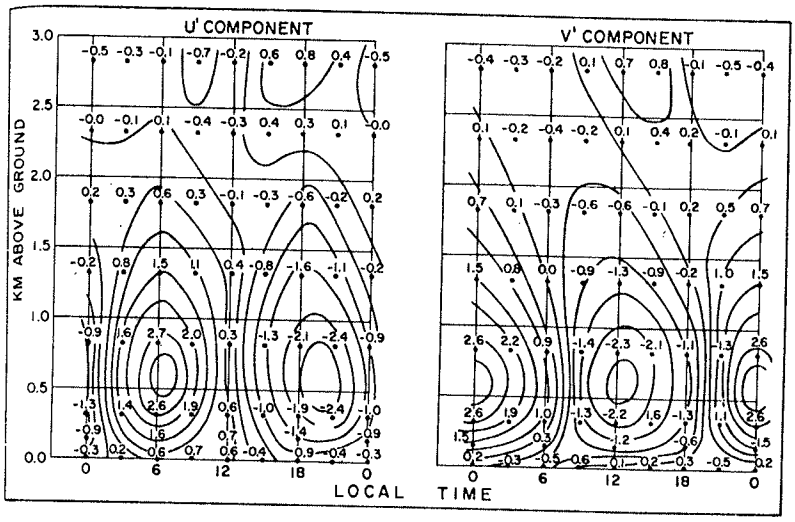


Fig. 3.33. Diurnal variation of wind components from their daily mean value at Fort Worth, Texas; values in m s^{-1} (after Bonner and Paegle, 1970).

The first of these is a nocturnal low-level wind maximum, distinct from the upper-level jet stream. This phenomenon is generally acknowledged to be a result of the de-coupling of the surface from the flow above it, as an inversion is established at night (Blackadar, 1957). This drastic reduction of the friction just above the atmospheric boundary layer creates an oscillation in the wind at that level about its equilibrium position (geostrophic balance). The period of the oscillation (called a half-pendulum day which, at 45°N, is about 17 h) is dependent on the local value for the Coriolis parameter, but the resulting flow speed generally reaches its maximum in the early morning hours (Fig. 3.33).

A second major phenomenon is the so-called low-level jet stream (the terminology here is somewhat confusing). This is a narrow ribbon of high-speed flow, analogous to the jet stream aloft, but restricted to low levels. An example is shown in Fig. 3.34. A variety of detailed theoretical explanations exist, but it is accepted generally that an

essential aspect in its creation is the sloping terrain, via a diurnal variation in the geostrophic wind.

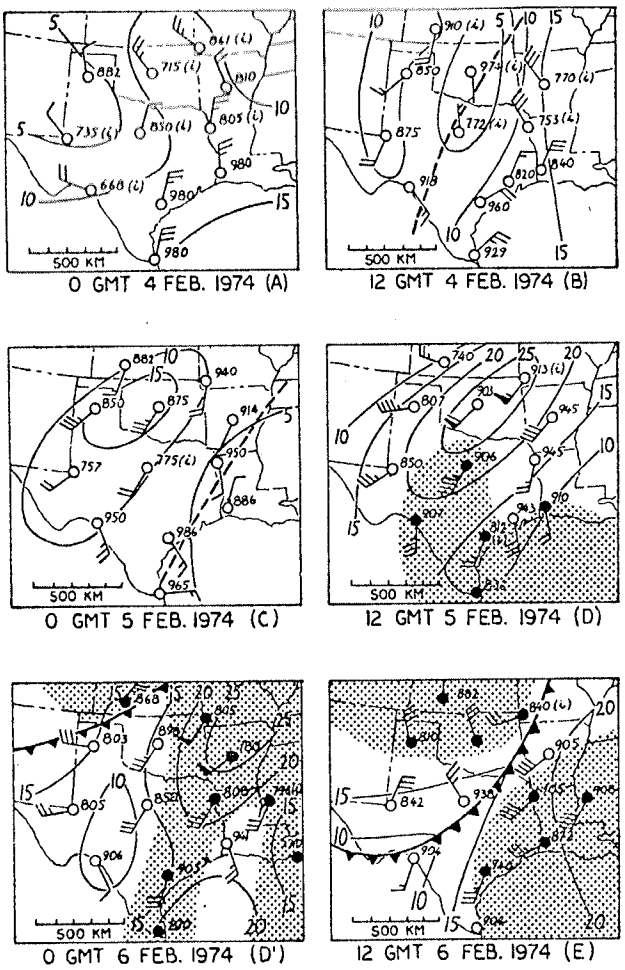


Fig. 3.34. Time evolution of a low level jet stream (after Djuric and Damiani, 1980). Winds shown are at the level of maximum wind in the lowest 2 km, isotachs are in m s^{-1} while relative humidities over 80% are stippled. The dashed line locates the ridge at 850 mb.

A basic element in establishing a classic low-level jet stream is the large-scale synoptic setting. Following passage of an anticyclone through the central United States, a region of anticyclonically curved return flow is created, westward from the Gulf of Mexico, and swinging northward into the plains. This current usually is augmented by establishment of a lee side trough (see III.C.2 above) and can evolve slowly over periods of up to several days (Djuric and Damiani, 1980). By this means, a favorable environment

for severe thunderstorms is re-established by low-level warm advection and a return of adequate moisture. Since the low-level jet stream is a means of enhancing the destabilization of the environment, it is of fundamental importance to the severe thunderstorm analysis/forecast problem. Uccellini and Johnson (1979) have presented evidence suggesting that the upper-level jet streak (a local maximum within the jet stream) and the low-level jet stream are coupled. This seems to be a reasonable hypothesis -- one needs to consider more than boundary layer processes to develop an adequate diagnosis of the low-level jet stream.

Having the appropriate synoptic environment, processes in the boundary layer (the lowest 1 to 2 km) enhance the tendency for development of the low-level jet stream. The speed maximum may be less than 1 km above the surface and the width of the zone of strong winds is generally several hundred km (see Fig. 3.35). Speeds in a well-developed low-level jet stream are in the range of 20-30 m s^{-1}.

In addition to the frictional changes, giving rise to the nocturnal boundary layer wind maximum, the sloping terrain forces a modification of the pressure gradients. This change is not always apparent in the "sea-level pressure" field since it does not properly account for terrain slope. The altimeter correction method of diagnosing the geostrophic wind (Bellamy, 1945) has been incorporated in the analysis of surface geostrophic winds by Sangster (1960). When this is done, it is found that some portion of the diurnal wind variation can be accounted for by the variations in pressure gradient. Sangster's method is examined further in III.F.3 and IV.B.

Numerical models (e.g., Bonner and Paegle, 1970; Chang, 1976) have been formulated which incorporate boundary layer processes. These theoretical models have generally succeeded in reproducing the basic features of the low-level jet. The means by which these phenomena interact with the large-scale setting to produce a concentrated jet stream are not completely understood. Considerations of potential vorticity (beyond the scope of these notes) suggest that the sloping terrain forces an increasing southerly component as the flow off the Gulf of Mexico encounters higher elevations. Also, when the geostrophic wind is parallel to the terrain, the frictional part of the ageostrophic wind is thereby directed upslope. This upslope flow may interact with the diurnal variations to create a concentrated core of southerly flow (Chang, 1976; Schaefer et al., 1982).

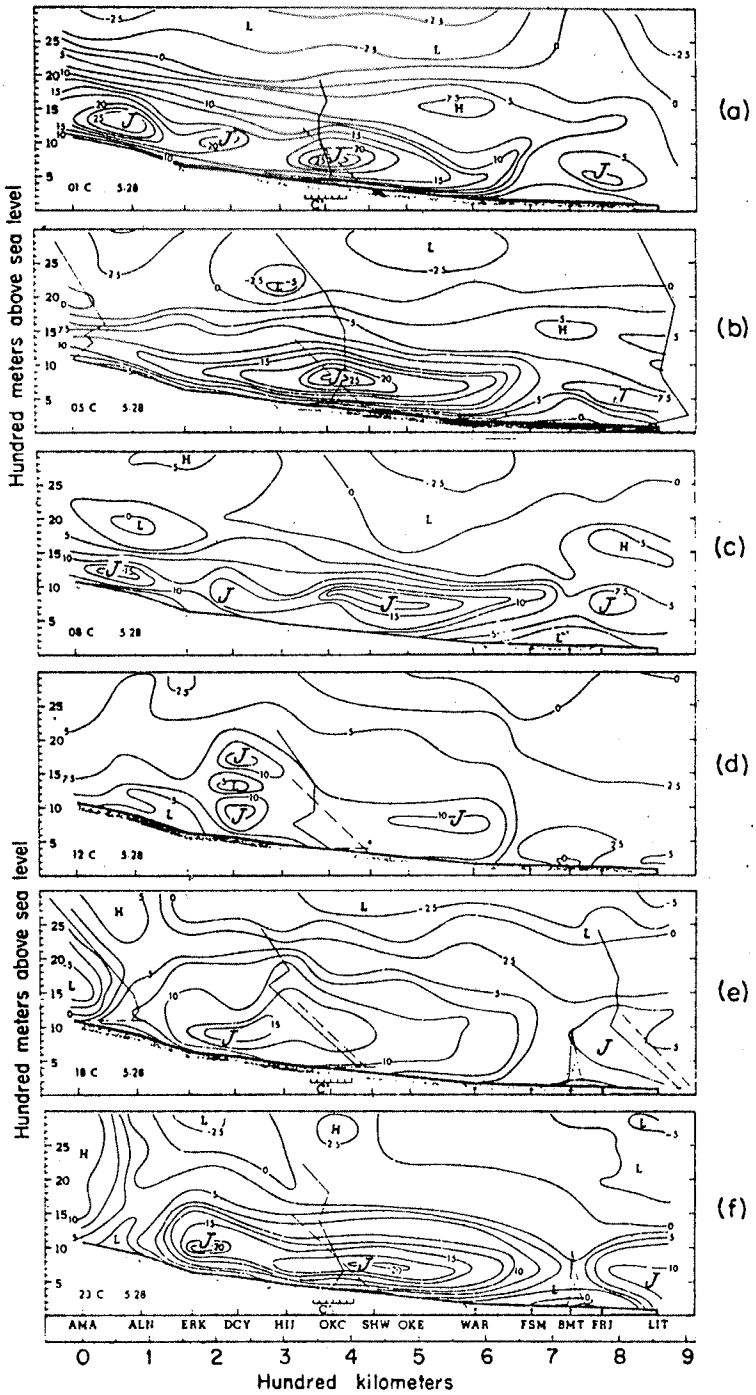


Fig. 3.35. Isotachs of southerly wind component (m s^{-1}) along the line from Amarillo, Texas (AMA) to Little Rock, Arkansas (LIT) during the day of 28 May 1961 (after Hoecker, 1963). Jet cores are indicated by "J" with other maxima and minima by an "H" or an "L", respectively.

Although its destabilizing influences via advection often are emphasized, the low-level jet stream also is associated with fields of vertical motion. While it is not as well-understood as the vertical motion field of the upper-level jet stream (see McNulty, 1978; Beebe and Bates, 1955), the low-level jet stream's vertical motion field is a likely explanation for the nocturnal thunderstorms which frequent the central plains of the United States (Sangster, 1958; Pitchford and London, 1962; Wallace, 1975). This can be easily seen in the pioneering work by Means (1944, 1952, 1954) which reveals a distinct nocturnal peak in low-level warm advection, obviously related to the low-level jet. If one recalls that warm advection is directly related to upward vertical motion,[5] a connection between the low-level jet stream and nocturnal thunderstorms can be seen readily.

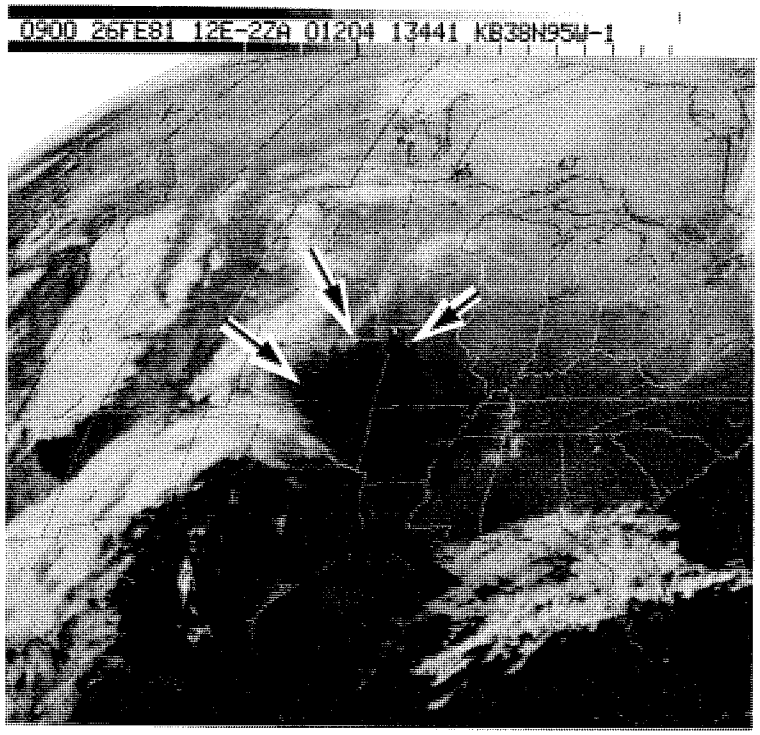


Fig. 3.36. Unenhanced infrared satellite image, showing the intrusion of moisture into eastern Kansas and western Missouri by means of "dark stratus" (arrows).

The space and time scales of this phenomenon are large enough that it often can be detected and monitored. The core of maximum winds usually can be seen easily in the 1200 GMT 850 mb analysis. The normal nocturnal inversion prevents its easy diagnosis in the surface wind field until

about 1500 GMT, after which the low-level wind maxima can be monitored via surface analysis through the remainder of the day. Satellite imagery can also be useful in this regard, since the cumulus field can often be tracked (subjectively) and may provide clues about the low-level winds when animated looping capability is available. Further, the advection of moisture may be seen and tracked via the infrared imagery: the so-called "dark stratus" (e.g., Parmenter, 1976; Gurka, 1980). This phenomenon results from the creation of stratiform clouds and fog in the regions of high low-level moisture. Since these clouds are low, their infrared (IR) emission temperature is relatively high, compared to surrounding clear areas. The clear, dry areas radiate more effectively and become cooler (brighter, in the gray scale used for IR satellite images) than the regions of high moisture and clouds (see Fig. 3.36). While clouds and/or fog are usually present, the effect can be produced by the moisture content alone, perhaps somewhat less dramatically.

4. Mesoscale Eddies

On occasion, the winds and terrain can interact to create mesoscale vortices. One fairly frequent observation of these is unrelated to convection - the wake vortex phenomenon. Zimmerman (1969), Chopra and Hubert (1965), Hubert and Krueger (1962) and others have described the occurrence of long "vortex streets" which can be seen in stratiform clouds downstream from an island. This process is well-recognized in fluid mechanics (Milne-Thomson, 1968, p. 377ff) and takes the form of a series of counterrotating vortices, which are shed into the wake of a flow past an obstacle (Fig. 3.37). These mesoscale eddies are typically observed in convectively stable environments, and are not considered significant in severe weather. Fujita and Grandoso (1968) have proposed that splitting thunderstorms (see II.III.A.5.b.(2)) may be the result of the creation of counterrotating vortices in the wake of a blocking updraft. While this is an attractive hypothesis, it is not generally accepted as the primary mechanism for storm splitting, nor is it terrain-associated.

Another example of a vortex which can be terrain-induced is that which can occur when the flow encounters a "corner". This phenomenon is also recognized in fluid mechanics (Prandtl and Tietjens, 1934, p. 217ff). It has been observed to occur, but it is uncertain how frequently this effect has an influence on severe convection. One possible example of this might be found in a study by Reed (1980). Once again, Reed's case is

non-convective in nature, but it involves a terrain-related small scale cyclone which was associated with damaging surface winds.

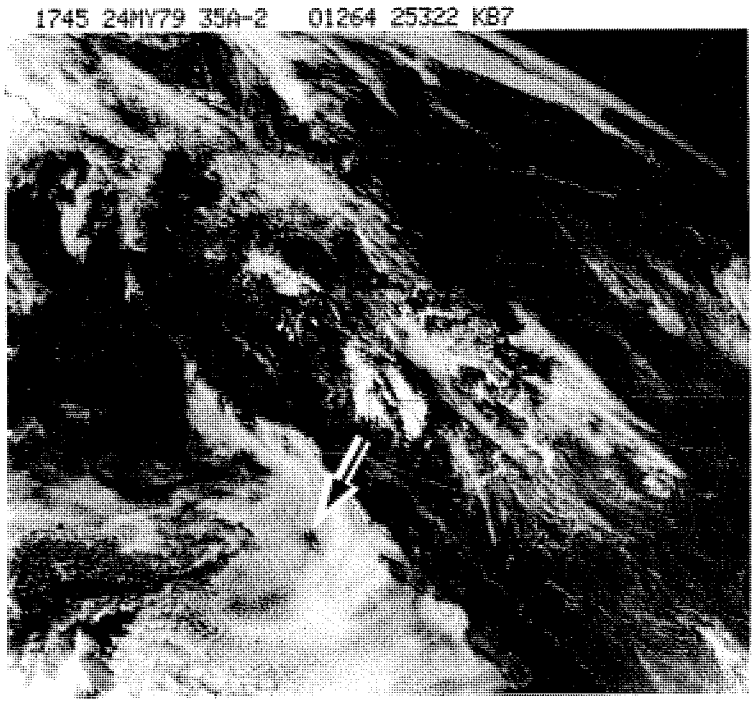


Fig. 3.37. Vortex trail revealed in wake of Guadalupe Island (arrow) in layer of stratocumulus clouds.

Recently, Johnston (1978, 1982) has documented the occurrence of what he calls Meso-scale Vorticity Centers (MVC). These appear during and after the dissipation of a Mesoscale Convective Complex (MCC - see Maddox, 1980b) as cyclonic circulations in mid-level clouds. They have been observed to persist for long periods (from several hours, to more than one day) and occasionally can serve to initiate new convection under certain conditions. Their reflection in surface data is rather subtle and they may not appear at the surface at all. Those which serve to develop new convection (less than 50% of those sampled) usually have a pronounced low-level convergence boundary and move into regions of unstably stratified air. They seem to be induced by the MCC rather than being the result of terrain effects.

It is certainly possible that mesoscale vortices might provide the initial disturbances in an environment where such a disturbance could intensify by other mechanisms. Once

a mesolow has developed, from these relatively obscure origins in a terrain-related or convectively-induced disturbance, convection may be forced by a larger system. The connection between such vortices and severe weather is obscure, but the analyst needs to be aware that such features could influence convection.

5. Miscellaneous Examples

a. The Black Hills Region

In contrast to the exceedingly complex features associated with the Rocky Mountains, the Black Hills area of southwestern South Dakota near Rapid City is a relatively isolated, compact region of uplands. As may be seen in high-resolution satellite images (Fig. 3.38), the region is clearly darker than surrounding plains, making its name quite appropriate. Initial thunderstorm development commonly occurs in this area (Kuo and Orville, 1973). There are two possible contributing factors to the tendency for convection develop earlier in the vicinity of the Black Hills than in the surrounding region.

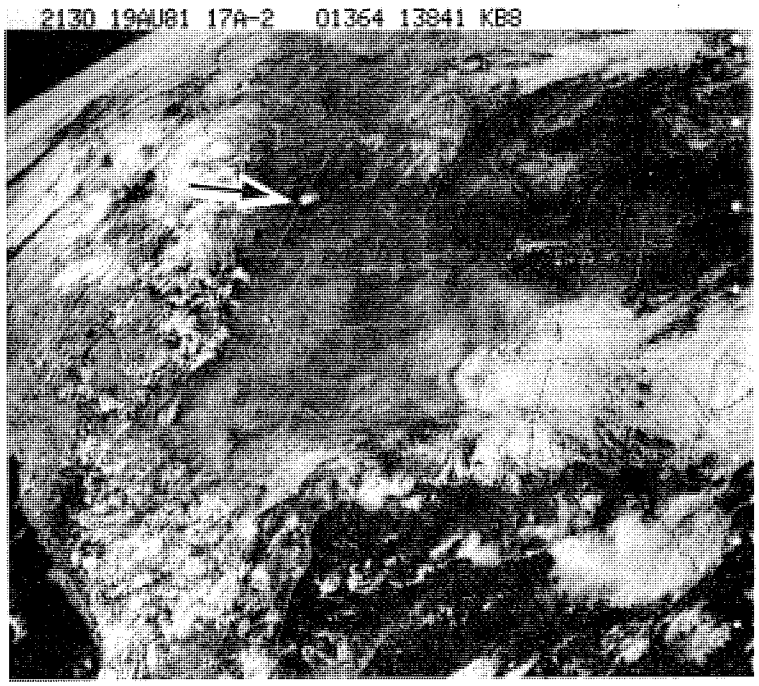


Fig. 3.38. Visible satellite image showing isolated thunderstorm complex developing over dark terrain of Black Hills area in southwestern South Dakota (arrow).

First, the rather abrupt transition from rolling plains to a region of uplands suggests that low-level flow impinging on the hills will be forced to rise. This may allow earlier breakthrough of the capping inversion than in the relatively uniform surrounding terrain. This feature has been exploited in the numerical modelling efforts of Orville (e.g., Orville and Sloan, 1970), to provide the initial impulse for the modelled convection. As may be clear from our earlier discussion of mountain/valley circulations, there is a natural tendency for upslope motion over at least the eastern portion of the Black Hills during the afternoon. This could augment any large-scale upslope flow dictated by the synoptic pattern.

A second contributor to early convection is the darker terrain, which should result in a local temperature anomaly. This would be like the "sea-breeze" around a small island in the afternoon, with a tendency for low-level flow to converge into the Black Hills from all directions. This effect would be distinct from the asymmetric mountain/valley flow, and should certainly be an enhancing factor in the development of convection.

b. The Caprock Escarpment

Much like the Black Hills, the Caprock Escarpment of west Texas combines several terrain-related phenomena in creating a favorable situation for convective storms (under the proper synoptic scale setting). This terrain feature is an extensive plateau, rising abrubtly from the plains to its east, about 100 to 300 m in a horizontal distance of a few km, along a roughly north-south line. Figure 3.39 shows this feature. Note that these terrain slopes yield a mesoscale upward motion of from 10 to 50 cm s^{-1} with an upslope wind component of 10 m s^{-1}. Further, the escarpment is cut by several canyons, with two major canyons being the White River Canyon to the east of Lubbock, extending southeastward, and the Palo Duro Canyon to the south of Amarillo (associated with a major fork of the Red River), which opens to the east-southeast (see Fig. 3.39a).

If this structure is considered, one can see that the escarpment rises relatively sharply from a region of gently sloped terrain, as do the Rocky Mountains further north. Although the barrier is certainly much less imposing than the Rockies, the same basic diurnal tendency for upslope flow and low-level wind structures exists in this region. In addition to the diurnal upslope effect over the plains to

the east, the escarpment provides an additional mechanical lift, since it forces easterly low-level flow upward and also acts as an early heat source, just as the east slopes of mountains do.

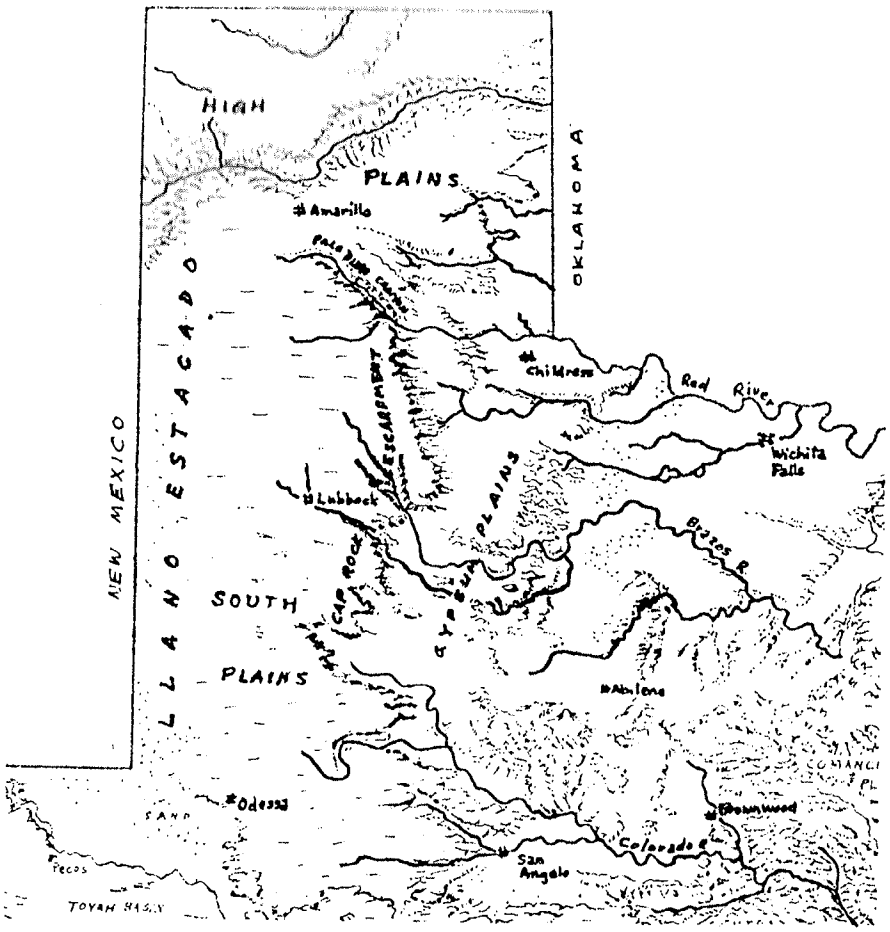


Fig. 3.39a. Physiographic map of west Texas showing the character of the Caprock Escarpment. The abrupt rise runs basically N-S from the Canadian River Valley (north of Amarillo) to well south of Lubbock. The "Llano Estacado" (Staked Plains), which forms the surface of the Caprock, is somewhat more resistant to erosion than the Gypsum Plains to the east. Observe the major canyons which trend WNW-ESE and the especially deep cut into the Caprock at the Palo Duro Canyon (see text for discussion) S and E of Amarillo. Note that the Caprock also has a western escarpment in New Mexico (although it is not as abrupt), part of which can be seen in this figure WSW of Amarillo.

Furthermore, under conditions of southeasterly to easterly low-level flow, moisture can be channelled into the canyons (Fig. 3.39b) and be forced upward into relatively

drier surroundings by the rising canyon floor. This establishes the area over canyons as a favored location for storm initiation or intensification, via forced lifting and locally enhanced moisture. Fankhauser (1971) and Marshall (1980) suggest that the local storm climatology reflects this preference.

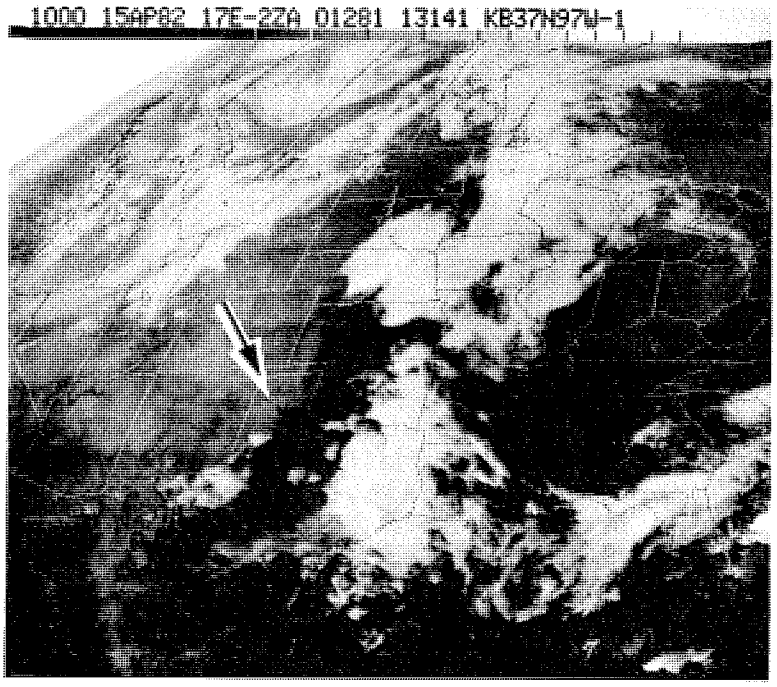


Fig. 3.39b. Illustration of the effect of the Caprock Escarpment and the Palo Duro Canyon. This unenhanced infrared image at 1000 GMT, 15 April 1982 shows the "dark stratus" associated with low-level moisture (see Fig. 3.36) backed up along the Caprock and, especially, into the Palo Duro Canyon (arrow). Compare with Fig. 3.39a.

Another role played by the canyon has been described by Marshall (1980). The south-facing northern walls of the canyons (Fig. 3.39a) receive more direct sunlight (and, hence, heating) during the day than do the canyon floor and north-facing walls. This is analogous to the east-facing slopes of the escarpment which received heating earlier in the morning. Thus, these walls should also be associated with a rising plume of heated air. This may be augmented under southerly flow conditions at low levels, which would tend to follow the terrain and be lifted along the northern canyon walls. As Marshall (1980) notes, the combined effects of the escarpment's topographic features is reflected in the area's precipitation climatology. It is noteworthy that, despite a relatively low population

density, the area of the Caprock is characterized by a well-defined tornado frequency maximum (Kelly et al., 1978).

c. Urban "Terrain"

Recently, it has been recognized that large urban areas have an effect on convection (Changnon, 1978; Braham and Dungey, 1978). With large metropolitan populations, there is enough industry and construction to have a demonstrable meteorological impact. This is generally seen as the combined influences of pollution and heat retention, although the details of the relative contributions by these effects are not completely resolved. Nevertheless, the data collected by Braham and Dungey show an unmistakable concentration of radar "first echoes" in the area adjacent to and downwind from a large metropolitan complex. Although Changnon's tornado data were too limited to make any clear conclusions, the hail and strong wind reports (as well as damaging lightning strikes) were sufficient to indicate that severe thunderstorm frequency is also enhanced by urban effects.

In severe weather climatology, urban areas provide an increase in severe thunderstorm reports merely by virtue of an increased population density. The carefully controlled experiments which provided data for Changnon's and Braham and Dungey's reports do indicate that some of this enhanced severe storm frequency is really meteorological, rather than a reporting anomaly.

Further, the tendency for enhanced convective activity may be present well downstream from the urban complex (Changnon, 1980). The evidence accumulated for the "La Porte anomaly" suggests that a variety of climatological factors, as well as microscale influences can create a condition wherein an isolated maximum in convection is established quite a distance downstream from the initiating urban complex. This maximum can shift in location and strength as a result of long-term changes in weather patterns.

This tendency for large urban areas to influence severe weather (and convective rainfall) is difficult to assess on a day-to-day basis. The natural tendency of the analyst/forecaster to ignore these effects in the face of active weather systems is understandable. However, the evidence indicates that an effort should be made to incorporate these findings into the forecast, when large metropolitan areas are involved.

G. Flash Floods and Severe Weather

While these notes have emphasized the analysis problem with respect to severe weather, convective flash flooding is certainly as potentially dangerous as any aspect of thunderstorms. Many of the analytical tools developed here for severe thunderstorms can be directly applied (shifting the emphasis in parameters somewhat, of course) to the convective flash flood problem (as done, for example, by Hales, 1977). This is a reflection of two related aspects of thunderstorms - first, it is not always obvious how to distinguish those situations which will produce severe weather from those which are predominately heavy rain producers. Second, the same thunderstorms which produce severe weather phenomena are also capable of copious rainfall, and vice versa.

To start with, Maddox et al. (1979) have provided an excellent summary of the features common to flash flood events. These are:

(1) Flash floods are associated with convective storms.

(2) Storms occur in regions with high surface dewpoint temperatures.

(3) Relatively high moisture contents are present through a deep tropospheric layer.

(4) Weak to moderate vertical shear of the horizontal wind is present through the cloud depth.

(5) Convective storms and/or cells repeatedly form and mov over the same area.

(6) A weak, mid-tropospheric, meso-scale trough helps to trigger and focus the storms.

(7) The storm area is very near the mid-tropospheric, large-scale ridge position.

(8) Storms often occur during night-time hours.

With the possible exception of points (3), (4), and (7) these findings could apply equally well to severe thunderstorms. Further, there is ample evidence to indicate that severe thunderstorms have occurred, even when (3), (4), and (7) are valid. Thus, while there is a slight shift in emphasis on certain features of the environment when the

analyst/forecaster is considering heavy precipitation potential, it is not always possible to be confident that severe weather is unlikely. The problem of dealing with convective weather in an operational environment is compounded when both severe weather and flash floods are occurring (see Maddox and Dietrich, 1981). This is also true for tropical cyclones, which are predominantly heavy precipitation producers. While hail is not often observed with tropical storms, tornadoes are certainly not rare (Novlan and Gray, 1974; Smith, 1965; Pearson and Sadowski, 1965). Tropical cyclone-associated tornadoes are still not very well understood, but it is now recognized that they are more common than formerly thought (Gentry, 1982).

Second, it should be clear that a large, long-lasting severe thunderstorm ingests large quantities of water vapor during its life cycle (estimated as high as 8000 metric tons per second!). This will be discussed in more detail in II.III.C, but it is adequate at this point to observe that many convective flash flood situations also involve severe weather, and vice versa. It is a natural consequence of the processes which produce severe weather (not all of which are well understood, of course) that heavy precipitation will often accompany severe phenomena.

CHAPTER III FOOTNOTES

[1] P. III-2: It is often mistakenly believed that altimeter setting is a station pressure value. This is not so; at most sites, the altimeter setting is read directly from an aneroid barometer - it is already a "sea level" pressure. The reduction is via the standard atmosphere. If one needs the station pressure, it can be found by inverting the reduction equation (see List, 1966), using altimeter setting. The so-called "Sea Level Pressure" cannot be used for finding station pressure, since there are "corrections" made in the reduction process which are difficult to reconstruct. Note that altimeter setting can be dangerous to use in regions of mountainous terrain.

[2] P. III-19: The reader should observe carefully several features in this example (Fig. 3.12), especially in regard to mixed layer models (e.g., Keyser and Anthes, 1977). There are extensive changes below about 650 mb from the 1200 GMT sounding to that at 0000 GMT, mostly in the "well-mixed layer". Observe the shallow superadiabatic "contact" layer just above the surface - this is a common feature of evening soundings in the dry air. Also, note that the afternoon well-mixed layer has a value of θ (potential temperature) very nearly that of the average θ seen in the morning, lending credence to the notion that no change of air mass is involved. The slight increase in moisture within the mixed layer during the day is not significant for our purposes here.

[3] P. III-29: It is, perhaps, worthwhile to describe the major weak points in time-to-space conversion, since it has been so widely applied in severe storms research. Naturally, the whole process lends itself much more readily to post-storm (rather than real-time) analysis. However, there are two more rather significant objections to the process. The first is that the "event" which is tracked to provide a conversion vector (a line, along which to plot successive observations) must be assumed to be steady-state. While it can be argued that severe storms are essentially steady-state, there is a growing acceptance that most severe storms are continuously evolving and the details of that evolution are crucial to the production of severe weather (Lemon and Doswell, 1979). A second major objection centers around the choice of a time-to-space conversion vector. It is typical to apply the same vector everywhere within the analysis region. It is not obvious that this should be the case, since individual storms, squall lines, fronts, etc.,

typically are characterized by different motions and to apply any given vector to all the surface observations is of questionable validity. Note that Holle and Maier (1980) have used two motion vectors to account for the motions of two separate outflow boundaries. If a field of vectors is allowed for, then the problem can become complicated beyond any hope of a plausible solution. A third objection can also be made: under certain circumstances, it can be shown that time-to-space conversion may distort the field erroneously - in effect, by creating nonlinear-appearing boundaries from ones which are in fact, linear. It can be argued that it is safer to "leave the observations in the place where they were made". They still may influence later analyses, but ought not to dictate the structure by rigorous application along the conversion vector(s).

[4] P. III-48: It is easy to argue in favor of replacing temperature/dewpoint analyses with potential temperature (θ) and mixing ratio versions. The conversion is simple, but these notes will continue to use more conventional fields.

[5] P. III-57: Based on scaling arguments, it is possible to argue that the connection between warm advection and upward motion (rooted as it is on quasigeostrophic theory) does not exist for low level jet streams. This is not necessarily the case. First, it is certainly plausible to suggest that parts of quasigeostrophic reasoning may well apply to flows where the whole scaling argument does not apply. Second, and more importantly, for adiabatic flow there is certainly isentropic uplift associated with warm advection.

IV. Objective Analysis Tools

While the overall emphasis in these notes is toward an analysis which is done by the analyst/forecaster, there are certain facets of the analysis which are more easily and precisely done via the computer. In general, machine interpolation has been overemphasized as a replacement for human analysis. The computer can draw lines beautifully and the pleasing appearance of the result, coupled with the assumption that "objective is always best", has led to a nearly universal replacement of hand-drawn operational charts with machine-produced versions.

The key factor in evaluating this trend is the general lack of a distinction between the terms "analysis" and "interpolation". Analysis is the process of developing the 4-dimensional understanding of atmospheric events described in I.A. To the extent that machine-drawn isopleths can facilitate the process of analysis, which is the province of the human analyst/forecaster, the computer is a valuable tool. When computer-based interpolation is used as an excuse to eliminate analysis by humans, a damaging precedent is established. This latter concept does not represent the correct "man-machine mix" and has led directly to "meteorological cancer" as described by Snellman (1977).

What, then, is the appropriate role for computer applications in the analysis process? Perhaps the most obvious is in the realm of large-scale analysis. The sheer volume of charts to be drawn indicates that much of the preliminary interpolation should be done by machines. The need to establish a smoothed representation of the large, synoptic-scale pattern should be apparent. In order to grasp the whole setting, upon which are superimposed the subsynoptic scale weather systems, one must necessarily examine the data on the large scale (e.g., all of North America, or the whole Northern Hemisphere) and smooth out the details (which are to be examined during the analysis phase). This is clearly most easily accomplished objectively.

Further, the plotting process for local mesoanalysis ideally should be done by machine. The apparatus to do so is in the AFOS system, and this capability should be expanded to include all available observations, as well as the capability to plot "change" variables (like short-term pressure changes). These programs are currently available at SELS, with the discussion of change variables having been included above.

Finally, having determined the overall pattern, there are many derived parameters which require far too much computation to be accomplished by hand. One can use the data to calculate divergence, vorticity, streamlines, and geostrophic winds, to name a few of the host of potentially valuable parameters. It is simply not possible to do this quantitatively with the eye, nor is it practical to compute them laboriously by hand. Meteorological literature abounds with parameters which some one has felt can make a contribution to analysis. Several of these approaches have already been included in the products routinely developed in SELS, and these are to be described here.

A. Moisture Convergence

Two of the primary factors in developing severe weather potential are low-level convergence and a supply of moisture. These may be combined in the field of moisture convergence (more properly, the moisture flux convergence). This field combines the influences of convergence and moisture advection. Thus, the divergence (a mathematical operation usually to be accomplished with data interpolated to a grid) of the product of the wind and some measure of moisture (e.g., mixing ratio) is computed. This can be done at any level in the atmosphere, but the surface data density and frequency make it the most-often chosen level. An example of the SELS version of moisture convergence (available on AFOS) is shown in Fig. 4.1.

A wealth of literature supports the basic idea (Hudson, 1971; Ostby, 1975; Doswell, 1977; Ulanski and Garstang, 1978) that this is a valuable parameter. In general, surface moisture convergence precedes the development of convection, allowing this diagnostic field to have short-range prognostic value. This makes physical sense, in that once moisture convergence begins, it should take a few hours to break through the capping inversion (normally present) and to accumulate a supply of moist air upon which the storms will draw.

It is worthwhile to consider some of the limitations in applying moisture convergence computations to the analysis. First, the strength of moisture convergence is highly scale-dependent. On the synoptic scale, values are generally in the range of 10^{-4} g kg^{-1} s^{-1}. As pointed out by Fritsch (1975) (see also Fritsch et al., 1976), this rate of moisture convergence can supply only about 20% of that needed to sustain a severe thunderstorm complex. Thus, on the subsynoptic (or meso-alpha) scale, there must be a large increase in moisture convergence (a factor of 5 or more) to

maintain a long-lived storm complex. Doswell's (1977) results substantiate that this increase does occur. Since the strength of moisture convergence is scale-dependent, the availability of data determines, to some extent, the resulting field. Clearly, in some situations, the detail necessary to diagnose the moisture convergence properly may not be available, and the computed field may not be representative of what the threat area is actually experiencing. The analyst should be wary of computed values in data void (or missing data) areas.

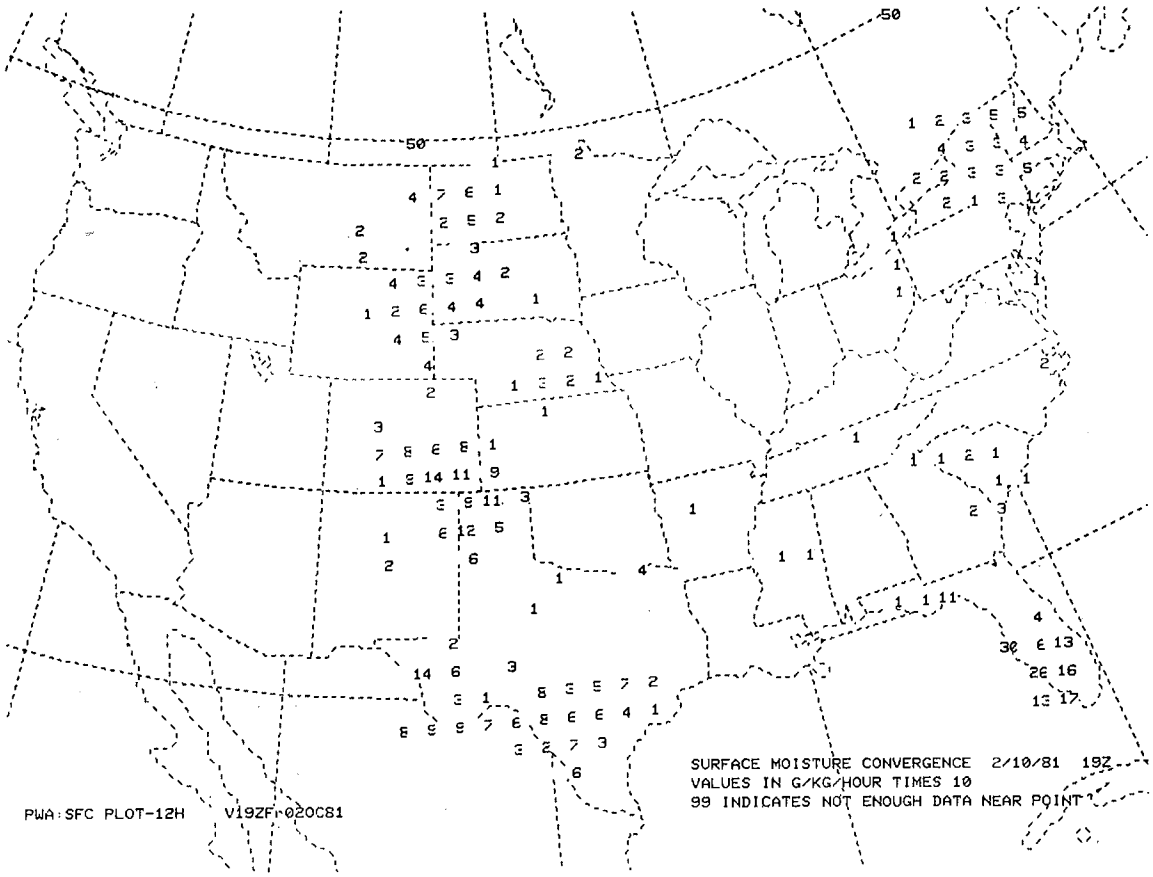


Fig. 4.1. Example of surface moisture convergence values at grid points produced at NSSFC and displayed on AFOS console.

Further, it is not always the case that conditions at the surface level reflect a clear picture of what is occurring. When an inversion is present, the roots of storm updrafts (and the zone of preceding moisture convergence) may be aloft. This is often true for nocturnal thunderstorm situations. In such cases, the surface moisture convergence may be completely unrelated to the weather, since surface-based sensors do not detect the physically significant events.

A problem related to the discussion in III.B.1 is that the surface wind observations are the primary factor in determination of moisture convergence. Errors and unrepresentative observations may produce completely fictitious centers of moisture convergence. These can usually be easily detected by the associated "bulls-eye" patterns, but the truly representative value may be lost irretrievably. A carefully thought-out screening procedure for the input data is essential. As discussed in III.A, this may not be entirely straightforward, since the observations may be both real and actually representative in some sense. Having the plotted input data available for examination alongside the resulting moisture convergence field is a way for the analyst to arrive at an assessment of the quality of the result.

Since moisture convergence usually includes both the product of moisture with convergence and the advection of moisture, this parameter may be somewhat misleading. It may be useful to have a separate measure of convergence alone (normally, the dominant term). If one has this, it is possible to keep track of travelling convergence centers which may be distinct from the moisture convergence pattern. That is, the moisture advection may combine with the convergence contribution to yield a relatively slow-moving moisture convergence field - yet what one really has is a moving convergence field intruding into a zone of strong moisture advection. The distinction could be significant.

Finally, satellite imagery can be a useful supplementary tool in checking on the moisture convergence field. Areas of enhanced cumuliform cloudiness frequently precede the onset of deep convection. As these develop and evolve, they should be compared with the moisture convergence field. It may be possible to locate moisture convergence zones in this manner which are not detected by conventional data. Small scale features in the moisture convergence can often be easily seen in the satellite data - e.g., the lines of cumulus congestus along "arc clouds" and fronts, compared to large areas of ordinary cumulus clouds.

B. Surface Geostrophic Winds

The production and application of surface data-generated geostrophic wind charts has been pioneered by Sangster (1960). An increasing acceptance of the value of this approach can be seen in its current availability via teletypewriter and its incorporation in the AFOS products. An example of the AFOS-available surface geostrophic wind

field is given in Fig. 4.2. Therefore, it is important to emphasize that this evaluation of geostrophic wind is not a straightforward application of the well-known geostrophic wind law to the conventional "sea-level pressure" field. Rather, it uses the so-called altimeter correction system (Bellamy, 1945) to incorporate the earth's surface topography. The details of the derivation are not important here, but the resulting product normally differs somewhat from what one might expect by looking at an NMC surface pressure analysis. This is partially caused by the influence of topography, and partially by the process by which a solution is obtained (a high degree of smoothing is done).

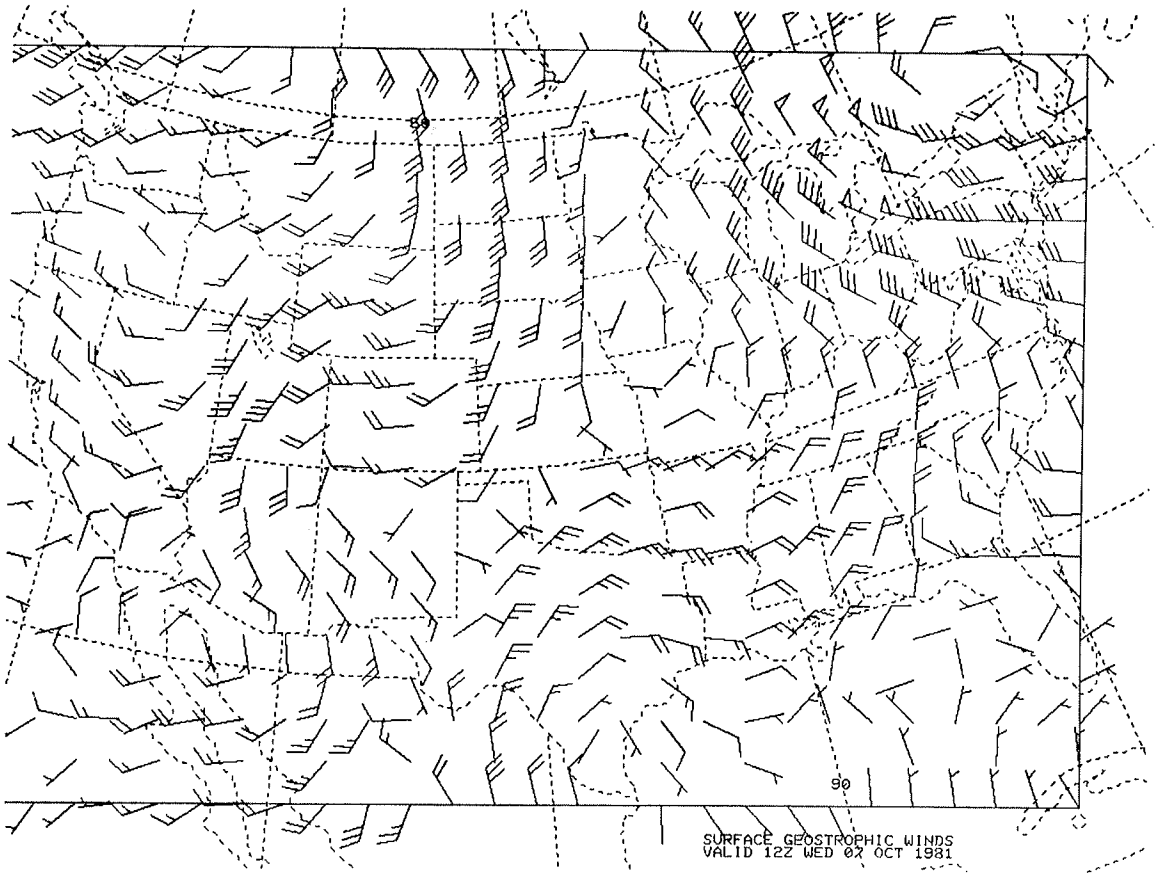


Fig. 4.2. Example of surface geostrophic wind field produced at NSSFC and displayed on AFOS console.

One of the benefits of Sangster's approach is that the diurnal variation in the geostrophic wind is directly incorporated. As described in III.F.3, this geostrophic wind variation is often reflected in the real winds and plays a role in the low-level jet stream (Sangster, 1967).

This diurnal wind variation, in turn, is an important factor in thunderstorms (Means, 1952; Sangster, 1958; Pitchford and London, 1962; Bonner, 1966).

Although the geostrophic wind is essentially non-divergent, this chart can be used effectively in convective forecasting. The zone of upward motion associated with the low-level jet lies generally to the left of the jet axis. Exceptions to this are either when the jet impinges on a boundary, or the calculated geostrophic speed decreases rapidly at the "nose" of the jet. These situations imply upward motion ahead of the maximum wind core along the axis, an implication which is generally substantiated in practice. This, as we have seen, is generally associated with strong warm advection.

Another advantage of the surface geostrophic wind chart and its main advantage over either the surface pressure map or the observed winds is its continuity. The changes in the field are relatively slow to occur and they accurately reflect the overall march of events. Observed winds fluctuate substantially and this makes it difficult to monitor the actual time evolution and movement of the low-level jet stream.

It is noteworthy that the surface geostrophic winds do not always relate clearly to the observed winds. It is obvious that much of the time, the observed winds are much slower than geostrophic. This is a natural consequence of surface friction. Schaefer and Doswell (1980) have recently incorporated surface friction into the force balance in an objective way, producing fields of the so-called antitriptic wind. This has not yet been implemented on an operational basis, but the suggestion is that by incorporating friction, a theoretical wind is obtained which is more appropriate at the surface than the geostrophic. This effort has been motivated in part, by the problem of inferring accelerations from the difference between observed and geostrophic winds.

The concept of ageostrophic acceleration is important to understand, and the reader should consult textbooks (e.g., Saucier, 1955, p. 240ff) for a more thorough discussion. Briefly stated, when the wind is not geostrophic, accelerations exist which act to turn and change the speed of the flow, in an effort to reach the balanced geostrophic equilibrium state. These accelerations are critical to understanding weather since, as we have discussed, the divergence (and, hence upward motion) associated with geostrophic flow is not physically significant.[1] As Schaefer and Doswell (1980) have pointed

out, the frictional contribution to the ageostrophic wind is large at the surface. Thus, a force balance ignoring friction is just not adequate. In order adequately to diagnose significant accelerations, one must first account for the friction -- what remains in the way of "non-antitriptic" winds (analogous to ageostrophic) is then more likely to be physically important.

Needless to say, since the antitriptic wind is slower than and directed to the left of the geostrophic, should one encounter observed surface winds in excess of geostrophic, then (assuming it is not an error) something significant certainly is occurring. The equations governing motion state that supergeostrophic flow is accelerated to the right of the geostrophic wind. A moment's thought should reveal to the reader that this is toward high pressure. Similarly, subgeostrophic winds are shunted toward low pressure (as may readily be concluded from examination of most surface maps). Supergeostrophic (or perhaps more appropriately, superantitriptic) winds are an indication that important events are underway and need to be monitored.

C. Filtering by Objective Interpolation

Although the primary surface data analysis responsibility lies with the analyst/forecaster, there are areas where an objective interpolation of the primary fields can provide new insights. Specifically, it has been shown that carefully designed objective interpolation schemes (Doswell, 1977; Maddox, 1980a) can be used to separate mesoscale features from the large-scale pattern. When using surface data, Doswell (1977) has included time series data (without going through the time-to-space conversion), in order to simulate the time continuity that subjective analysts can impose. This has a variety of beneficial effects, including a reduction of the tendency that most objective interpolation techniques have to place extreme values between observing sites. It also limits the hour-to-hour "jumps" in the location and strength of the extrema.

The advantage this objective interpolation scheme provides the analyst is that it helps to isolate and enhance those mesoscale features that are truly supported by the observations. Further, it can be used to develop the derived parameters, such as divergence and vorticity, associated with those mesoscale phenomena. This technique probably comes the closest to reproducing what the human can do. By providing the opportunity to examine derived parameters at the small-scale limits of the data, a substantial

benefit is gained. It can also be an important analysis aid to the less experienced individual.

An obvious disadvantage is the introduction of an additional processing step which requires substantial on-site computer capability. The speed and timeliness of this analysis aid is directly proportional to that capability, which for the time being makes it impractical for universal application. In order to compensate for the time delay, an objective analysis tool should provide the analyst with some parameter or insight otherwise unavailable.

D. Upper-Level Divergence

Having discussed low-level moisture convergence analysis in IV.A, some mention of the required upper-level divergence is necessary. As in the case of low-level data, it is simply not possible to diagnose upper divergence by eye. It is possible for the subjective analyst to locate certain indirect indicators of upper divergence. We have already considered some of these, especially positive vorticity advection. Another commonly used indicator of upper divergence is "difluence", but it should be recognized that difluence and divergence are not equivalent! As detailed by McNulty (1978), the left front and right rear quadrants of upper jet maxima are favored for divergence (with the justification calling upon vorticity advection arguments). No doubt a myriad of empirical rules governing the use of upper-level charts for convective forecasting can be theoretically justified through some connection with upper-level divergence.

With all these indirect methods, of varying quality, one could legitimately ask: Why not compute upper-level divergence directly? There are several problems with direct calculation of divergence which we have already considered (III.B.1), and which are described in textbooks (e.g., Haltiner and Martin, 1957, p.314ff). The main argument is that it is difficult to do so because the two terms used are relatively large and of opposite sign, so we end up taking the difference of two large numbers. Such an operation can result in the creation of substantial errors. Another not as well-recognized problem is the quality and quantity of upper-level wind data. When the sonde has reached, say 300 mb, it may well have traversed enough horizontal distance that it is low on the horizon. Low elevation angles can create substantial errors in wind computations (see, e.g., Middleton and Spilhaus, 1953, Ch. VII). In fact, in situations of greatest interest -- i.e., those with strong winds aloft--this low elevation angle problem is at its worst.

Therefore, winds above 300 mb may have substantial errors associated with them. Finally, each rawinsonde terminated below 300 mb reduces the overall number of observations. It is not a rare event that at least one or two rawinsondes do not reach 300 mb in a given set of synoptic observations.

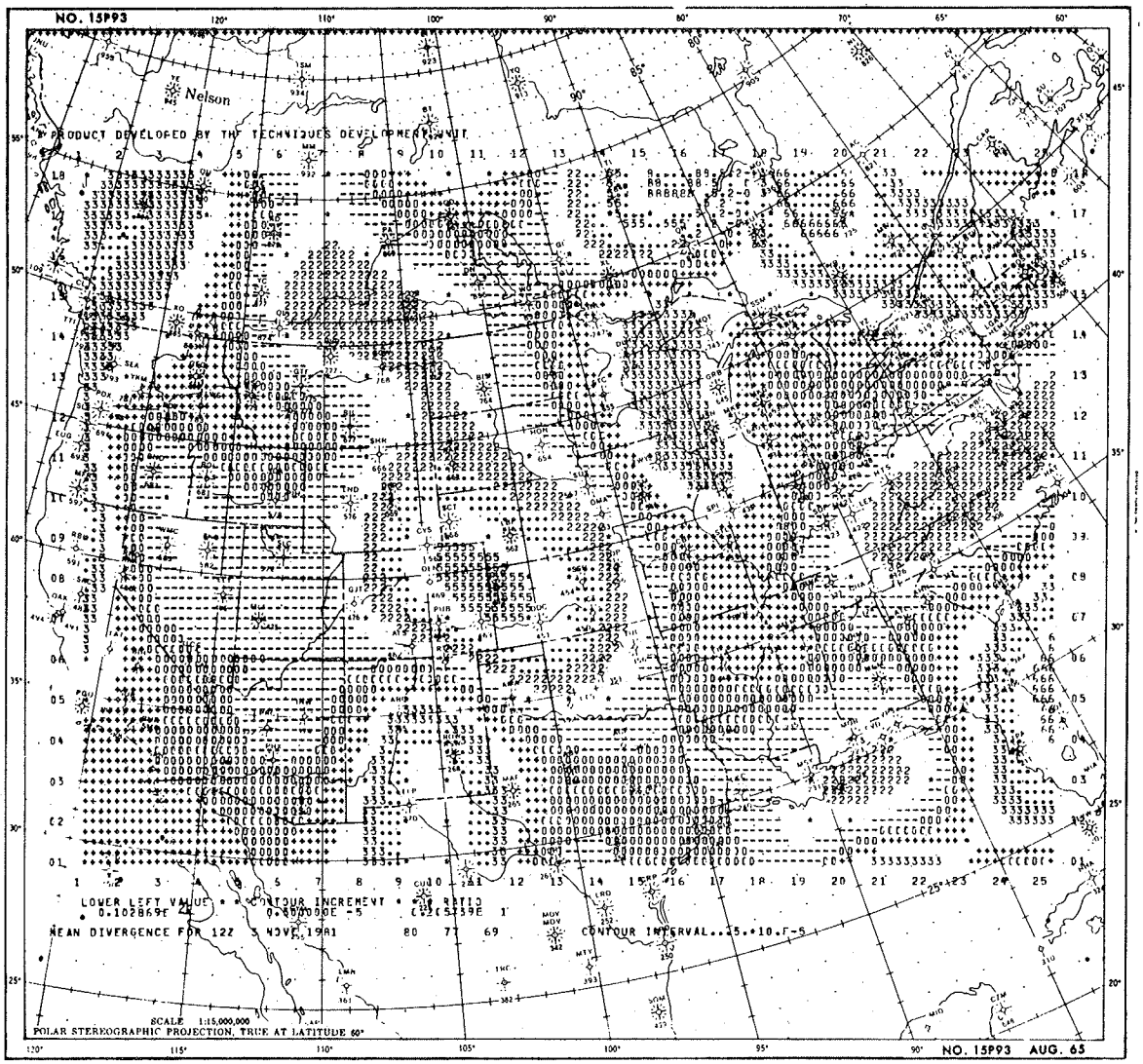


Fig. 4.3. Example of mean 300-200 mb divergence contours produced at NSSFC on map background. The zero line is along the boundary between "0" and "-", with contours at intervals of $\pm 5 \times 10^{-5}\ s^{-1}$.

In spite of these problems, there is good reason to believe that a careful analysis of divergence is possible. The keys to producing a meaningful divergence calculation are a pre-analysis screening for obviously erroneous data and the proper smoothing of the wind field to be used in the divergence calculation (see Panofsky, 1964, p.33f).

McNulty's (1978) approach provides a mean divergence between 300 and 200 mb (in effect, an upper level vertical motion field).[2] This vertical averaging tends to smooth out the irregularities which might be found at any given level. His results (an example is shown in Fig. 4.3) suggest that the analysis does contain meaningful information about the divergence field. In a related area, Schaefer and Doswell (1979) have shown that a different method for calculating divergence and vorticity (using line integrals) is inherently superior to the conventional method (using derivatives of the wind components). In some unpublished examples, McNulty has found that the line integral method does, indeed, produce a field which better relates computed divergence to satellite images of cloud patterns. The line integral approach to upper divergence remains to be tested operationally, as it is somewhat more time-consuming than conventional methods.

E. Kinematic Analyses and Trajectories

We have already mentioned vorticity and divergence many times. These parameters are two of the four main properties of the wind field. The other two are stretching and shearing deformation. It is possible to use either divergence and vorticity or the two components of deformation to reconstruct the wind.[3] When using divergence and vorticity, for example, the divergence can be used to find the "irrotational" wind contribution and the vorticity to find the "non-divergent" contribution. Non-divergent winds have been used extensively in the initialization process of numerical models and in the study of atmospheric dynamics (e.g., Haltiner and Williams, 1980). The reader should recall that the geostrophic wind is an example of an essentially non-divergent wind (to a first approximation). The isallobaric "wind" is an example of an irrotational field, but it should be noted that it does not represent a true wind (see III.D.2).

While these notes cannot provide much working knowledge of the often-neglected topic of kinematic analysis, some topics deserve special attention as they relate to objective "analysis" and also to forecasting. Readers are urged to pursue the whole range of kinematics in the references (esp. Saucier, 1955, Ch. 10; or Petterssen, 1956a, Ch. 2).

Deformation is a kinematic property to which relatively little attention is given. This is unfortunate since deformation is that property which is characteristic of fluid flow. Note that there are two components: stretching and shearing. It is possible to combine these into a single,

resultant deformation (which is always non-negative) with a resultant axis of dilatation. This is analogous to combining the two components of the vector wind into a (non-negative) wind speed and a result direction. Shown in Fig. 4.4 are the two components of deformation. Note that each has both an axis where the winds are "converging" and an axis where the winds are "diverging" (perpendicular to the former). When the two components of deformation are combined mathematically, the resultant deformation has an axis of "diverging" winds along the resultant dilatation axis. The orientation of the resultant dilatation axis shifts, depending on the relative contributions of the shearing and stretching components (just as in the wind component analogy). The variation of this axis with the components is shown in Fig. 4.5. Note that the range of directions is only 180°, since the axis of dilation is not directed.[4]

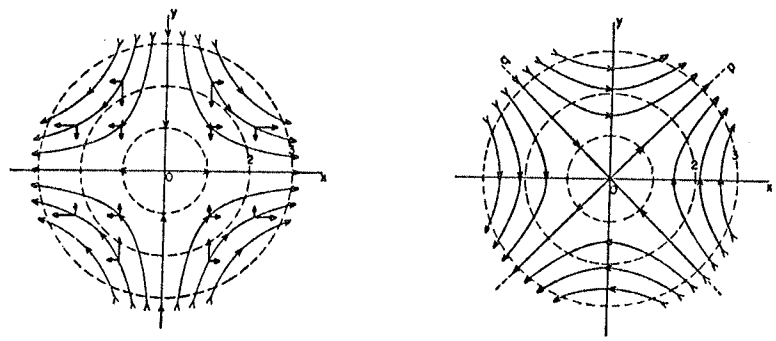


Fig. 4.4. The two types of horizontal deformation (after Saucier, 1955). On the right is "shearing" deformation and on the left is "stretching" deformation.

One should be careful, in considering Fig. 4.4, to note that these two idealized components of the total deformation need not be apparent in actual flows. It is all too common to ignore deformation unless the flow has this "hyperbolic" appearance - this is simply not valid. The pictures of Fig. 4.4 should be thought of as relative flows, since the total flow can have intense deformation without any such pattern apparent (see Saucier, 1955; Doswell, 1982b).

Distribution of atmospheric properties like temperature, moisture, etc., are influenced dramatically by deformations. This is apparent to anyone who has observed cloud motions in an animated satellite loop. Cloud patterns are stretched and sheared by the flow, and this process often results in the formation of cloud lines. Not only do

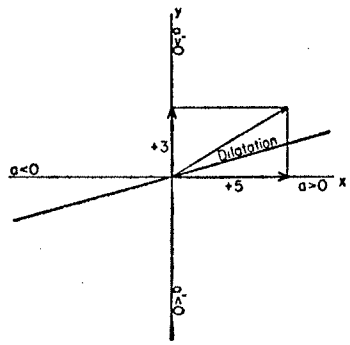


Fig. 4.5. Illustration of how the two components of deformation may be combined into a single "resultant" deformation (after Saucier, 1955). The stretching deformation value (a) is plotted along the x-axis, while the shearing component (a') is plotted along the y-axis. In this example, a = +5 and a' = +3, to give an angle (at the upper right of the rectangle) with respect to the x-axis of 31° for the axis of dilatation, the angle is half of this value, or 15.5°.

thunderstorms frequently develop in lines, but all cloud types can occur in linear features. The banded character of cloud patterns may have its origins in processes other than deformation at times (see e.g., Kuettner, 1959), but deformation is a highly visible aspect of atmospheric flow.

Since clouds typically arise from vertical motion, it can be seen that divergence patterns should also be influenced by deformation. The banded characteristic often is not apparent when objectively analyzed fields of divergence are examined. Such by more or less circular patterns, rather than bands, of divergence and convergence.

Why is this the case? A large part of the explanation is the nature of objective interpolation to uniformly distributed grid points. This has been recognized, but is not often emphasized. Wiin-Nielson (1959) has pointed out:

> ... there is a tendency to deform the initial rather regular pattern into the structure of elongated bands ... The stretching is in some regions so large that the bands disappear between the grid points ... We are therefore bound to get a smooth picture. This does, however, not mean that we should neglect the deformation properties of the fields.

Naturally, these statements apply to grid point models as well as to grid point analyses. This is not a fault of the grid point analysis scheme chosen or of the parameters of the scheme. As Barnes' (1964) method exemplifies, it is possible to interpolate objectively so as to reproduce the

observations to whatever degree is desired. Rather, it is a characteristic of all such schemes.

Can this defect of objective interpolation be overcome? As discussed in II.A, the human analyst can draw (or re-draw, if using machine-prepared isopleths) long, narrow ribbons during the analysis. To guide this process, use may be made of satellite images and temporal continuity (i.e., trends toward developing the bands). Another, more objective approach for negating this tendency (inherent in using grid points) is trajectory analysis and forecasting. As with kinematics, the details of trajectory approaches are beyond the scope of this text. References should be consulted (e.g., Saucier, 1955; Doswell, 1982a, Wiin-Nielson, 1959; Reap, 1968, 1972).

In the most simple terms, trajectories trace out the paths of air parcels over some finite time period. One should be aware that trajectories and streamlines are not generally the same. Unless the flow is completely steady-state, the trajectories will differ from streamlines in possibly important ways. The advantage to calculation and use of trajectories (usually called Lagrangian methods) is that trajectories can account for the deformation in the flow directly. Parcels tend to collect in deformation zones, allowing for more resolution in precisely those areas where gradients are becoming most intense.

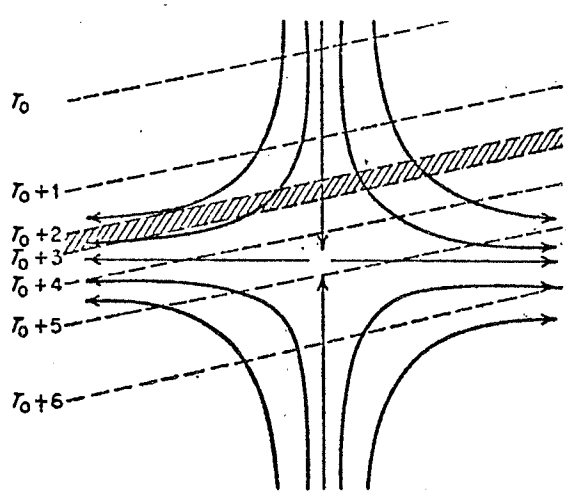


Fig. 4.6. Illustrating frontogenesis. The line of frontogenesis (hatched) moves toward the axis of dilatation, while the temperature contrast increases (after Petterssen, 1956a).

At this point, we should digress briefly to consider a topic implied by the previous sentence: frontogenesis. In essence, frontogenesis is the process by which gradients of atmospheric quantities are intensified. Deformation fields, as one might expect by our initial discussion on the subject, have a crucial role in frontogenesis (Saucier, 1955; Petterssen, 1956a; Miller, 1948; Stone, 1966; Hoskins and Bretherton, 1972, etc.). The circulations involved are not forced by the frontogenesis but, rather, accompany it. The topic is complex and still the subject of ongoing research. However, we should be aware of the strong linkages among deformation, frontogenesis, vertical motion, and divergence.

Briefly, if we consider frontogenesis to mean the accumulation of gradient for, say, potential temperature (as in the common definition), then there are several factors which need examination. The horizontal accumulation of gradient can be shown to be the product of the original gradient value and the combined effects of convergence and contraction via deformation. Thus, the larger the initial gradient, the more rapid the frontogenesis. Convergence is clearly accumulative, but in order for the deformation to contribute to frontogenesis, the axis of dilation must be tilted at less than a 45° angle to the property lines (see Fig. 4.6).

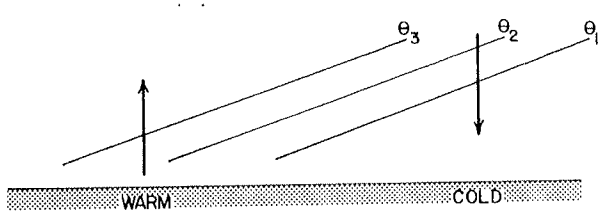


Fig. 4.7. Illustration of the effect of a direct circulation (warm air rising; cold air sinking) on horizontal gradients of potential temperature. The same picture arises if the air is rising (or sinking) everywhere, but the warm air is rising relative to the cold.

One should also note that horizontal differences in vertical velocity can act to "tilt" the normally strong vertical gradients into the horizontal, resulting in an increase of horizontal gradients. It is a process which is frequently ignored, but this neglect is perilous. The justification usually states that vertical motions are so weak, that gradients in vertical velocity are unimportant. This is substantially in error, since vertical gradients of atmospheric variables are often quite strong. The contribution by differential vertical advection can easily be as

strong as, or stonger than, that produced by horizontal advective effects. One also should observe that, in general, a vertical circulation in which relatively warm air rises and cold air sinks (a direct circulation) acts to destroy horizontal gradients of potential temperature. This can be seen easily in Fig. 4.7. Clearly, a vertical circulation of the opposite sense (an indirect circulation) increases the horizontal potential temperature gradient.

By increasing resolution in deformation zones, the trajectory method offers some distinct advantages over grid point schemes (usually called Eulerian methods). There are, of course, some disadvantages. For example, it is hard to construct "weather maps" from the distorted structures which results (see e.g., Welander, 1955). Reap (1968) uses backward trajectories to compensate for this -- i.e., he traces parcels backward in time from a uniform grid. However, this creates an end product which still "suffers" from the uniform resolution implicit in a grid. Other problems exist, but it suffices to say that the grid point approach is best suited for numerical models which forecast for periods beyond 12 h. Many of the problems are of substantially lesser significance for forecasts of 12 h or less (Doswell, 1982a).

CHAPTER IV FOOTNOTES

[1] P. IV-6: Observe, however, that this is not strictly true. When the geostrophic wind is strong and directed northward (southward), the convergence (divergence) associated with the changing Coriolis parameter may reach significant values.

[2] P. IV-10: Since we can calculate the mean upper-level divergence, one might reasonably ask why not compute mean divergence at low levels also, and find vertical motion? The answer is that this has been tried and the results do not repay the effort. What one finds is a diagnostic field with little or no prognostic value. This may be related to the 1200 and 0000 GMT rawinsonde times. However, it does seem that the low-level rawinsonde divergence field is more susceptible to weather "contamination" and has less continuity from synoptic time to synoptic time. While the upper-level fields are also "noisy" in this way, they have been found to be more useful in a prognostic sense than the old low-level (Sfc to 10000 ft, MSL) vertical motion charts once produced at SELS.

[3] P. IV-10: Actually, this can be done only to within a constant vector - the "translation" property of the wind. Since translation is everywhere constant within the field, it cannot be described by derivatives. See Saucier (1955, Ch. 10), Schaefer and Doswell (1979), Doswell (1982b) and their references for further discussion.

[4] P. IV-11: In visual terms of the wind analogy, a wind vector has one "arrowhead" showing which direction along the line is specified; whereas, the axis of dilation has two "arrowheads", so neither direction is preferred.

V. Interpretation of Numerical Guidance

A. General Remarks

The task of analysis is diagnostic in character. Knowing what is going on now is an essential beginning to forecasting what will be going on in the future. Prior to the advent of numerical prognosis, this first step of analysis received the benefit of considerable attention. The act of forecasting had to proceed in a largely intuitive way, with a heavy emphasis on "rule of thumb" (i.e., if this happens, then something else will follow) and extrapolation (i.e., it's been moving this way, it should continue to do so).

With the development of increasingly sophisticated numerical models, it has been possible to incorporate much of what we understand about the weather into the models. Not only can they extrapolate, they can predict new developments, dissipate old systems, and filter out "noise". The amount of detailed meteorological theory actually brought to bear by the models far exceeds that available to the analyst/forecaster. So what role does the human have in the process of producing a forecast? Perhaps the obvious answer is that despite their sophistication, the models still err. At times, their errors are far in excess of what a human forecast would create. Also, the machines occasionally break down for one reason or another and a human is still needed to salvage the product. Petterssen (1956a) says it quite well:

> While the machines provide the answers that can be computed routinely, the forecaster will have the opportunity to concentrate on the problems which can be solved only by resort to scientific insight and experience. Furthermore, since the machine-made forecasts are derived, at least in part, from idealized models, there will always be an unexplained residual which invites study. It is important, therefore, that the forecaster be conversant with the underlying theories, assumptions, and models. In particular, it is important that he be able to identify the "abnormal situations" when the idealized models (be they dynamical or statistical) are likely to be inadequate.

This text is not really the appropriate forum for a discussion in detail of the human forecaster's role, but the forecaster concerned with convective weather needs to be aware of the inherent limitations within the models (see Doswell, et al., 1981). With time, the current limitations may be superceded by others, as our understanding of the atmosphere (as reflected by numerical models) changes.

As discussed in Weiss and Ferguson (1982), numerical guidance in severe storm forecasting (and, perhaps, other areas as well) poses two different dilemmas. The first is to understand how the large-scale analyses and forecasts relate to the production of severe convection. It is not a straightforward process to go from an analysis to a depiction of where severe storms will occur. The same statement applies even to a perfect forecast. Presumably, combinations of large-scale parameters (perhaps involving different parameter sets in different synoptic-scale settings) can be developed to guide this process. Our current level of understanding is such that even a perfect model forecast (and/or analysis) often can lead to an imperfect forecast of convective weather.

The second dilemma concerns how well the models actually perform. This is especially true in forecasting convection, since all the parameters which are conceivably of interest to thunderstorm forecasting are not currently available directly from model output. It may be possible to construct most, if not all, of these parameters from the internal variables in a given model. However, it is not clear that that model (or any other) does as good a job with those variables as it does with, say, 500 mb heights. Thus, model guidance needs to be considered in light of which parameters can reliably be derived from model output (present or future). This is a cornerstone in the model output statistics (MOS) approach to forecasting (Glahn and Lowry, 1972). As Weiss and Ferguson (1982) suggest, we are far from a perfect knowledge of this, as well.

B. Short and Long Term Error History

At the current time, a multiplicity of models exists and the forecaster is sometimes faced with the dilemma of contradictory output from the different models. Some of this can be clarified by keeping track of how the various models have been behaving. If a particular model is doing quite well with system movement, but has been treating the development/decay of systems badly over, say the past week, then this should be considered when evaluating its latest forecast.

Naturally, short-term model behavior can change from day to day, and the way any given model handles any given weather system can change. For example, primitive equation models (the LFM is one) have a tendency for what is called "locked-in error" (Fawcett, 1969). This is a model-specific error (and, as such, really represents a long-term error) which can routinely be treated, and can (potentially) be eliminated or reduced by changes in the model. At certain stages in the evolution of a weather system, a given model can do rather poorly, whereas later in its life cycle, the system is well-handled.

A basic element in the short-term evaluation of any given model is the forecaster's knowledge and experience of how the atmosphere behaves. When model output is examined, the analyst/forecaster already should have in mind what is anticipated in terms of the overall trend. Should the model output contradict this trend, a further examination of the possible explanations for this difference is called for. If the analyst/forecaster examines the model output first, then there is a tendency to be biased by what is seen, and to accept the model results less critically.

For any given model, it is possible to develop a statistical picture of the model's behavior over a long period. There may be biases and consistent errors which can be accounted for. This sort of error analysis should be routinely done at the large forecast centers where the models are run, and transmitted to the field offices on a regular basis (e.g., Fawcett, 1969; Brown and Fawcett, 1972). With this information at hand, it should be possible to modify the model results subjectively to account for these consistent errors.

While this process is as yet imperfectly accomplished, some effort along these lines is being made (Leary, 1971; Tsui and Brody, 1982). Part of the reason for the imperfect application of this concept is the need for lengthy compilation of the appropriate statistics. Further, the models are altered at fairly frequent intervals, compromising the value of any accumulated statistics. The problems and magnitude of the task tend to inhibit its proper execution. Also, the lack of a communications channel by which forecast offices can be made aware of problems perceived in the field limits the process. Field offices are only vaguely aware of how the various models work, so their knowledge of model limitations is correspondingly vague. Further, they are often not aware of development efforts at NMC until after their implementation, if at all.

C. Initialization and Adjustment

One area of great potential value for the analyst/forecaster's contribution to model output enhancement is in an assessment of its initialization. The numerical forecast begins with initial data which have been interpolated to the model grid. If the analyst has done a careful job of producing an internally consistent 4-dimensional picture of the atmosphere's structure, he/she is prepared to evaluate how adequately the models are initialized. This presupposes that the analyst has available for examination the initial fields upon which the model operates.

Hales (1979a) has provided several good examples of how this can be accomplished when the model input has deficiencies in the data-void ocean areas. This sort of careful examination of model input, in comparison with observations, can be applied to data-rich areas as well. For example, if the analyst has concluded that a significant feature is found in the data by considering all available information, including satellite imagery, the model's initialization should be studied. If the model fields seem inconsistent with the analyst's assessment (the feature may have been poorly located or subjected to excessive smoothing), then the model forecast should be appropriately modified.

Not infrequently, this sort of initialization error may represent the major reason for disagreement between the subjective forecast and the model output. By changing the location and/or strength of an analyzed feature based on detailed analysis procedures, the analyst/forecaster can provide what is lacking in the models. Mesoscale details are simply too complex and too small for current numerical models and it is unrealistic to expect them in the models. Current and foreseeable models have very limited (if any!) capability to forecast convection and the forecaster/analyst should anticipate that a major effort is required to provide mesoscale detail to model output.

Further, this process should recognize that current models (and expected future revisions) are least accurate in the very long-term and very short-term ranges. The long-term range is not of overwhelming concern to the forecast of convection, but the short-term (12 hr or less) range is the area of greatest concern. Even with careful initialization procedures, numerical models start their forecasts with a period of "adjustment", wherein the variables within the model come into a state of balance dictated by the

model's governing equations. This adjustment period can take 6-12 hrs of forecast time, and is characterized by rapidly oscillating variables, while the mass and velocity fields reach a "mutual understanding". During this period, the output is unreliable.

The reasons for this adjustment period need not concern us here, but the result is that the 12-hr forecast may not be as reliable as the 24-hr prognosis. This fact, coupled with the problems posed by coarse initial data and inadequate convective physics, places great responsibility on the analyst/forecaster. The model output can generally be relied upon (within the limits supplied by its error history) to provide a broad-scale background of change, upon which are superimposed the details of specific interest to the convection forecast. It is up to the analyst/forecaster to provide those details, based on physical understanding not currently incorporated in the models.

D. Statistical Convective Weather Guidance

Within the last few years, a statistical approach to convective storms forecasting has been developed and put into operation. There are two distinct products, one for short-range use (Charba, 1979a) analogous to the SELS "Watch" product, and the other for medium-range use (Reap and Foster, 1979) analogous to the SELS "Convective Outlook" product. Both take the form of a map of convective storm probability, and both are derived by a mathematical process known as screening regression.

Briefly, screening regression proceeds in the following manner. Given a data set of occurrences ("predictands") for a particular event, a parameter set is offered to the screening regression program. The list of candidate parameters ("predictors") may run into the hundreds. The program searches the parameters to find the one parameter which explains the greatest amount of the variation (in space and time) in the event's occurrence. Then, given the first such parameter, what parameter in combination with the first explains the most variation? The process continues in this fashion until some chosen threshold is reached, where adding new parameters has reached the point of diminishing returns. The result is an equation which forms a weighted sum of all the chosen predictors such that, when values of the actual predictors are plugged into the equation, a probability for the event is produced.

The short-range statistical guidance makes heavy use of observed data, from the surface network and from radar, to

provide predictors. The medium-range product uses LFM-derived forecast predictors. There is considerable art involved in developing predictors, owing to certain limitations imposed by the screening regression method. Both products find that a substantial amount of the natural variation is associated with what they have termed an "interactive" predictor. This is essentially a modulated climatology, in which the climatological frequency is modified according to the value of another parameter (surface pressure for severe convection, the K-index for general thunderstorms).

In order to transform the probability values into a yes-no forecast, thresholds have been developed. These thresholds are still being experimented upon, and the exact methods await further research, especially in the medium-range products.

Comparative verification of these statistical products with those produced at SELS has not led to any definitive conclusions. Generally speaking, the short-range products do not detect as many severe occurrences as do the SELS watches, but they can have somewhat better "lead time" (Charba, 1979a). Conversely, the medium-range products have a greater chance of including the severe events within their thresholds, but they may do so at the expense of falsely alerting a substantially larger area (Weiss et al., 1980a, Weiss et al., 1980b). Recently, experimental medium-range statistical forecasts have demonstrated better skill at reducing the falsely alerted area (Reap et al., 1982).

From an analyst's viewpoint, these statistical guidance products are best dealt with in the same way as more conventional guidance. That is, the best strategy is to form a conception of when and where severe weather is likely to develop without having seen the guidance. Then, if there is a difference, the analyst/forecaster should try to understand the difference and make adjustments (if necessary) to the first conception, based on an examination of the differences.

An obstacle to this procedure (which also applies to all facets of interpreting numerical guidance) is that it is not always clear why the guidance is performing in the way it does. If the analyst is uncertain what went into the guidance product, there is not much available which can provide any insights, since the entire process is out of his/her hands. This is further complicated by unannounced changes to the model and to the statistical routines. Hopefully, enough stability in the system will eventually exist

that the analyst can begin to determine those circumstances when the guidance is best and when the guidance is most likely to fail. Charba's (1979b) efforts to document these synoptic situations when his product performs most poorly are a step in the proper direction. Much more of this sort of self-evaluation is necessary.

VI. Concluding Remarks on Mesoanalysis

It is hoped that this text has conveyed one idea above all. That idea is that mesoanalysis must be based upon physical understanding. Rather than tie the analysis process to a particular weather chart (or charts) and develop some all-powerful parameter (or set of parameters), the concept of integrating all available analysis tools into a physically consistent picture is heavily stressed.

If one works on the assumption that severe thunderstorms occur when unstably stratified air having sufficient moisture is lifted, then the analysis (which includes the physical interpretation) of data is dramatically simplified. "Features" in the data are no longer mysteriously combined to produce severe thunderstorm forecasts. Of course, this places a burden on the analyst/forecaster -- namely, he/she must understand how the "features" contribute to (a) development of unstably stratified air, (b) the presence [and the sufficiency] of moisture, and (c) the occurrence of upward vertical motion. Further, an awareness of how weather systems work (as we currently understand them) is essential to the proper accomplishment of the analysis. The development of conditions favorable to severe thunderstorms involves complex but basically understandable interactions among the three basic ingredients. That is, for example, lifting can act to destabilize the stratification, and the advective processes which introduce moisture and instability can also create upward motion. The basic elements of synoptic analysis are also applicable to the convective forecast -- the focus is modified from the evolution of large-scale weather systems by having to concentrate on details which are usually not important to the synoptic scale. However, as we have learned, the storms themselves and their mesoscale effects can have a dramatic influence on the large-scale systems, as well.

With a better physical understanding of how the large-scale pattern sets the stage, we no longer need to depend on a different set of rules for each situation. Severe weather episodes under unusual meteorological circumstances, such as northwesterly flow aloft (Johns, 1977, 1982a,b), or with exceptionally low dewpoints (Johns, 1982c), and in unique geographical locations like New England (David, 1977) or the High Plains (Doswell, 1980) should not be seen as anomalies, but as elements of the same basic picture.

Similarly, we should be aware that severe thunderstorms are not exclusively confined to springtime situations

involving strong cyclogenesis, although the majority of strong storms occur in the period April through June (Kelly et al., 1978). When wintertime severe storms do develop (Galway and Pearson, 1979; Burgess and Davies-Jones, 1979), they can easily be integrated into the patterns we have described. Summertime severe thunderstorms also fit (Maddox and Doswell, 1982). The seasonal variations in weather patterns can create a severe weather threat in a variety of ways. Different seasons are dominated by different parameters, but the basic building blocks of unstably stratified, moist, rising air are vital to severe storms and the analyst's job is to diagnose if, when, and where those basic building blocks will come together.

This concept also rules out a rigidly structured analysis program. Since the primary ingredients may be developed in a large variety of ways, it is not productive to lay out a rigid set of rules for analysis. Charts, parameters, and even concepts valuable to forecast a given situation may not be significant in the next. Further, although the basic elements of unstably stratified, rising moist air are necessary conditions, they are not by themselves sufficient. There are many unknown or poorly understood factors (e.g., microphysical interactions, the exact role of vertical shear, etc.) and it is naive to expect that our current concepts of how these factors interact to produce severe weather shall survive unchallenged for very long.

We have not spent a great deal of time giving specific details about how weather map analysis relates directly to severe weather. Some applications of the contents in this volume (and the next) are explored in Vol. III, but it is not possible to be exhaustive in any treatise of this sort. Rather, the interested reader will consult the references. It is the reader's responsibility (and distinctly to his/her advantage) to pursue further the topics mentioned in these notes -- to avoid doing so is to miss the point.

While we have asserted that there is a "big picture" based on physical understanding, one should not be deceived into thinking that anyone (especially the author!) fully understands that big picture. However, the analyst/forecaster can apply some fairly simple dynamical ideas which are valid on the synoptic and subsynoptic scale and use them to improve everyday weather analysis. By similar reasoning, the reader should be aware of the basic physical processes occurring on the thunderstorm scale. This is the goal of Volume II, Storm Scale Analysis.

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